

Tropical systems from the southwest Indian Ocean into southern Africa: impacts, dynamics and projected changes

Report to the
Water Research Commission

by

J Malherbe^{1,4}, FA Engelbrecht², WA Landman^{2,4}, T Lumsden³ & C Theale⁴

¹ ARC-Institute for Soil, Climate and Water

² CSIR

³ School of Bioresources Engineering and Environmental Hydrology, University of
KwaZulu-Natal

⁴ University of Pretoria

**WRC Report No. 1847/1/12
ISBN 978-1-4312-0384-0**

March 2013

Obtainable from

Water Research Commission
Private Bag X03
GEZINA, 0031

orders@wrc.org.za or download from www.wrc.org.za

DISCLAIMER

This report has been reviewed by the Water Research Commission (WRC) and approved for publication. Approval does not signify that the contents necessarily reflect the views and policies of the WRC, nor does mention of trade names or commercial products constitute endorsement or recommendation for use.

Executive Summary

The project considers the influence of tropical systems (cyclones, storms, depressions) from the Southwest Indian Ocean (SWIO) over the Limpopo River Basin. These systems have been linked to widespread heavy rainfall and subsequent flooding over the region. Due to the semi-arid nature of this area, variation and change in time of significant rain contributing synoptic systems are very relevant to the agricultural community. A wide range of climate datasets are examined from 1948 to 2012 to appreciate the climatology of these systems over the past few decades while simulations of an Atmospheric Global Circulation Model (AGCM) based on SST simulations by several Assessment Report 4 (AR4) Global Coupled Models (GCMs) are scrutinized to appreciate the projected change in occurrence under anthropogenic forcing towards the end of the 21st century.

Major findings of the study are:

- *The rainfall resulting from these systems over the Limpopo River Basin historically is confined to the January-March period, with a maximum around early February.*
- *Less than 10% of the average annual rainfall can be attributed to these systems but...*
- *...widespread heavy rainfall events are mostly the result of these systems (Ch. 5).*
- *The rainfall associated with these systems over the region varies with a quasi bi-decadal cycle, in phase with the Dyer-Tyson cycle.*
- *The extent and intensity of rainfall over the region related to these types of systems are linked to the track after landfall and, related to this and the wind vectors at various levels, to upper air divergence, surface convergence, convective available potential energy (CAPE) and precipitable water (Ch. 6).*
- *The quasi bi-decadal variation in the influence of these systems over the Limpopo River Basin is related to a change in the zonal nature of tracks after landfall.*
- *Most strongly related to landfall tendencies in the region, as previously noted, is strengthened easterlies in the subtropics with an anticyclonic anomaly to the southeast of southern Africa and a lower to mid-tropospheric ridge separating the systems from the westerlies to the south in the region (Ch. 7 & Ch. 8).*
- *Deep negative vorticity values are usually confined to regions far to the south of the southern African subcontinent while tropical systems are steered towards and into the subcontinent within a warm subtropical upper air ridge.*
- *The cyclic behaviour of the contribution to rainfall by these systems over the Limpopo River Basin is related to a broadening of the Hadley cell in the region (Ch. 9) and the interaction between Standing Wave 1 and 3 over the Atlantic and Indian Oceans respectively. On a hemispheric scale, these conditions are related to the tendency towards a positive Southern Annular Mode (SAM). Conversely, decadal scale variability of the tendency of upper air troughs to develop to the east of southern Africa during negative SAM is negatively related to multi-year periods with increased influence by tropical systems from the SWIO.*
- *On climate change time scales, the influence of tropical systems from the SWIO over the Limpopo River Basin is expected to decrease by the late 21st century under enhanced anthropogenic forcing (Ch. 10) due to...*
- *...a simulated intensification of the subtropical high pressure systems over the northern and eastern parts of the southern African subcontinent responsible for steering these systems further north, where an increase is simulated. In fact, ...*

- *...thermal and dynamical properties of the troposphere over the southern parts of the SWIO are expected to cause a decrease in the occurrence of tropical cyclones and also a decrease in the landfall of these systems.*
- *The projected decrease in landfall of tropical systems from the SWIO over the Limpopo River Basin is related to a relatively larger strengthening of the subtropical high pressure over parts of the subcontinent.*
- *Rainfall is expected to decrease over the region in response to the change in tracks of the systems from the SWIO and the strengthening of the ridge with...*
- *...a concurrent decrease in simulated design flood estimates for parts of the Limpopo River Basin by the end of this century (Ch. 11).*

Not only considering climate variability in terms of total rainfall, but in terms of contributing synoptic weather system also, helps create a better understanding of regional circulation types involved towards explaining the drivers of the variability.

Acknowledgements

The project team would like to thank the Water Research Commission for supporting the research conducted towards this report.

The following Reference Group members are thanked for their guidance and support throughout the project:

Mr C Moseki	: Water Research Commission (Chairman)
Dr GC Green	: Private Consultant
Dr M Rouault	: University of Cape Town
Prof C Reason	: University of Cape Town
Dr M Tadross	: University of Cape Town
Ms L Dyson	: University of Pretoria
Dr R Dube	: Private Consultant
Mrs CM Smit	: Research Coordinator

The South African Weather Service is acknowledged for the historic rainfall data used during the project. The Agricultural Research Council is also acknowledged for providing resources utilized and also for weather data used.

The editorial assistance by Dr TP Fyfield (ARC-ISCW) is gratefully acknowledged.

Table of Contents

Chapter 1 Introduction	1
Chapter 2 Identification of Tropical Systems	3
Chapter 3 Areas of Interest and Data	5
3.1 Rainfall data	5
3.2 Synoptic data – NCEP Reanalysis	6
3.3 Simulated synoptic data from CCAM for analysis of projected anthropogenic forcing	8
3.3.1 Introduction	8
3.3.2 Methodology	8
Chapter 4 Tracking of Closed Low Pressure Systems	11
Chapter 5 Historical Overview	12
5.1 Introduction	12
5.2 Results and Discussion	13
5.2.1 Impact of tropical systems from the SWIO over Limpopo River Basin	13
5.2.2 Cycles in the occurrence of landfalling tropical systems and associated rainfall	23
5.2.3 Trends in rainfall contributed by tropical systems over the Limpopo River Basin	27
Chapter 6 Factors Contributing to Intensity and Extent of Precipitation Associated with Tropical Systems from the SWIO over the Limpopo River Basin	30
6.1 Introduction	30
6.2 Discussion	30
6.2.1 3-8 February 2000 (Case 1)	30
6.2.2 21-28 February 2000 (Case 2)	35
6.2.3 7-13 February 1996 (Case 3)	39
6.2.4 17-20 February 1995 (Case 4)	43
6.2.5 3-7 March 2003 (Case 5)	47
6.2.6 2-6 March 1999 (Case 6)	51
6.3 Summary	55
Chapter 7 Case Studies of Steering Mechanisms of Landfalling versus Non-landfalling Tropical Systems	57
7.1 Introduction	57
7.2 Methodology	58
7.3 Results and Discussion	60
7.3.1 Landfalling systems – westward tracking	60
7.3.2 Landfalling system – northwestward tracking	60
7.3.3 Eastward-moving system	61
7.3.4 Southward-moving system	61
7.4 Discussion	69

Chapter 8 Synoptic Climatology of Landfalling Tropical Systems	77
8.1 Introduction	77
8.2 Methodology	77
8.3 Results and Discussion	78
8.3.1 Geopotential heights	78
8.3.2 Zonal winds.....	80
8.3.3 Meridional winds	81
8.3.4 Mixing ratio	83
8.3.5 Precipitable water content.....	85
8.3.6 Vertical average of zonal and meridional humidity flux	86
8.3.7 Vertical structure of SOM-derived atmospheric signals related to landfall versus non-landfall events.....	90
8.4 Conclusion	92
Chapter 9 Decadal-scale Dynamics as Related to Landfall of Tropical Systems and Subsequent Rainfall Contribution to the Limpopo River Basin	93
9.1 Introduction	93
9.2 Data and Methodology	95
9.3 Results and Discussion	96
9.3.1 Synoptic patterns and time series.....	96
9.3.2 Anomalous patterns during the previous maximum	104
9.4 Conclusion	110
Chapter 10 Projections of Change in the Climatology of Tropical Systems over the SWIO and into Southern Africa under Enhanced Anthropogenic Forcing	112
10.1 Introduction	112
10.2 Methodology	114
10.3 Results and Discussion	114
10.3.1 Simulation of warm-cored closed low pressure systems and TCLVs compared to observed data	114
10.3.2 Projected changes as simulated by the CCAM ensemble.....	120
10.4 Conclusions	124
Chapter 11 Projected Impact of Climate Change on Extreme Hydrological Events in Two Primary Drainage Basins in Limpopo.....	125
11.1 Introduction	126
11.2 Methodology	126
11.2.1 Study area.....	126
11.2.2 Hydrological modelling and input data.....	126
11.2.3 Estimation of design floods	130
11.2.4 Assessing climate change impacts on design floods.....	130
11.3 Results	130

11.3.1	Design floods based on observed climate data	130
11.3.2	Climate change impacts on design floods for different return periods.....	131
11.3.2	Climate change impacts on design floods for different event durations.....	135
11.4	Discussion and Conclusions.....	140
Chapter 12	Conclusions and Recommendations for Further Research	141
12.1	Decadal-scale variability	141
12.2	Climate change research	141
Chapter 13	Knowledge Dissemination and Capacity Development.....	143
13.1	Presentations at conferences	143
13.2	Scientific papers.....	143
Chapter 14	References	144

List of Tables

Table 5.1: Average contribution to yearly rainfall at the stations indicated in Fig. 3.1 over the period 1948 to 2008, by tropical systems from the SWIO.....	22
Table 5.2: Occurrence of widespread heavy rainfall events over the eastern parts of the Limpopo Province of South Africa and the contribution of well defined tropical systems from the SWIO to these events.....	22
Table 5.3: Results of the Mann-Kendall test for trend at the stations as indicated.....	28
Table 7.1: Tropical cyclone events used in the dataset showing the months they occurred, days of data, year and if they made landfall or not.....	59
Table 7.2: Characteristic steering mechanism differences between landfalling and non-landfalling tropical systems in the SWIO-southern Africa region.....	69
Table 10.1: Correlation matrix of CCAM ensemble and NCEP Reanalysis occurrence of tracks per grid point	119
Table 11.1: Definition of degrees of confidence relating to patterns of change in design flood estimates	135

List of Figures

Fig. 2.1: Thermal infrared image from METEOSAT 7 on 5 February 2000.....	3
Fig. 3.1: The Limpopo River Basin. Six rainfall recording stations used for throughout the study of historical impacts (Ch. 5 and 9) are also indicated	5
Fig. 3.2: Area of Interest 1 (entire region) and Area of Interest 2 (sub-region)	6
Fig. 3.3: Area of Interest 3	7
Fig. 3.4: Quasi-uniform C48 CCAM grid, having a horizontal resolution of about 2° in latitude and longitude.....	9
Fig. 3.5: Stretched C64 CCAM grid, having a horizontal resolution of about 0.5° over southern Africa and the SWIO.....	10
Fig. 5.1: Multi-day rainfall sequence indicating the influence of a tropical system from the SWIO over the northern parts of South Africa	13
Fig. 5.2: Objectively obtained occurrence of tropical systems making landfall from the SWIO at each of the 2.5° grid points over the area surrounding the Limpopo River Basin....	14
Fig. 5.3: Number of tropical systems from the SWIO per year responsible for rainfall over the Limpopo River Basin.....	15
Fig. 5.4: The average distribution per ten-day period of tropical systems responsible for rainfall over the Limpopo River Basin since 1948	15
Fig. 5.5a: Total rainfall in South Africa during inland passage of a tropical cyclone in 1948 (top left), tropical cyclone “A” in 1956 (top right), tropical cyclone Astrid in 1958 (bottom left) and tropical cyclone Bridget in 1960 (bottom right)	17
Fig. 5.5b: Total rainfall in South Africa during inland passage of tropical cyclone Claude in 1966 (top left), tropical cyclone Caroline in 1972 (top right), tropical cyclone Eugenie in 1972 (bottom left) and tropical cyclone Danae in 1970 (bottom right).....	18
Fig. 5.5c: Total rainfall in South Africa during inland passage of tropical cyclone Emlie in 1977 (top left), tropical cyclone Domoina in 1984 (top right), a tropical depression in 1995 (bottom left) and tropical cyclone Bonita in 1996 (bottom right)	19

Fig. 5.5d: Total rainfall in South Africa during inland passage of a tropical depression in 1996 (top left), a tropical depression in 2000 (top right), tropical cyclone Eline in 2000 (bottom left) and tropical cyclone Japhet in 2003 (bottom right).....	20
Fig. 5.5e: Total rainfall in South Africa during inland passage of tropical cyclone Favio in 2007 (left) and tropical depression Dando in 2012 (right)	21
Fig. 5.6: Contribution of tropical systems from the SWIO to 5-day rainfall events exceeding various threshold amounts at Entabeni on the eastern escarpment.....	23
Fig. 5.7: Normalized rainfall contributed by tropical systems from the SWIO as an average between all stations considered (Fig. 3.1).....	24
Fig. 5.8: Periodograms obtained for the 5-year moving averages of rainfall caused by tropical systems from the SWIO (left) and for total annual rainfall (right) at Entabeni. The 90% and 95% confidence levels are indicated (similar results were obtained for the other stations located in South Africa).....	25
Fig. 5.9: 5-year moving average rainfall for all 6 stations accumulated (thick solid line) and the 5-year moving average annual rainfall contributed by tropical systems from the SWIO (dotted line).....	26
Fig. 5.10: Normalized 5-year average rainfall contributed by tropical systems from the SWIO for all six stations (Fig. 3.1) and a linear trend line fitted to the curve.....	27
Fig. 5.11: Normalized 20-year average rainfall for all stations.....	28
Fig. 5.12: Time series of the yearly amount of events where more than 400 mm was measured within five days at Entabeni since 1923	29
Fig. 6.1: Total rainfall (mm) over southern Africa on 7 February 2000.....	31
Fig. 6.2: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of a tropical depression on 3 February (top) and 7 February 2000 (bottom)	32
Fig. 6.3: Moisture convergence and wind vectors on 7 February 2000 at 850 hPa (top) as well as vertical velocity (ω – Pa/s) at 500 hPa. The 700 hPa heights are indicated on both maps while the thick solid contours on the bottom map are the 500 hPa heights	33
Fig. 6.4: Wind vectors and divergence at 250 hPa (top) for 7 February 2000 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated.....	34
Fig. 6.5: Total rainfall (mm) over southern Africa on 24 February 2000.....	35
Fig. 6.6: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of tropical cyclone Eline on 19 February (top) and 24 February 2000 (bottom).....	36
Fig. 6.7: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 24 February 2000. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom).....	37
Fig. 6.8: Wind vectors and divergence at 250 hPa (top) for 24 February 2000 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated.....	38
Fig. 6.9: Total rainfall (mm) over southern Africa on 10 February 1996.....	39
Fig. 6.10: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of a tropical depression on 7 February (top) and 10 February 1996 (bottom).....	40
Fig. 6.11: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 10 February 1996. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom).....	41

Fig. 6.12: Wind vectors and divergence at 250 hPa (top) for 10 February 1996 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated	42
Fig. 6.13: Total rainfall (mm) over southern Africa on 20 February 1995	43
Fig. 6.14: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of tropical of a tropical depression on 16 February (top) and 20 February 1995 (bottom)	44
Fig. 6.15: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 20 February 1995. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)	45
Fig. 6.16: Wind vectors and divergence at 250 hPa (top) for 20 February 1995 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated	46
Fig. 6.17: Total rainfall (mm) over southern Africa on 5 March 2003	47
Fig. 6.18: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of tropical cyclone Japhet on 2 March (top) and 5 March 2003 (bottom)	48
Fig. 6.19: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 5 March 2003. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)	49
Fig. 6.20: Wind vectors and divergence at 250 hPa (top) for 5 March 2003 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated	50
Fig. 6.21: Total rainfall (mm) over southern Africa on 5 March 1999	51
Fig. 6.22: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of a tropical depression on 3 March (top) and 6 March 1999 (bottom)	52
Fig. 6.23: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 6 March 1999. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)	53
Fig. 6.24: Wind vectors and divergence at 250 hPa (top) for 6 March 1999 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated	54
Fig. 6.25: Zonal and less zonal tracks followed by tropical systems from the SWIO	56
Fig. 7.1: Steering flow (700-500 hPa) vectors (arrows) and zonal component (shading) for a landfalling tropical depression in February 2000 on 4 February (top) and after landfall on 7 February (bottom). Black contours represent the synoptic situation at 700 hPa on the 4 th and white contours represent the situation on the 7 th	63
Fig. 7.2: Temperatures at the 250 hPa level during the landfall of a tropical depression in February 2000 on 4 February (top) and after landfall on 7 February (bottom). Solid contours represent the synoptic situation at 700 hPa on the 4 th and dashed contours represent the situation on the 7 th	64
Fig. 7.3: Wind vectors and potential vorticity at 250 hPa during the landfall of a tropical depression in February 2000 on 4 February (top) and after landfall on 7 February (bottom). For each image, black contours represent the synoptic situation at 700 hPa at that specific time and the white contours represent the situation during the other relevant time	65

Fig. 7.4: Steering flow (700-500 hPa) vectors (arrows) and zonal component (shading) for a southward moving tropical cyclone (Fabriola) in January 1997 on 7 January (top) and on 9 January (bottom). For each image, black contours represent the synoptic situation at 700 hPa at that specific time and the white contours represent the situation during the other relevant time	66
Fig. 7.5: Temperatures at the 250 hPa level during the southward movement of a tropical cyclone (Fabriola) in January 1997 on 7 January (top) and on 9 January (bottom). For each image, solid contours represent the synoptic situation at 700 hPa at that specific time and dashed contours represent the situation during the other relevant time.....	67
Fig. 7.6: Wind vectors and potential vorticity at 250 hPa during the southward movement of a tropical cyclone (Fabriola) in January 1997 on 7 January (top) and on 9 January (bottom). For each image, black contours represent the synoptic situation at 700 hPa at that specific time and the white contours represent the situation during the other relevant time...	68
Fig. 7.7: 700 hPa height average (contour) and differences (landfall - non-landfall) for all cases while the tropical system is still in the Mozambique Channel (top) and for the entire sequence (bottom). Average values are shown in contours while differences are indicated in shading	71
Fig. 7.8: 700-500 hPa average wind vectors (top) for all cases (landfall and other), difference (landfall - other) in wind vectors for all cases while still in the Mozambique Channel (middle) and differences (landfall - other) in wind vectors for the entire sequences (bottom). Wind vectors are represented by arrows while the difference in the zonal wind component is represented by shading	72
Fig. 7.9: Surface temperature averages and differences (landfall - non-landfall) for all cases while the tropical system is still in the Mozambique Channel (top) and for the entire sequence (bottom). Average values are represented by contours while differences are represented by shading	73
Fig. 7.10: 250 hPa height average and differences (landfall - non-landfall) for all cases while the tropical system is still in the Mozambique Channel (top) and for the entire sequence (bottom). Average values are shown in contours while differences are indicated in shading	74
Fig. 7.11: 250 hPa average wind vectors (top) for all cases (landfall and other), difference (landfall - other) in wind vectors for all cases while still in the Mozambique Channel (middle) and differences (landfall - other) in wind vectors for the entire sequences (bottom). Wind vectors are represented by arrows while the difference in the zonal wind component is represented by shading	75
Fig. 7.12: 500-250 hPa average thickness (top) for all cases (landfall and other), difference (landfall - other) in thickness for all cases while still in the Mozambique Channel (middle) and differences (landfall - other) in thickness for the entire sequences (bottom). Difference values are represented by shading while the average values for the landfall and other cases are shown in black and white contours respectively	76
Fig. 8.1: The 700 hPa SOM for geopotential heights with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average value for the 26 tropical cyclone events is shown on the right	79
Fig. 8.2: The 700 hPa SOM for 600 hPa zonal wind with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average value for the 26 tropical cyclone events is shown on the right	81

Fig. 8.3: The 600 hPa SOM for meridional winds with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right).....	82
Fig. 8.4: The 600 hPa SOM for mixing ratios with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right).....	84
Fig. 8.5: The SOM for the precipitable water content with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right).....	85
Fig. 8.6: The SOM for the vertical average of zonal humidity flux with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right).....	87
Fig. 8.7: Shows the SOM for the vertical average of meridional humidity flux with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right).....	89
Fig. 8.8: The index values for geopotential height and temperature showing at which level the SOM has most significance in terms of tropical cyclone movement.....	90
Fig. 8.9: The index values for the zonal and meridional winds showing at which level these variables have most significance in terms of tropical cyclone movement according to the SOM analysis.....	91
Fig. 8.10: The index values for the relative humidity and mixing ratios showing at which level the SOM has most significance in terms of tropical cyclone movement.....	91
Fig. 9.1: Scaled annual rainfall totals contributed by tropical systems from the SWIO at four locations over the period 1948-2008 (similar results were obtained for Pafuri and Entabeni)	93
Fig. 9.2: SOM based on the standardized values with lighter (darker) shades showing positive (negative) deviations for the 850 hPa level (top) and Sammon map (bottom) showing the occurrence of days during landfall events within the identified nodes with lighter shades indicating a higher number of days with landfall events	97
Fig. 9.3: SOM based on the standardized values with lighter (darker) shades showing positive (negative) deviations for the 250 hPa level (top) and Sammon map (bottom) showing the occurrence of days during landfall events within the identified nodes with lighter shades indicating a higher number of days with landfall events	98
Fig. 9.4: SOM based on the standardized values with lighter (darker) shades showing positive (negative) deviations for the 100 hPa level (top) and Sammon map (bottom) showing the occurrence of days during landfall events within the identified nodes with lighter shades indicating a higher number of days with landfall events	99
Fig. 9.5: Time series of the occurrence of days within the 25 nodes of the SOM for the 250 hPa level. All landfall events responsible for at least some rain over any of the 6 stations used for rainfall analysis have been indicated with circles and those responsible for more than 50 mm in one day at any of the stations with dark circles.....	101
Fig. 9.6: Variation in concentration within each SOM of landfall event days per pressure level	102
Fig. 9.7: Time series of the difference between Pattern B Positive and Pattern B Negative (line) and annual rainfall (bars), Pattern B Positive and Pattern B Negative (bottom)	103

Fig. 9.8: Time series of the yearly number of days with Pattern B Negative (top) and Pattern B Positive (bottom) of the subcontinent for the lower (850 and 700 hPa), middle (600 and 500 hPa) and upper (250 and 100 hPa) levels in the atmosphere	104
Fig. 9.9: Average zonal wind along 40°E during the 1993-2001 period (solid contours), 1979-1992 and 2002-2011 periods (dashed contours) and the difference (shaded)	105
Fig. 9.10: Difference wind vectors of the 700-500 hPa flow for the period 1993-2001 with the 1979-1992 and 2002-2011 periods with the difference in zonal component shaded.....	106
Fig. 9.11: Difference wind vectors of the 250 hPa flow for the period 1993-2001 with the 1979-1992 and 2002-2011 periods with the difference in zonal component shaded	107
Fig. 9.12: Difference on 850 hPa convergence and difference in wind vectors for the period 1993-2001 with the 1979-1992 and 2002-2011 periods with the difference in zonal component shaded.....	107
Fig. 9.13: Difference in 850 hPa relative vorticity (shaded) and difference in wind vectors for the period 1993-2001 with the 1979-1992 and 2002-2011 periods	108
Fig. 9.14: 500 hPa height difference (m) for the period 1993-2001 with the 1979-1992 and 2002-2012 periods	109
Fig. 9.15: Categorized January-March rainfall (bars), 9-year moving average of SAM index (brown line), 9-year moving average of rainfall category (black line) and years with largest rainfall events associated with tropical systems from the SWIO (circles).....	110
Fig. 10.1: Tracks per grid point (shaded) of warm-cored closed lows originating over the SWIO and making landfall into southern Africa as simulated by the CCAM ensemble for 1961-1990 (top) and observed tracks per grid point according to NCEP Reanalysis data (bottom). Actual tracks are also indicated (for the CCAM-CSIROmk3.5 on the CCAM-Ensemble map).....	115
Fig. 10.2: Latitudinal occurrence of warm-cored closed low pressure systems from the SWIO moving across 28°E as simulated by CCAM ensemble (grey) and observed in NCEP (black) for the period 1961-1990	116
Fig. 10.3: Tracks per grid point (shaded) of TCLVs originating or moving across the SWIO as simulated by the CCAM ensemble for 1961-1990 (top) and observed tracks per grid point according to NCEP Reanalysis data (bottom). Actual tracks are also indicated (for the CCAM-CSIROmk3.5 on the CCAM-Ensemble map).....	117
Fig. 10.4: Distribution of areas where TCLVs were only simulated by CCAM (light grey), observed in NCEP Reanalysis data and not simulated in the CCAM ensemble (grey) and where TCLVs were both observed in NCEP Reanalysis data and simulated by the CCAM ensemble	118
Fig. 10.5: Correlation between the CCAM ensemble simulation (x-axis) and observed NCEP Reanalysis data (y-axis) of TCLVs in the region of interest	118
Fig. 10.6: Taylor diagram showing the statistical comparison of the simulation of TCLVs frequency of occurrence per grid point of the individual CCAM ensemble members (circles) and the CCAM ensemble mean (triangle) over the SWIO with the reanalysis distribution (cross)	119
Fig. 10.7: Ensemble mean difference (top) and number of ensemble members indicating an increase/decrease (bottom) in number of tracks of warm-cored closed lows (left) and TCLVs (right) originating or moving across the SWIO between 1961-1990 and 2071-2100	120
Fig. 10.8: CCAM ensemble average projected change in January-March rainfall (shaded) and 700 hPa heights (contours) between the 1961-1990 and the 2071-2100 periods	121

Fig. 10.9: Median change in the Convective Tropical Cyclone Genesis Parameter (unit: approximate average 10-yearly total number of tropical cyclones per 2° grid point) as calculated from the ensemble of CCAM members for the period 2071-2100 relative to 1961-1990	122
Fig. 10.10: The change in the dynamic (top left) and convective (top right) components of the Convective Tropical Cyclone Genesis Parameter, change in the average wind shear (ms^{-1} / 600 hPa) between 800 and 200 hPa (middle left) and the change in vorticity (10^{-6}s^{-1} – bottom left) dynamical components and change in the gradient of potential temperature (10^{-1}°C) between the surface and 500 hPa (middle right) and the change in relative humidity (%) at 600 hPa (bottom right) as calculated from the ensemble of CCAM members between the 1961-1990 and 2071-2100 periods	123
Fig. 11.1: Map of the study area for assessing projected climate change impacts on extreme hydrological events	127
Fig. 11.2: The <i>ACRU</i> agrohydrological modelling system: general structure (after Schulze, 1995).....	127
Fig. 11.3: Quinary Catchments in the study area overlaid by a 0.5° resolution grid representing the scale of the regional climate projections	129
Fig. 11.4: Design flood estimates based on observed climate (1950-1999) for 7-day duration events at a) 2 year, b) 5 year, c) 10 year, d) 20 year and e) 50 year return periods.	132
Fig. 11.5: Design flood estimates based on observed climate (1950-1999) for a 20-year return period for a) 1 day, b) 2 day, c) 3 day, d) 4 day, e) 5 day, f) 6 day and g) 7 day duration events	133
Fig. 11.6: Ratios of future to present 1-day duration design flood estimates derived from different regional climate projections for the 2, 10 and 20 year return periods.....	134
Fig. 11.7: Degree of confidence that design flood estimates will increase or decrease by 10% or more for the 2, 10 and 20 year return periods	136
Fig. 11.8: Degree of confidence that design flood estimates will increase or decrease by 20% or more for the 2, 10 and 20 year return periods	137
Fig. 11.9: Degree of confidence that design flood estimates will increase or decrease by 30% or more for the 2, 10 and 20 year return periods	138
Fig. 11.10: Ratios of future to present design flood estimates derived from the CCAM-CSIROmk3.5 climate projection for the 1 to 6 day event durations for the 2, 10 and 20 year return periods	139

List of Abbreviations

AGCM	Atmospheric Global Circulation Model
AR4	Assessment Report 4
CAPE	Convective Available Potential Energy
CCAM	Conformal Cubic Atmospheric Model
CGCM	Coupled Global Climate Model
CTCGP	Convective Tropical Cyclone Genesis Parameter
ENSO	El Niño Southern Oscillation
GCM	Global Coupled Model
IPCC	Intergovernmental Panel on Climate Change
ITCZ	Inter-Tropical Convergence Zone
PV	Potential Vorticity
PWC	Precipitable Water Content
SAM	Southern Annular Mode
SOM	Self-Organizing Map
SWIO	Southwest Indian Ocean
TCLV	Tropical Cyclone-Like Vortex
TCSGP	Tropical Cyclone Seasonal Genesis Parameter

Chapter 1

Introduction

The study focuses on all tropical systems following an inland track that enhances the contribution to widespread heavy rainfall over the Limpopo River Basin. Most of this region is a semi-arid area with rainfall ranging between 300 and 600 mm per annum and high temperatures during summer (Schulze, 1965). However, within this climatic context, it supports a large rural population dependent on rainfed agriculture as well as large national parks that leads to vulnerability to the impacts of rainfall variability (Vogel and O'Brien, 2003). The highest rainfall occurs over the eastern escarpment – stretching from the Drakensberg in the south, northwards over the eastern parts of Zimbabwe. Some of these areas receive more than 1500 mm of rain per annum. The region receives about 85% of its annual average rainfall during the October to March period with the peak of the rainy season coinciding with the maximum in the annual cycle of tropical disturbances that control the summer rainfall season to a large degree (Preston-Whyte and Tyson, 2000). These systems invade southern Africa in the form of tropical cyclones, tropical lows and easterly waves (van Heerden and Taljaardt, contributing to Karoly and Vincent, 1998) during this important part of the rainy season.

Each tropical cyclone season approximately nine such cyclones form over the Southwest Indian Ocean (SWIO; Mavume et al., 2009). These tropical systems form during the summer months between November and April, mainly occurring in the late summer period (Jury, 1993). Tropical cyclones form in the region of 40-80°E and 5-30°S where tropical cyclone genesis is favoured specifically in the region of 10-15°S (Jury, 1993). Favouring conditions in this region include sufficient coriolis, low level cyclonic vorticity, weak divergent easterlies in the upper levels, sea-surface temperatures greater than 27°C and a deep layer of moisture and convective instability (Jury, 1993). On average only a small percentage of tropical cyclones have made landfall each season in the past 50 years (Mavume et al., 2009). Mozambique and southern Africa seldom experience tropical cyclones with approximately one per season, if any (Vitart et al., 2003; Reason and Keibel, 2004; Reason, 2007).

Tropical cyclones that originate over the SWIO tend to have a southward re-curvature before they reach the African coast (Vitart et al., 2003). Tropical cyclone genesis seldom occurs within the Mozambique Channel but in the rare occurrence of this happening, they tend to move off in a southeasterly direction (Vitart et al., 2003). Often once a tropical cyclone has entered into the Mozambique Channel it moves towards the south and curves to the southeast back into the Indian Ocean (Reason and Keibel, 2004). A condition over southern Africa, associated with a tropical cyclone within the channel, includes a dominating high pressure region and dry conditions over the interior (Reason and Keibel, 2004; Reason, 2007).

Predicting the landfall of tropical cyclones, using synoptic conditions, is only possible up to a few days in advance (Vitart et al., 2003). It has been suggested, however, that large-scale conditions may aid in forecasting the likelihood months in advance since they influence both the frequency and track of such systems (Vitart et al., 2003). The major problem in the predictability in the risk of landfall of tropical cyclones is that their track varies considerably each season (Vitart et al., 2003). This has led to the suggestion that the zonal steering flow (the zonal winds in the middle to upper troposphere) could be linked to the risk of tropical cyclones making landfall, particularly over Mozambique, where tracks were found to be more zonal with a low steering flow. This means that during a stronger easterly flow over the region, tropical cyclones are more likely to make landfall.

Unusual tracks in tropical cyclones such as Eline (2000) and Favio (2007) as well as development of tropical cyclone Dera (2001) within the Mozambique Channel are particular cases that have been examined and documented, contributing to literature studies required to better understand the tropical cyclone systems affecting southern Africa (Reason and Keibel, 2004; Klinman and Reason, 2008; Reason, 2007). In February 2000 tropical cyclone Eline caused devastating floods in many regions of southern Africa as a result of its unusual penetration far inland over the southern African interior (Reason and Keibel, 2004). It was argued that three precursor synoptic conditions existed allowing for penetration over southern Africa (Reason and Keibel, 2004). These precursor conditions were a ridge to the south of the Mozambique Channel, a preceding trough over southern Africa and an easterly steering current at mid-levels. The effect of La Niña also had an influence, enhancing the favourable conditions for landfall (Reason and Keibel, 2004). This allowed for anomalously warm SSTs in the Mozambique Channel and over the Indian Ocean north and northeast of Madagascar. The following season, 2000/2001, was found to be below normal in the amount of cyclones that formed, but tropical cyclone Dera formed later in the season (Reason, 2007). This particular cyclone was unusual in that it formed in the Mozambique Channel and tracked southwards out of the channel without making landfall but contributed to significant rainfall over Mozambique (Reason, 2007). In February 2007 tropical cyclone Favio formed in the SWIO and had a very unusual track in that it did not recurve to the southeast as tropical cyclones usually do, but rather tracked southwards around Madagascar and turned northwestwards making landfall over Mozambique (Klinman and Reason, 2008).

During the 1995-2000 period, a relatively large number of tropical systems from the SWIO contributed to rainfall over the Limpopo River Basin with a concomitant increase in major flood events. The total rainfall attributed to these systems during this period superseded that of any other 6-year period since the beginning of atmospheric records over southern Africa. However, two other periods over the previous five decades have displayed similar increases in rainfall contributed by these systems. The impact and variability in the region creates a need for research on tropical cyclones that affect southern Africa and a better understanding of the synoptic conditions within the atmosphere during landfall which could aid in future predictions on their risk for making landfall (Reason and Keibel, 2004).

Given these observations, the aim of the project is to understand the contribution to rainfall over the region and also appreciate the variation in time of these systems collectively, considering a broader series of events than Dunn (1985). Hence the specific project objectives are to:

- Determine from historical synoptic-scale weather data and rainfall records the impact of tropical systems from the SWIO over the Limpopo Province and also the variation in time over the record length.
- Describe the systems in terms of the meso- to synoptic-scale dynamics of the systems.
- Investigate the occurrence of natural variability of the impact of tropical systems from the SWIO over the eastern parts of southern Africa and describe this within the context of broader climate variability in the region.
- Simulate historical cases of westward moving low-pressure systems with the conformal-cubic atmospheric model (CCAM).
- Analyze the simulations of climate change over southern Africa with the emphasis on changes projected in the frequency and occurrence of westward moving tropical lows over the SWIO, and associated changes in rainfall over Limpopo and Mozambique.

Chapter 2

Identification of Tropical Systems

Tropical systems (cyclones, depressions, lows) are considered to be warm-cored, upright systems characterized by strong low-level convergence and relatively warm upper atmospheric temperatures due to the latent heating of the atmosphere resulting from condensation. While strong uplift is found over the middle layers of the atmosphere, this is replaced by a divergent high pressure area in the upper layers (e.g. Dyson and van Heerden, 2002). The identification of tropical systems from synoptic data focuses on the identification of a closed low pressure area in the lower to middle levels of the troposphere also identified by a column of warmer air over the surrounding area (Nguyen and Walsh, 2001; Sugi and Noda, 2002). Within this study, the identification of closed tropical (warm-cored) systems therefore relies on the following criteria:

- (1) A closed low (minimum in geopotential height) at 700 hPa and 500 hPa that exists for at least 24 hours.
- (2) The closed low is replaced by relatively high temperatures compared to that of the surrounding areas at 250 hPa.
- (3) The closed low as described above can be identified while the system occurs over the SWIO.
- (4) A vorticity value lower than -0.000035 s^{-1} at 700 hPa – this vorticity threshold is identified by considering several low-pressure systems at 700 hPa from historical records and reanalysis simulations.

Fig. 2.1 is an example of such a warm-cored system in the form of a tropical depression moving into the Limpopo River Basin in February 2000.

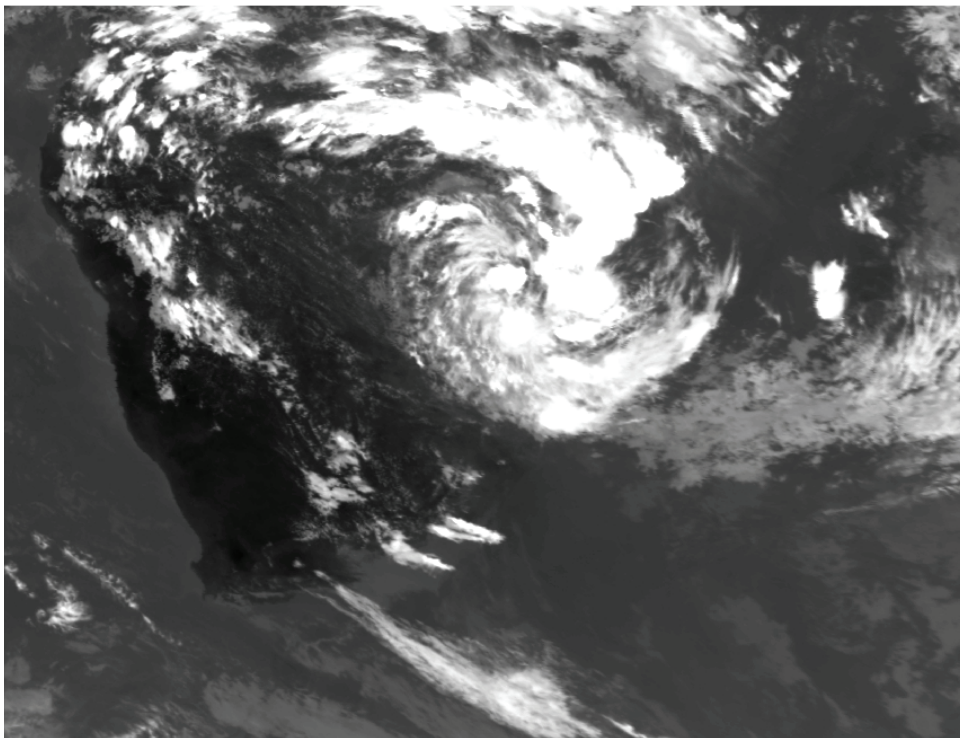


Fig. 2.1: Thermal infrared image from METEOSAT 7 on 5 February 2000

To distinguish between tropical cyclones and weaker warm-cored closed low pressure systems, the 10 m wind speed close to the centre of the 700 hPa low is required to exceed a threshold based on the spatial resolution of the input synoptic data at one grid point at least. For 2° data this value is 13 m/s and at 2.5° spatial resolution this value is 11 m/s (Walsh et al., 2007). NCEP Reanalysis data are resampled to a 2° grid similar to the CCAM C48 resolution for Intercomparison (but the minimum 10 m wind threshold is kept at 11 m/s).

The study focuses mainly on systems following an inland track that enhances the contribution to widespread heavy rainfall over the Limpopo River Basin.

Chapter 3

Areas of Interest and Data

3.1 Rainfall data

A large focus within this study is on the influence of these systems over the Limpopo River Basin, a relatively low-lying area characterized by a semi-arid climate with summer rainfall. This is indicated in Fig. 3.1 together with weather stations used in the study. These stations have continuous rainfall data records spanning the period 1948 to 2008 at least.

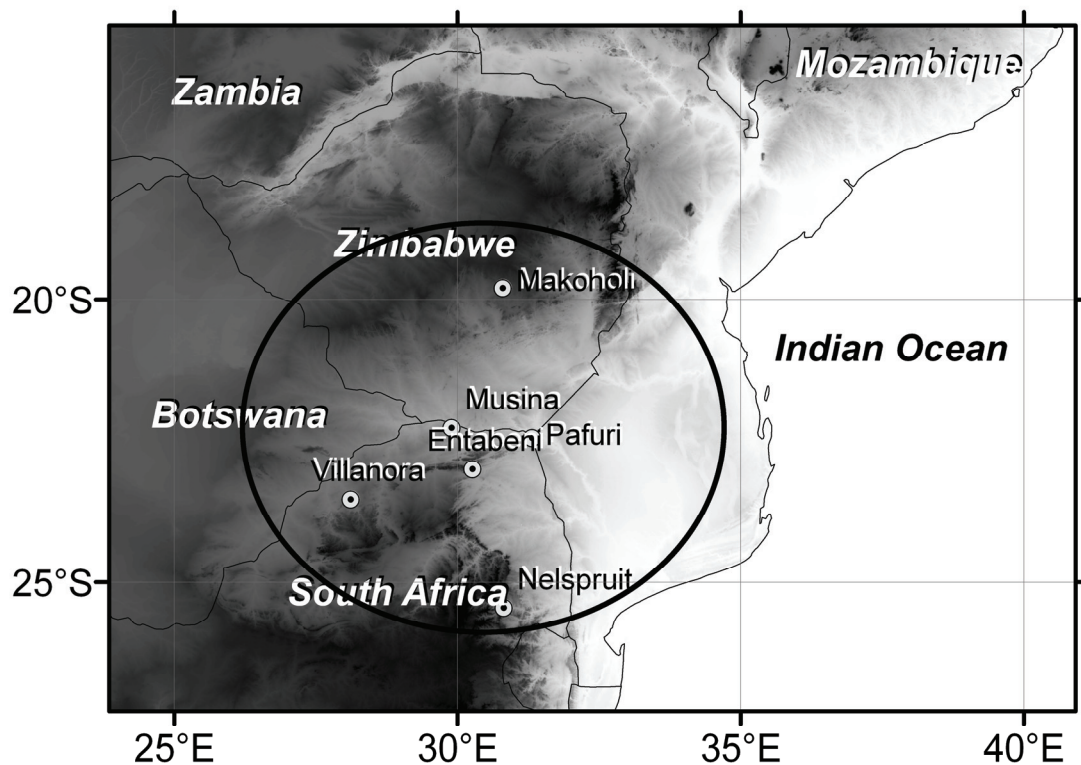


Fig. 3.1: The Limpopo River Basin. Six rainfall recording stations used for throughout the study of historical impacts (Ch. 5 and 9) are also indicated

Daily rainfall data for 1000-2000 rainfall stations since 1948 covering the whole of South Africa (from the Agricultural Research Council - Institute for Soil, Climate and Water (ARC-ISCW) Climate Information System which also contains data from the South African Weather Service (SAWS)) are used to compile daily rainfall maps used in Ch. 5 as one of the datasets to identify tropical systems from the SWIO contributing to rainfall over northeastern South Africa.

In Ch. 6, the rainfall contribution by individual tropical systems over the SADC region is considered. The daily 8 km resolution rainfall estimate (Xie and Arkin, 1997) produced by the National Oceanic and Atmospheric Administration Climate Prediction Centre (NOAA-CPC) is used. The daily totals are summed over the relevant periods for each system.

3.2 Synoptic data – NCEP Reanalysis

The historical synoptic data used during this project are the NCEP Reanalysis (Kalnay et al., 1996) 6-hourly synoptic data that cover the entire period since 1948 with a spatial resolution of 2.5°. Various parts of the project utilize a range of data from this dataset. For the work on historical periods (Ch. 6 to 9) the data are mostly spatially subsetting to the Area of Interest (AOI) indicated in Fig. 3.2. Several authors have cited improvements to the quality of NCEP Reanalysis throughout the time series of data. Improvements happened especially during the mid to late 1970s when satellite data were introduced (Kanamitsu et al., 1997; Sturaro, 2003; Tennant, 2004). The data over the higher latitudes towards the south were improved due to the inclusion of more surface observations since the early 1970s (Hines et al., 2000). In Ch. 9, NCEP Reanalysis data for the years 1979 to 2012 are used in the analysis of large-scale synoptic fields in the southern African region focussing on landfall. The reason for the limited period is the increase in accuracy in the data and the degree of spatial accuracy needed for the analysis performed. In this section, the geopotential height at all standard pressure levels available from 1000 hPa to 30 hPa is used.

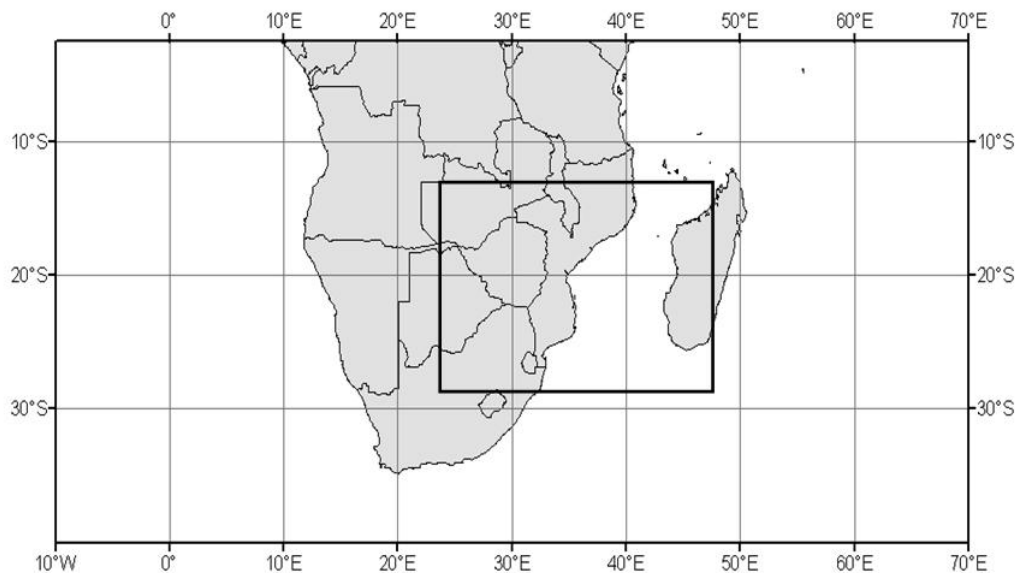


Fig. 3.2: Area of Interest 1 (entire region) and Area of Interest 2 (sub-region)

Certain sub-sections focus on the smaller domain indicated in Fig. 3.2. In the section on climate change (Ch. 10), the focus is more towards the east, as indicated in Fig. 3.3.

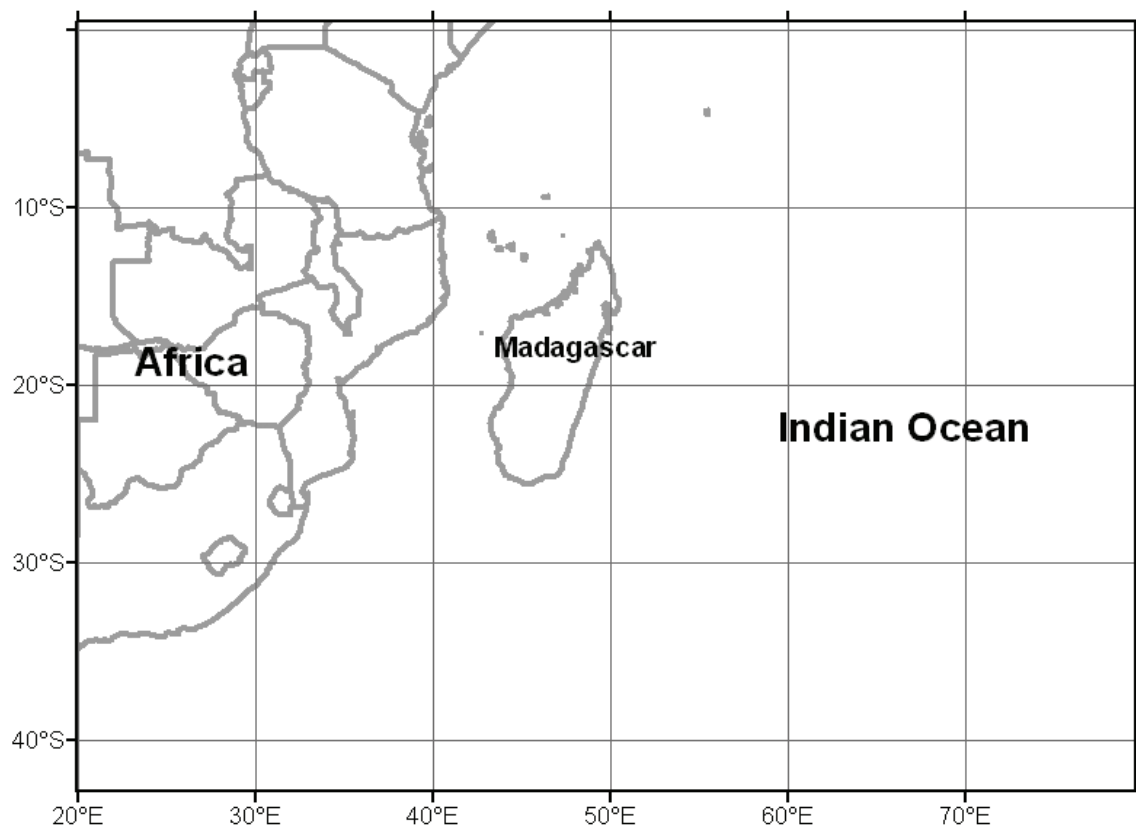


Fig. 3.3: Area of Interest 3

The two larger AOIs permit the tracking of tropical systems traversing the SWIO in southern Africa and also allow for an identification of the local circulation patterns associated with the driving mechanisms of these systems.

For a historical perspective of the influence of tropical systems from the SWIO over southern Africa (Ch. 5), the identification and tracking of tropical systems in the region relies in part on data for the longest period possible – starting in 1948.

For a comparison of cyclone climatologies as simulated by the CCAM ensemble (Ch. 10) and the observed, the data for the period 1961-1990 are used. For this section specifically, the NCEP Reanalysis data are resampled to a 2° grid.

The tracking algorithm employed for both the historical perspective (since 1948) and the tropical cyclone climatology (1961-1990) is based on the following data from this dataset:

Geopotential Heights

700 hPa

500 hPa

250 hPa

Zonal and Meridional Wind Vectors

10 m

700 hPa

Temperatures

250 hPa

For the study of dynamics of tropical systems moving over the subcontinents from the SWIO (Ch. 6) and study of steering mechanisms of observed cases (Ch. 7) as well as the study of the

synoptic dynamics in the southern African region during landfall/non-landfall (Ch. 8), the 2.5° spatial resolution 6-hourly data are also downscaled to a 50 km resolution using CCAM for the entire period from January 1989 to March 2010. Downscaling is performed using a multiple-nudging approach. CCAM is integrated in stretched-grid mode over southern Africa and the SWIO, at C64 resolution (approximately 0.5° in latitude and longitude over the area of interest). The high resolution grid is centred at 28°E and 25°S, which provides highest resolution over the Mozambique Channel and the eastern subcontinent of southern Africa – with the purpose of resolving the westward propagating tropical storms with sufficient detail. The high resolution simulations are nudged within the NCEP Reanalysis fields, through the application of a digital filter using a 1500 km length scale. The filter is applied at six-hourly intervals and for all the standard levels of the atmosphere. The choice of the digital filter is based on the CCAM's ability to capture the existence of actual tropical depressions and low pressure systems (the weaker systems that made landfall and also contributed towards rainfall over the eastern parts of the southern African subcontinent) and simulate them.

Output fields used include:

- Height of standard pressure levels
- Temperature at standard pressure levels
- Zonal/Meridional wind at standard pressure levels and 10 m
- Vertical velocity at standard pressure levels
- Humidity at standard pressure levels
- Mixing ratio at standard pressure levels
- Total column precipitable water
- Surface temperature
- Total column integrated temperature and moisture fluxes

3.3 Simulated synoptic data from CCAM for analysis of projected anthropogenic forcing

3.3.1 Introduction

An ensemble of high-resolution simulations of regional climate change over southern Africa and the SWIO has been performed at the Council for Scientific and Industrial Research (CSIR) Natural Resources and the Environment (NRE). These simulations may provide insight into the potential impact of enhanced anthropogenic forcing on the frequency of occurrence, intensity and tracks of tropical storms occurring over the SWIO. The conformal-cubic atmospheric model (CCAM), a variable-resolution global atmospheric model of the Commonwealth Scientific and Industrial Research Organization (CSIRO) Marine and Atmospheric Research in Australia (McGregor, 2005; Engelbrecht et al., 2009) was used to perform the simulations.

The resulting six-member ensemble is studied with regards to the impact of climate change on the frequency and tracks of tropical systems in the SWIO and the tendency to make landfall over eastern southern Africa.

3.3.2 Methodology

CCAM is forced with the bias-corrected sea-surface temperatures and sea-ice simulations (Katzfey et al., 2009) of six coupled global climate models (CGCMs) that contributed to Assessment Report 4 (AR4) of the Intergovernmental Panel on Climate Change (IPCC). All six CGCMs responded to greenhouse gas forcing corresponding to the A2 (business as usual)

emission scenario of the Special Report on Emission Scenarios (SRES), for the period 1961-2100.

The models are:

- GFDL-CM2.0 [The version 2.0 CGCM of the Geophysical Fluid Dynamics Laboratory (GFDL) of the National Oceanic and Atmospheric Administration (NOAA) in the USA]
- GFDL-CM2.1 [The version 2.1 CGCM of the Geophysical Fluid Dynamics Laboratory (GFDL) of the NOAA in the USA]
- ECHAM5/MPI-Ocean Model [The CGCM from MPI in Germany]
- UKMO-HadCM3 [The Met Office Third Hadley Centre Coupled Ocean-Atmosphere GCM in the UK]
- MIROC3.2-medres [Model for Interdisciplinary Research on Climate 3.2, medium resolution version, of the Japanese Agency for Marine-Earth Science and Technology]
- CSIRO Mark 3.5 [The version 3.5 CGCM of the Commonwealth Scientific and Industrial Research Organization (CSIRO) in Australia]

Data storage limitations and computational restrictions place a limit on the number of models that can be downscaled – these six models are selected based on their relative superiority in providing realistic simulations of El Niño Southern Oscillation (ENSO) events. Apart from their relative superiority in providing realistic simulations of ENSO events, these six models have also been shown to simulate average sea level pressure fields in the southern hemisphere and tropical belt (van Ulden and van Oldenborgh, 2006), response of rainfall over the Western Pacific to SST anomalies associated with the ENSO signal (Ose and Arakawa, 2009) and the Madden Julian Oscillation (MJO; Sato et al., 2009) relatively well compared to the full set of IPCC models. The ENSO phenomenon and the MJO have been shown to have an influence on the behaviour of tropical cyclones worldwide (Wang and Chan, 2002; Ho et al., 2006).

Downscaling is performed using a multiple-nudging approach. First, CCAM was applied at quasi-uniform C48 resolution (about 2° in latitude and longitude), with forcing from the host models as specified. The model grid is shown in Fig. 3.4.

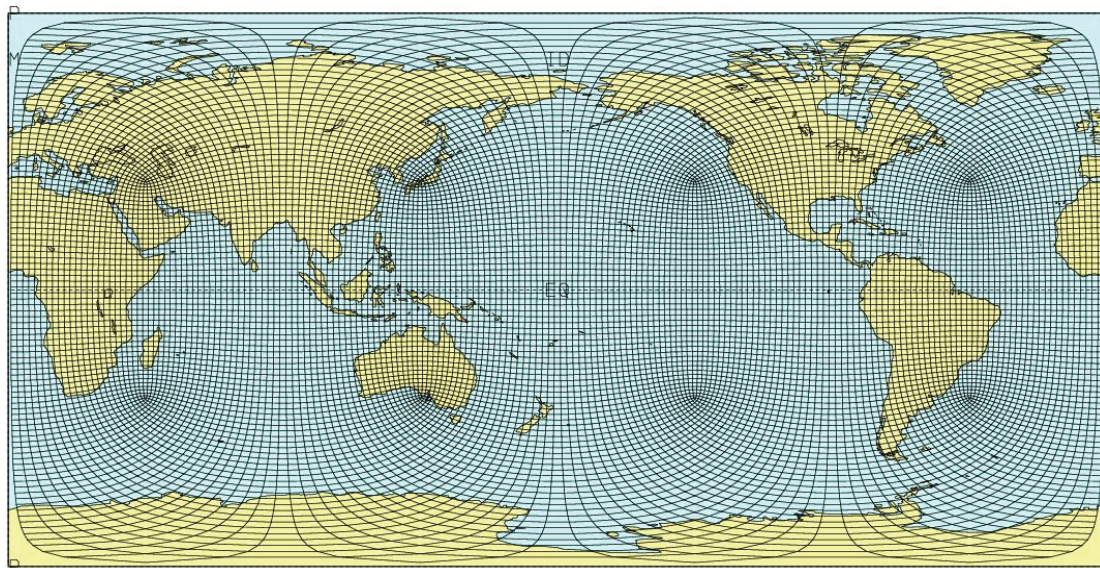


Fig. 3.4: Quasi-uniform C48 CCAM grid, having a horizontal resolution of about 2° in latitude and longitude

In a second phase of the downscaling, CCAM is integrated in stretched-grid mode over southern Africa and the SWIO, at C64 resolution (about 0.5° in latitude and longitude over the area of interest). The high resolution grid is centred at 28° E and 25° S (Fig. 3.5), which provides highest resolution over the Mozambique Channel and the eastern subcontinent of southern Africa – with the purpose of resolving the westward propagating tropical storms with sufficient detail. The higher resolution simulations are nudged within the quasi-uniform C48 simulations, through the application of a digital filter using a 4000 km length scale. The filter is applied at six-hourly intervals and from 900 hPa upwards. Model output for the six C48 resolution ensemble members is available at six-hourly intervals, whilst output for the C64 runs was written at hourly intervals.

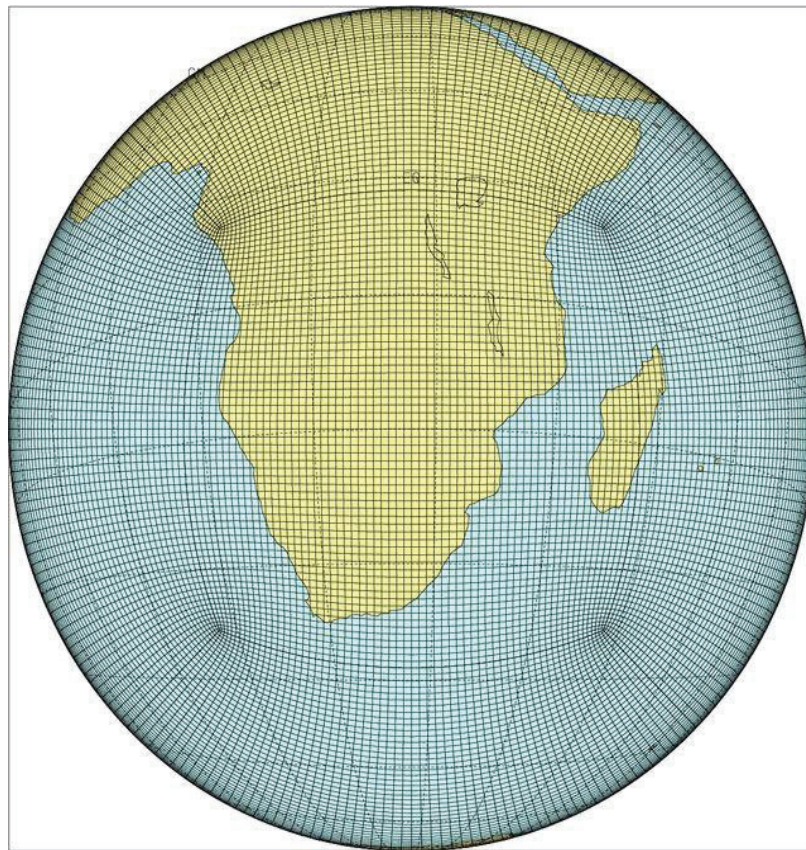


Fig. 3.5: Stretched C64 CCAM grid, having a horizontal resolution of about 0.5° over southern Africa and the SWIO

Chapter 4

Tracking of Closed Low Pressure Systems

Tracking algorithms within simulated data follow various atmospheric parameters in time and space in order to follow the position of a tropical cyclone-like vortex (TCLV). While certain variables are chosen to identify warm-cored low pressure systems as opposed to cyclones associated with the westerlies, cut-off values are chosen to help in the identification of intense warm-cored systems (TCLVs). The position of the TCLV is determined by a lower level geopotential height or sea level pressure anomaly (Walsh and Watterson, 1997; Tsutsui, 2002; Sugi and Noda, 2002) sometimes in the vicinity of a vorticity anomaly also at the lower levels (Bengtsson et al., 1995; Walsh and Watterson, 1997). To distinguish warm-core closed lows from cold closed lows (e.g. Taljaard, 1985), sometimes a prerequisite is that the wind speed in the lower troposphere needs to exceed that in the upper troposphere (Bengtsson et al., 1995; Nguyen and Walsh, 2001) and temperature and height anomalies up to a height of 300 hPa are also sometimes required to exceed that of the surrounding areas (Nguyen and Walsh, 2001; Tsutsui, 2002). Certain criteria are relaxed during tracking after identification of a TCLV. The wind speed needs to exceed a certain spatial resolution dependent threshold for distinguishing a TCLV from a weaker warm-cored closed low (Walsh et al., 2007) while the minimum distance requirements for the tracking of the system are also spatial and temporal resolution dependent (Blender and Schubert, 2000).

In this study, for tracking of warm-cored closed lows and TCLVs, the first step is to identify all local geopotential minima (closed lows) by comparing each grid point to the heights of the surrounding grid points on the desired pressure levels. Geopotential minima are identified by considering the geopotential height at all grid points relative to their surrounding 9-grid point stencil on the latitude-longitude grid – making exceptions where more than one local minimum defines the centre of a low pressure area by expanding the stencil and considering the algebraic average of the longitudinal and latitudinal coordinates of the grid points sharing the geopotential minimum value as the position of the closed low at that time level (e.g. Lambert, 1988).

The area of interest (Fig. 3.2) has been selected to include Madagascar and the Indian Ocean to the east of the island, to allow the tracking of tropical systems from 80°E westwards into southern Africa. Tracking is carried out on the geopotential height of the 700 hPa level only after identification of a warm-cored closed low (Ch. 2) through the application of an objective closed low tracking algorithm (Malherbe et al., 2012; Engelbrecht et al., 2012). It is done in an iterative procedure where all the height minima identified are subjected to a tracking procedure in time by finding the closest pairs of height minima over single time steps considering also the movement of the system as identified from a previous time step while taking minimum distance (700 km) and intensity change requirements into consideration (Malherbe et al., 2012).

Chapter 5

Historical Overview

5.1 Introduction

In order to quantify the contribution of tropical systems to rainfall over the eastern interior of southern Africa, it is necessary to objectively identify these systems throughout the length of a data record from a number of different data sources. Within this study, the focus is on tropical systems that can be identified as having formed a geopotential minimum (a closed low) in the mid-levels of the troposphere, and that have developed over the SWIO before making landfall over southern Africa. Such systems include tropical low pressure systems, tropical depressions, tropical storms and tropical cyclones.

The study area is the Limpopo River Basin and adjacent eastern escarpment in northeastern South Africa, southern Mozambique and southern Zimbabwe. In the context of the study, relevant systems are identified as having the properties (1 to 4) listed in Ch. 2 and also satisfy the following criteria:

- (5) The centre of the region of low pressure must make landfall (either in the mid-levels or in the lower atmosphere, but not necessarily in closed low form).
- (6) Be responsible for rainfall over the eastern interior of southern Africa – over the escarpment of South Africa and/or the Limpopo River Basin, within South Africa or Zimbabwe. Rainfall had to be measured at at least one of the six stations indicated in Fig. 3.1.

The availability of synoptic-scale circulation data limits the start of the period considered to 1948. Various datasets are analyzed in order to identify tropical systems that satisfy the six criteria mentioned above:

- 6-hourly National Centres for Environmental Prediction (NCEP) - National Centre for Atmospheric Research (NCAR) reanalysis data (Kalnay et al., 1996 – Ch. 3) for AOI 1 (Fig. 3.2)
- Daily synoptic data and weather maps from the South African Weather Service (SAWS)
- Daily rainfall data covering South Africa (from the ARC-ISCW Climate Information System which also contains data from the SAWS). Daily rainfall maps were interpolated from 1000-2000 rainfall stations from 1948
- Data for a rainfall recording station from the Zimbabwean Meteorological Services
- La Réunion cyclone tracks data – a database of tropical cyclones and tropical depressions that have occurred over the SWIO region since 1848

The following steps are taken to identify tropical systems making landfall and causing rainfall over the interior of the Limpopo River Basin:

- Visual inspection of NCEP Reanalysis data time lapse
- Consideration of the La Réunion cyclone tracks database
- Objective identification of closed-low tracks over the SWIO (Ch. 4). All the systems identified by the objective tracker formed part of the larger set of systems identified by visual inspection
- Consideration of daily weather bulletins obtained from the SAWS – more detailed sea level and 850 hPa fields than the NCEP Reanalysis data covering South Africa and the adjacent area – also supplemented with comments from forecasters

- Inspection of time series of interpolated daily rainfall maps over South Africa to identify possible influence of a tropical system over the far eastern/northeastern parts of the country

A rainfall sequence associated with the passage of tropical systems into areas adjacent to South Africa (Fig. 5.1) is used as further evidence. Daily rainfall maps have been interpolated for each day of the January-March period since 1948.

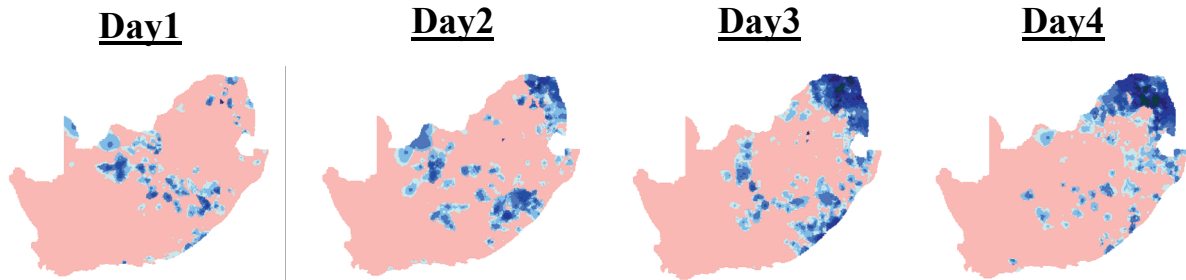


Fig. 5.1: Multi-day rainfall sequence indicating the influence of a tropical system from the SWIO over the northern parts of South Africa

In order to be added to an initial dataset, the presence of a tropical weather system has to be identifiable in at least three of the above-mentioned datasets. As a final prerequisite to be part of the final dataset, some rain had to occur at one or more of the three stations positioned furthest east in the area of interest (Fig. 3.1). These are Makoholi, Pafuri, Entabeni and Nelspruit.

Based on the annual rainfall totals associated with the identified tropical systems from the SWIO, Fourier analysis is performed for all stations separately to evaluate any evidence of cyclic behaviour. To emphasize the existence of cycles at periods of several years, the Fourier analysis is performed on the five-year moving averages of the original time series of rainfall, calculated for each station for the period 1948 to 2008. Statistical significance is established by comparing results with a set of 5000 Fourier analysis results for each station based on resampling the real rainfall data through a Monte Carlo process (Livezey and Chen, 1983; Wilks, 2006).

5.2 Results and Discussion

5.2.1 Impact of tropical systems from the SWIO over Limpopo River Basin

Fig. 5.2 shows the time series of the annual number of tropical systems from the SWIO that were responsible for precipitation over the Limpopo River Basin in South Africa or Zimbabwe.

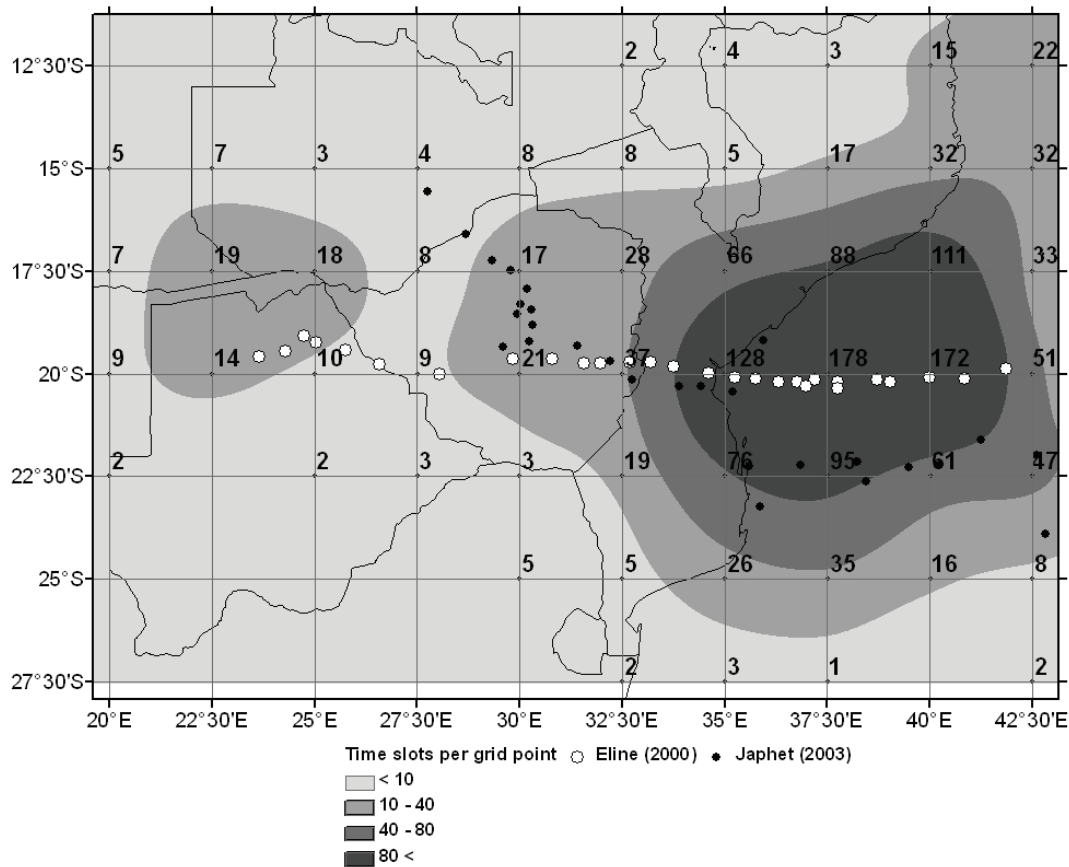


Fig. 5.2: Objectively obtained occurrence of tropical systems making landfall from the SWIO at each of the 2.5° grid points over the area surrounding the Limpopo River Basin

Fig. 5.2 reveals that the preferred tracks of westward penetrating tropical systems from the SWIO into the southern African subcontinent concentrate in the latitudinal band between and including 17.5°S and 20°S, coinciding spatially with the northern half of the Limpopo River Basin. The sharp west-east gradient near the coastline is the result of both the deflection to the north or south of some landfalling systems, as well as the loss of a closed circulation when some systems move westwards over land.

The number of tropical systems from the SWIO per year that are responsible for rainfall over the Limpopo River Basin, as identified through the combination of synoptic and rainfall datasets, is indicated in Fig. 5.3. The complete list of these systems is given in Malherbe et al. (2012).

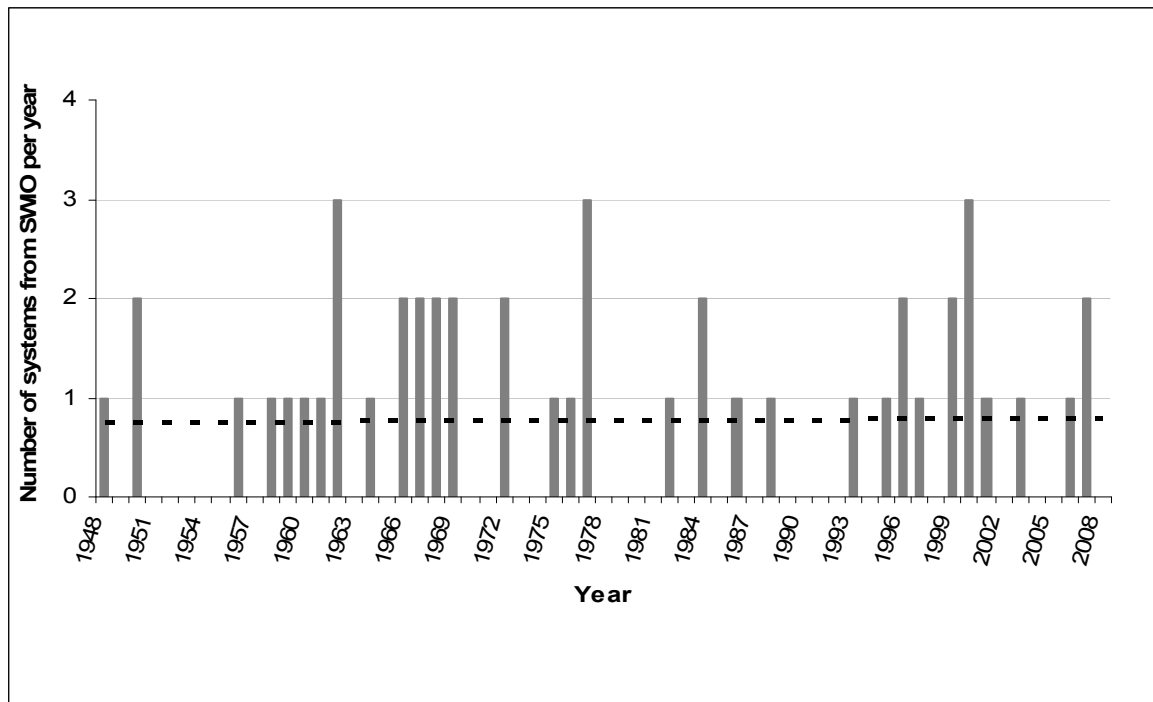


Fig. 5.3: Number of tropical systems from the SWIO per year responsible for rainfall over the Limpopo River Basin

There were two periods with a relatively high frequency of landfalling systems causing some rain over the area of interest – the 1960s/70s and the late 1990s, with fewer systems in the 1980s. There is also no strong trend visible in the time series. The average distribution per ten-day period is shown in Fig. 5.4.

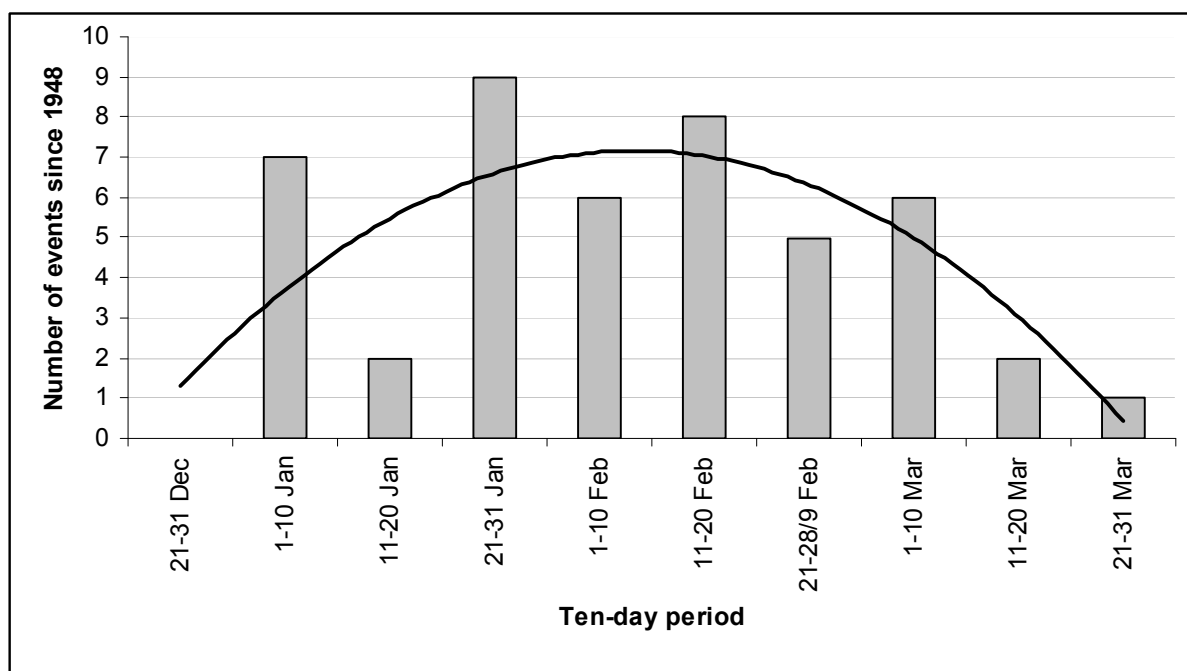


Fig. 5.4: The average distribution per ten-day period of tropical systems responsible for rainfall over the Limpopo River Basin since 1948

Fig. 5.4 shows that the average seasonal distribution of tropical systems causing rainfall over the Limpopo River Basin reaches a maximum around late January to mid-February, with a very definite start early in January and a more gradual end in late March. No such systems were responsible for rainfall over the AOI since 1948 during the last ten days of December.

Of the 44 systems identified, 18 were responsible for daily rainfall amounts greater than 50 mm at one or more rainfall stations shown in Fig. 3.1. This value (50 mm) in a 24-hour period is considered heavy rainfall by the SAWS (e.g. Dyson, 2009). The rainfall maps during the periods when these systems were active over southern Africa are shown in Fig. 5.5a-e. On several occasions, the interior trough was also responsible for rainfall over the central, southern and western parts of the country, not related to the rainfall caused by the tropical system in the northeast/north.

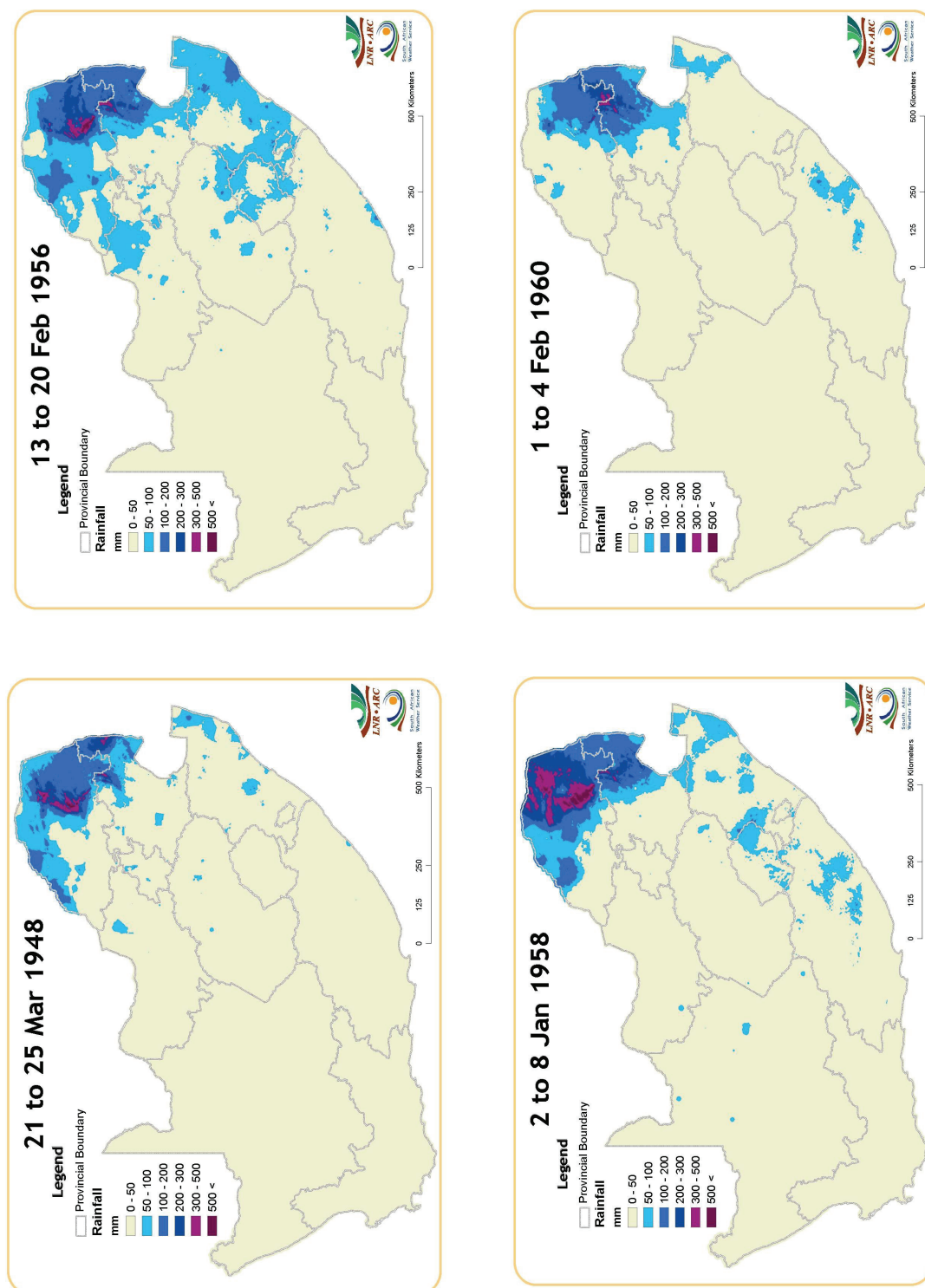


Fig. 5.5a: Total rainfall in South Africa during inland passage of a tropical cyclone in 1948 (top left), tropical cyclone "A" in 1956 (top right), tropical cyclone Astrid in 1958 (bottom left) and tropical cyclone Bridget in 1960 (bottom right)

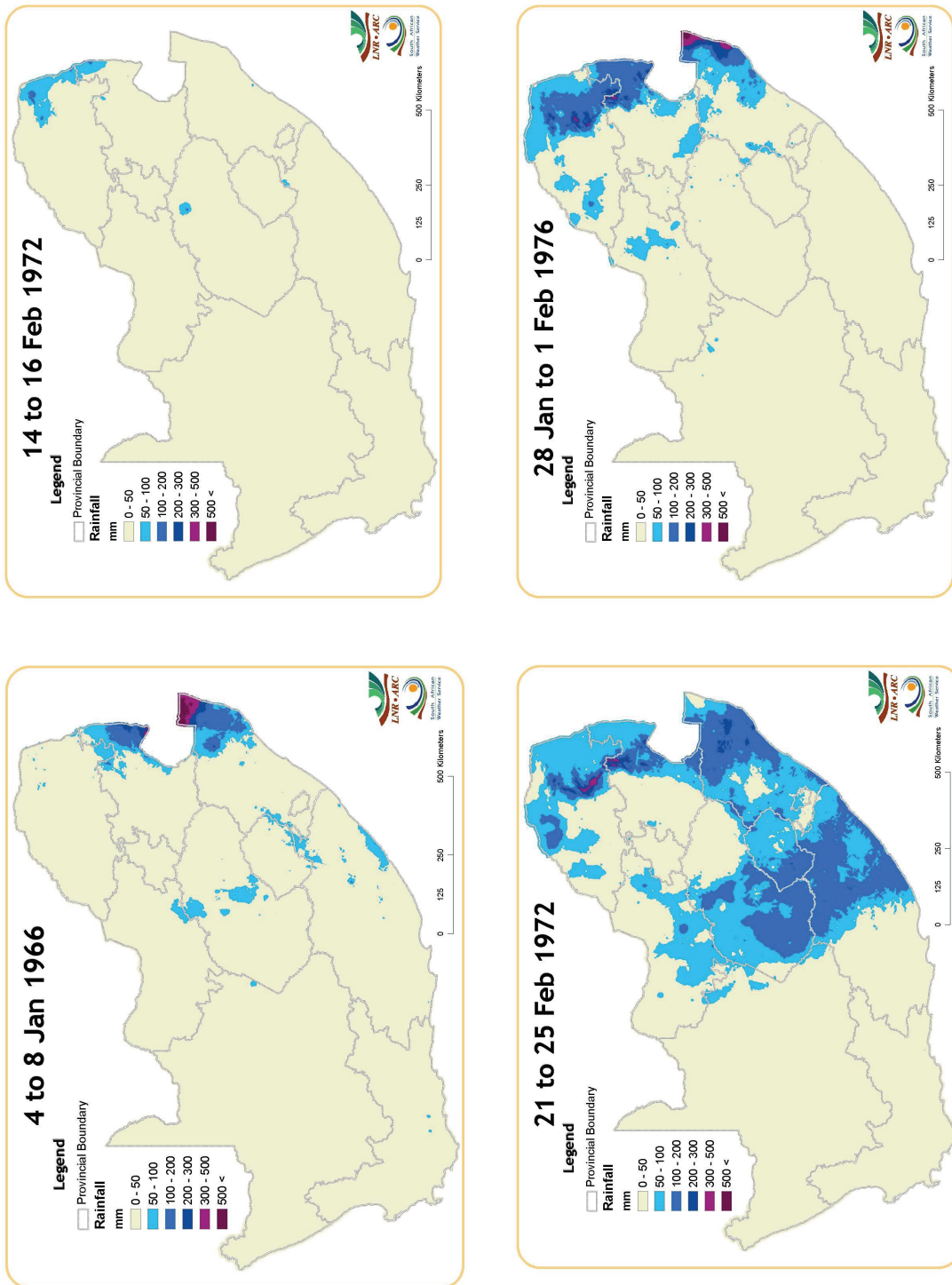


Fig. 5.5b: Total rainfall in South Africa during inland passage of tropical cyclone Claude in 1966 (top left), tropical cyclone Caroline in 1972 (top right), tropical cyclone Eugene in 1970 (bottom left) and tropical cyclone Danae in 1970 (bottom right)

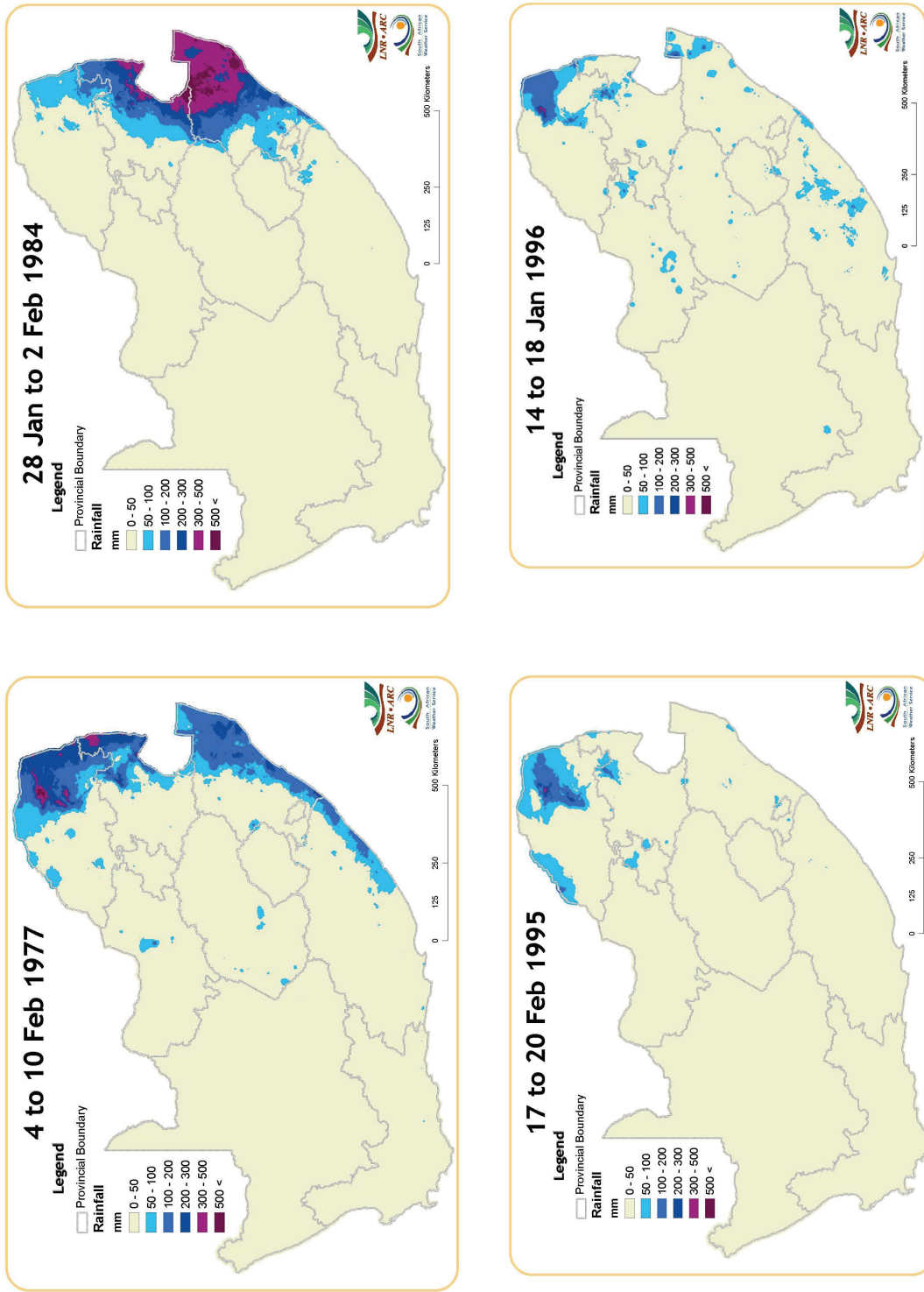


Fig. 5.5c: Total rainfall in South Africa during inland passage of tropical cyclone Emile in 1977 (top left), tropical cyclone Domoina in 1984 (top right), a tropical depression in 1995 (bottom left) and tropical cyclone Bonita in 1996 (bottom right)

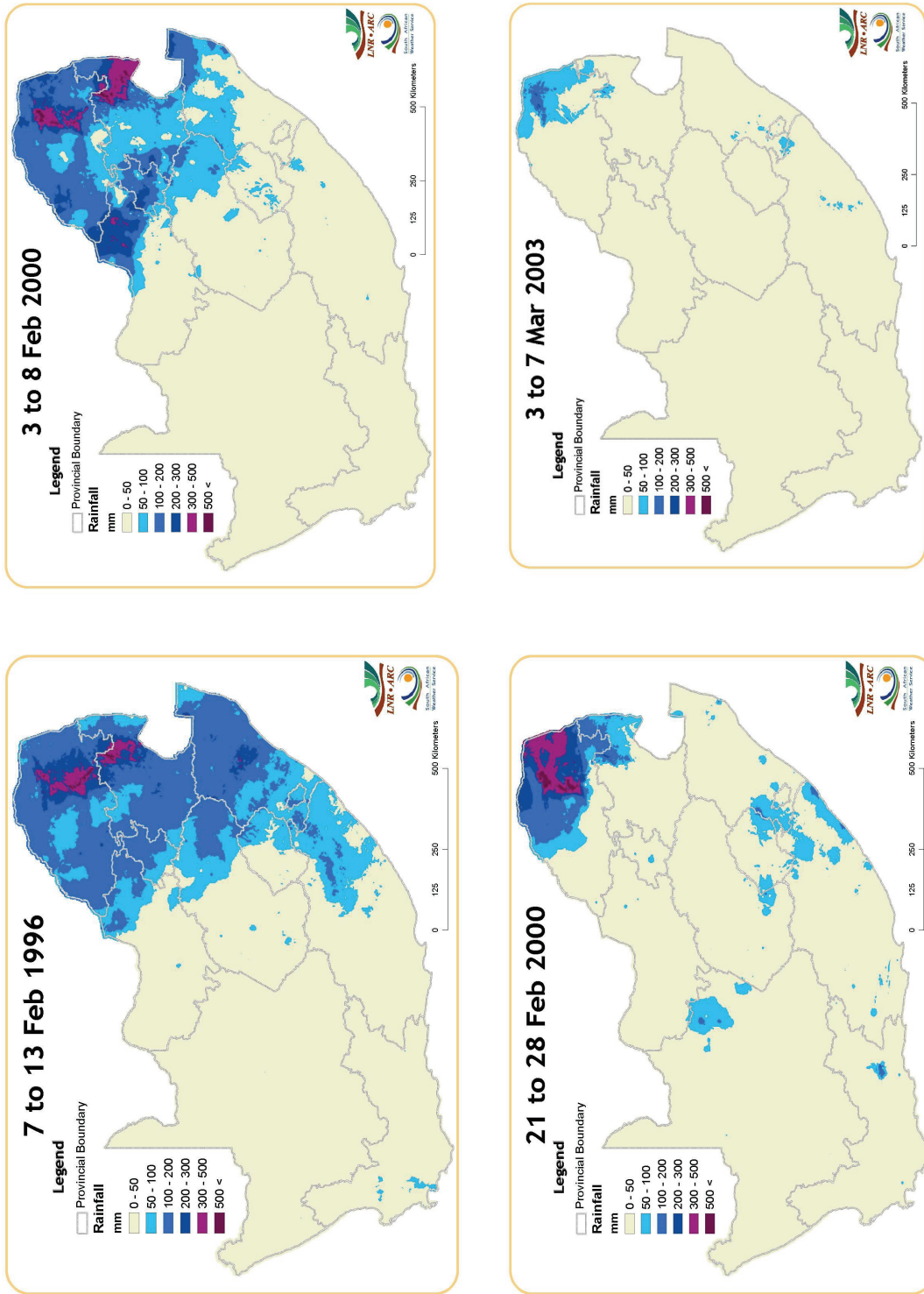


Fig. 5.5d: Total rainfall in South Africa during inland passage of a tropical depression in 1996 (top left), a tropical depression in 2000 (top right), tropical cyclone Eline in 2000 (bottom left) and tropical cyclone Japhet in 2003 (bottom right)

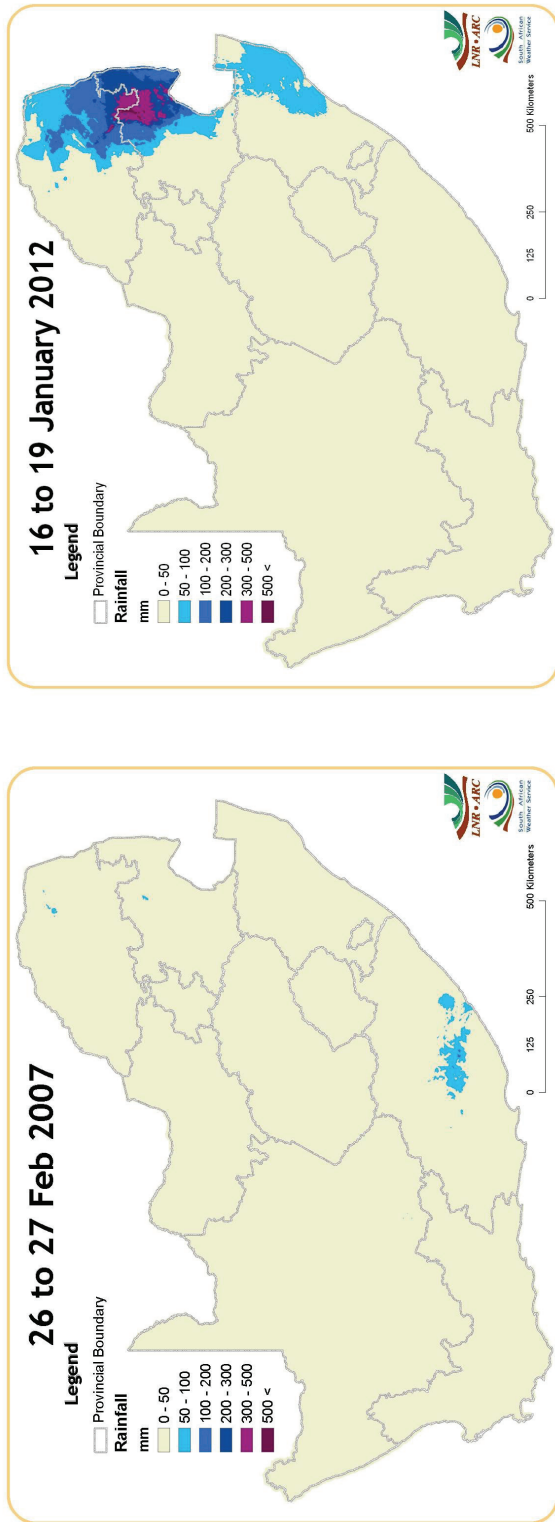


Fig. 5.5e: Total rainfall in South Africa during inland passage of tropical cyclone Favio in 2007 (left) and tropical depression Dando in 2012 (right)

For the analysis of rainfall events typically associated with these systems, a moving period of 5 days with daily increments is used – based on the average duration of rainfall events associated with the passage of these systems over the interior obtained from station rainfall data in the northeastern parts of South Africa. From the data of available stations over South Africa and Zimbabwe (Fig. 3.1), the total contribution to rainfall during the years since 1948 by tropical systems from the SWIO is around 7% over the escarpment (Entabeni) and Limpopo River Basin of northeastern South Africa (Musina, Pafuri) and southern Zimbabwe (Makoholi), with lower contributions further to the south and west. Table 5.1 summarizes the contribution of these systems to the total annual rainfall and late summer (here defined as January to March) rainfall for the six selected stations in the northeast of South Africa and southern Zimbabwe.

Table 5.1: Average contribution to yearly rainfall at the stations indicated in Fig. 3.1 over the period 1948 to 2008, by tropical systems from the SWIO

Station	Nelspruit	Entabeni	Pafuri	Musina	Villanora	Makoholi (Zim)
Annual rainfall (mm)	847	1795	440	360	421	647
Tropical system contribution - All year (%)	4	7	7	8	3	7

The westward and southward reduction in total rainfall contribution by tropical systems can be seen from Musina (8% average contribution to annual total rainfall) to Villanora (3%) in the west and Nelspruit (4%) in the south. It is in terms of widespread heavy rainfall events, though, that these systems become important contributing factors.

The importance of tropical systems from the SWIO over the area of interest in terms of widespread heavy multi-day total rainfall can be seen in Table 5.2. The table is based on the simultaneous rainfall at Musina, Pafuri and Entabeni (Fig. 3.1). The area within the triangle connecting these three stations is about 4 400 km². Simultaneous rainfall events at all three of these stations are identified and the relative contribution of tropical systems from the SWIO to such events is calculated. A total of 50 mm within 3 days or 100 mm within 5 days exceeds the 95th percentile of total rainfall simultaneously at all three stations (cut-off values used in Table 5.2).

Table 5.2: Occurrence of widespread heavy rainfall events over the eastern parts of the Limpopo Province of South Africa and the contribution of well-defined tropical systems from the SWIO to these events

	3-day total rainfall exceeding 50 mm	5-day total rainfall exceeding 100 mm
Total	33	7
Jan-Mar	55%	86%
Tropical system from SWIO	27%	57%

Table 5.2 puts the contribution of tropical systems into perspective. For only the widespread heavy rainfall events, Table 5.2 shows in the first row the total number of occurrences of such events according to the definition for each column. The second row shows what percentage of the number of these events occurred during the second part of summer (January to March) while the last row shows what percentage was as a direct result of rainfall caused by tropical systems from the SWIO. Tropical systems from the SWIO were responsible for more than a

quarter of very high 3-day total rainfall events and for more than half of 5-day widespread heavy rainfall events over the area. The increasing importance with higher multi-day rainfall events along the escarpment is shown in Fig. 5.6 which indicates also that the highest 5-day total rainfall amounts (at Entabeni) since 1948 were associated with the passage of tropical systems from the SWIO.

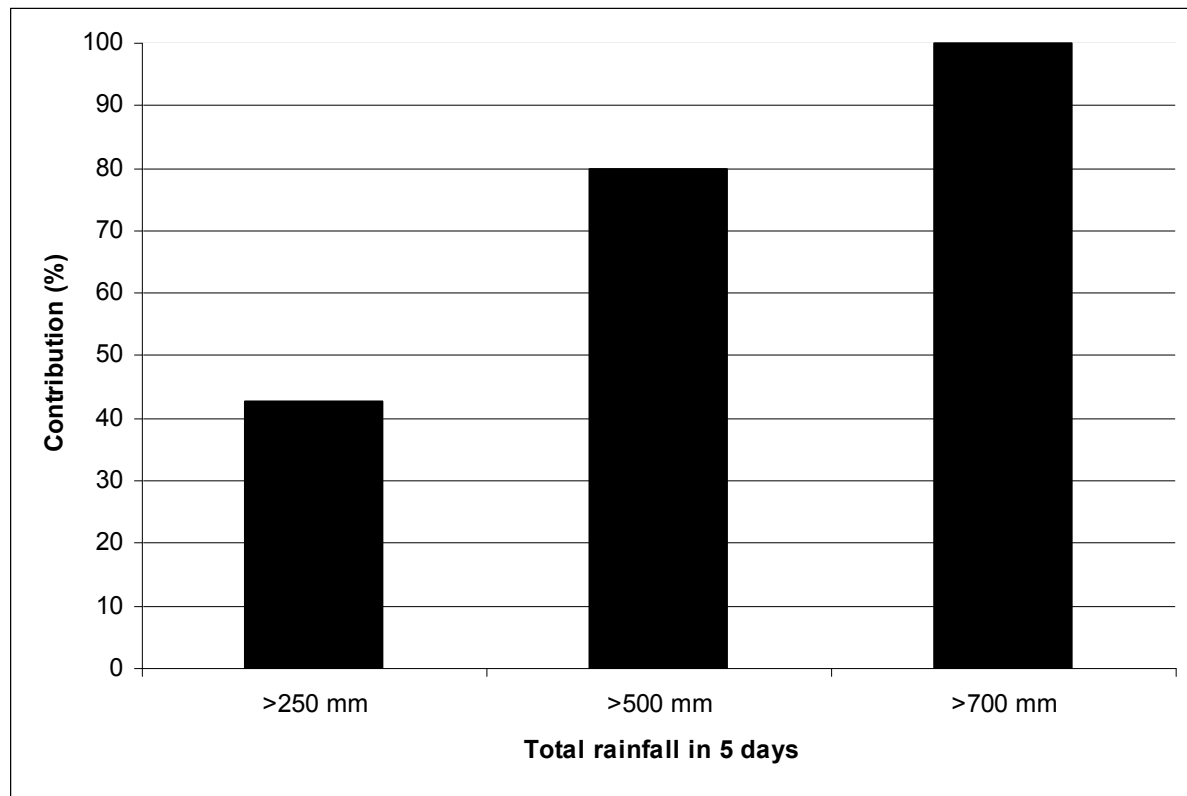


Fig. 5.6: Contribution of tropical systems from the SWIO to 5-day rainfall events exceeding various threshold amounts at Entabeni on the eastern escarpment

5.2.2 Cycles in the occurrence of landfalling tropical systems and associated rainfall

Because cycles have been observed in the landfall of tropical cyclones over various ocean basins (e.g. Chan and Xu, 2009) and to investigate any similar behaviour over the Limpopo River Basin, a Fourier analysis is performed on the station rainfall data in order to establish whether there exists any cyclic behaviour in the occurrence of rainfall caused by landfalling tropical systems over the area of interest, over the period 1948 and 2011.

Fig. 5.7 shows the normalized annual rainfall contributed by tropical systems from the SWIO as an average between the six stations used.

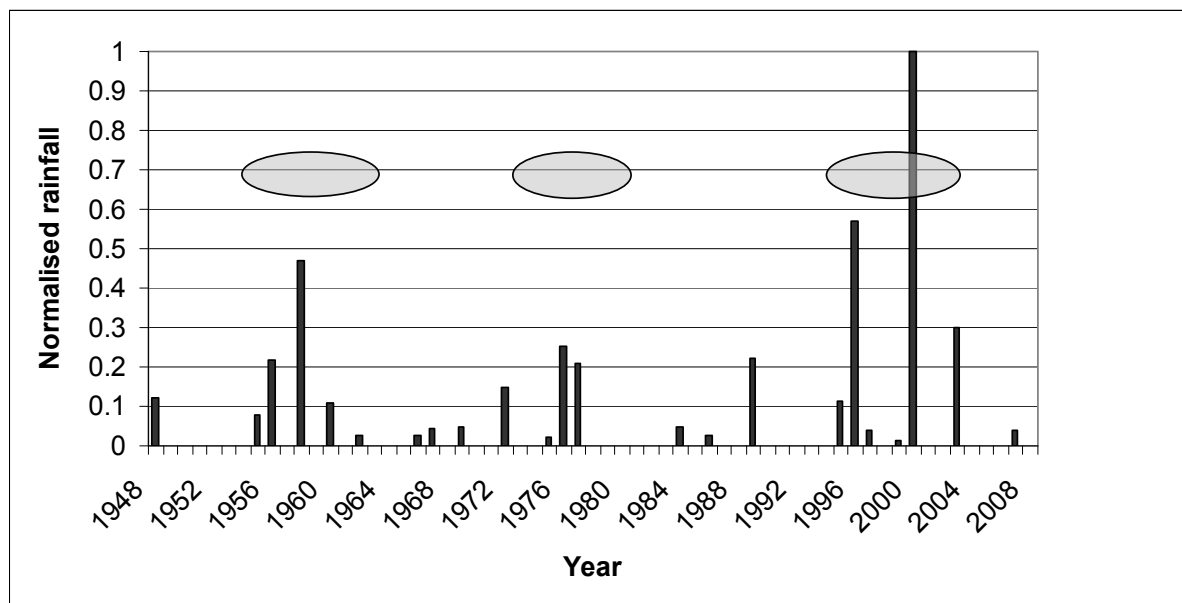


Fig. 5.7: Normalized rainfall contributed by tropical systems from the SWIO as an average between all stations considered (Fig. 3.1)

It is clear that there were multi-year periods since 1948 during which more rainfall was contributed by these systems over the Limpopo River Basin. This behaviour is further highlighted by a quasi-bidecadal peak in the periodograms of the resulting Fourier analysis performed on the annual rainfall contributed by tropical systems over the stations located over northeastern South Africa. Fig. 5.8 shows the periodogram performed on the rainfall data for Entabeni. To emphasize the existence of cycles at periods of several years, the Fourier analysis was performed on the five-year moving averages of the original time series of rainfall.

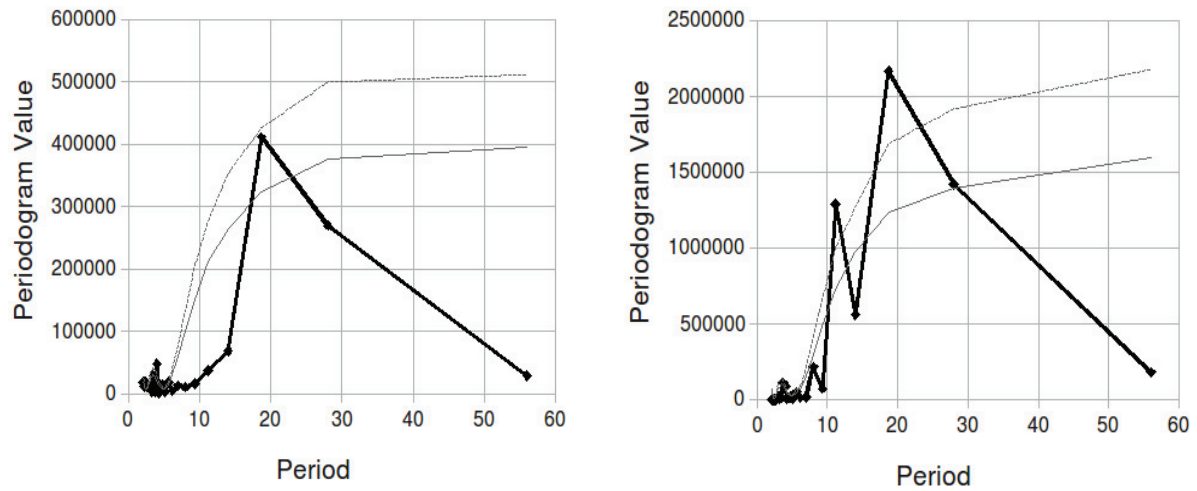


Fig. 5.8: Periodograms obtained for the 5-year moving averages of rainfall caused by tropical systems from the SWIO (left) and for total annual rainfall (right) at Entabeni. The 90% and 95% confidence levels are indicated (similar results were obtained for the other stations located in South Africa)

On the left-hand side (Fig. 5.8) the periodograms for the rainfall caused by tropical systems from the SWIO are shown, while the right-hand side shows the periodograms resulting from Fourier analysis performed on the total annual rainfall at each station. The 90% and 95% confidence levels are also indicated. Similar results were obtained for the other stations as well except for Makoholi in Zimbabwe where the quasi-bidecadal cycle was not significant (Malherbe et al., 2012). This quasi-18-year oscillation is statistically significant above the 90% confidence level for all the South African stations considered and remains significant at that level also for the total annual total rainfall at Musina, Entabeni and Pafuri. The quasi-bidecadal cycle in rainfall further also coincides with the cycle in total rainfall (Fig. 5.9).

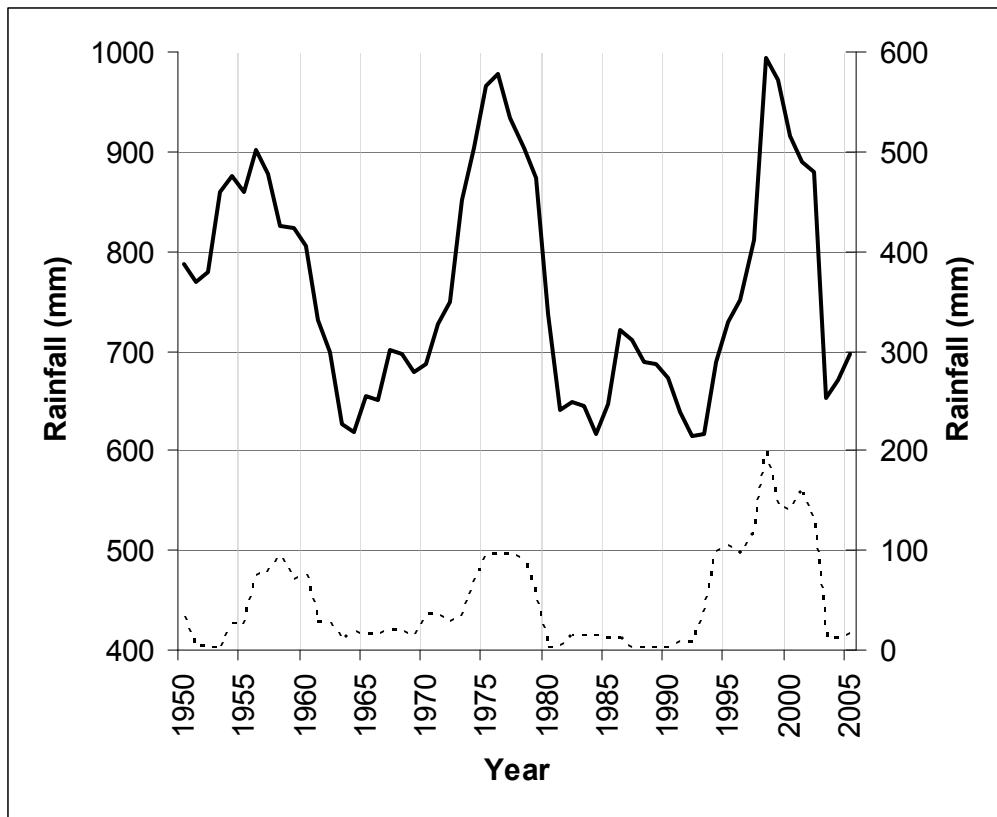


Fig. 5.9: 5-year moving average rainfall for all 6 stations accumulated (thick solid line) and the 5-year moving average annual rainfall contributed by tropical systems from the SWIO (dotted line)

The average contribution by tropical systems from the SWIO ranges between 2% for years with below-average rainfall to 8% for years with above-average rainfall, calculated by considering the average for all six stations considered (Fig. 3.1).

This rainfall cycle, with peaks around the late 1950s, late 1970s and late 1990s, is exactly the climate oscillation noted earlier by Dyer and Tyson (1977), confined to the subtropical latitudes south of 15°S and clearest in the band extending from 20 to 30°S across the subcontinent (Tyson et al., 2002), known as the Dyer-Tyson cycle. This is also shown by the weakness of this oscillation in the rainfall time series at Makoholi in Zimbabwe compared to the South African stations.

The range in annual total rainfall far exceeds that contributed by tropical systems from the SWIO. Furthermore, there is no relation between initial landfall along the coast of Mozambique and the rainfall contributed by these systems. For example, during 1962, 1967, 1969 and 1986, tropical cyclones that made landfall moved westward or northwestward following a track too far north to contribute rainfall over the South African stations where the quasi-bidecadal cycle was noted. It can, therefore, not be said that these systems drive the cycle, but rather that altered atmospheric and/or surface conditions during the wet phase of the cycle alters the behaviour of these systems in terms of track and possibly in terms of intensity towards a greater contribution in rainfall over the Limpopo River Basin.

5.2.3 Trends in rainfall contributed by tropical systems over the Limpopo River Basin

The 5-year running mean of the average normalized rainfall contributed by tropical systems from the SWIO at all six stations demonstrates cyclic behaviour (Fig. 5.10) while a linear trend line fitted over the data shows a slight upward trend in rainfall contributed by these systems over the period.

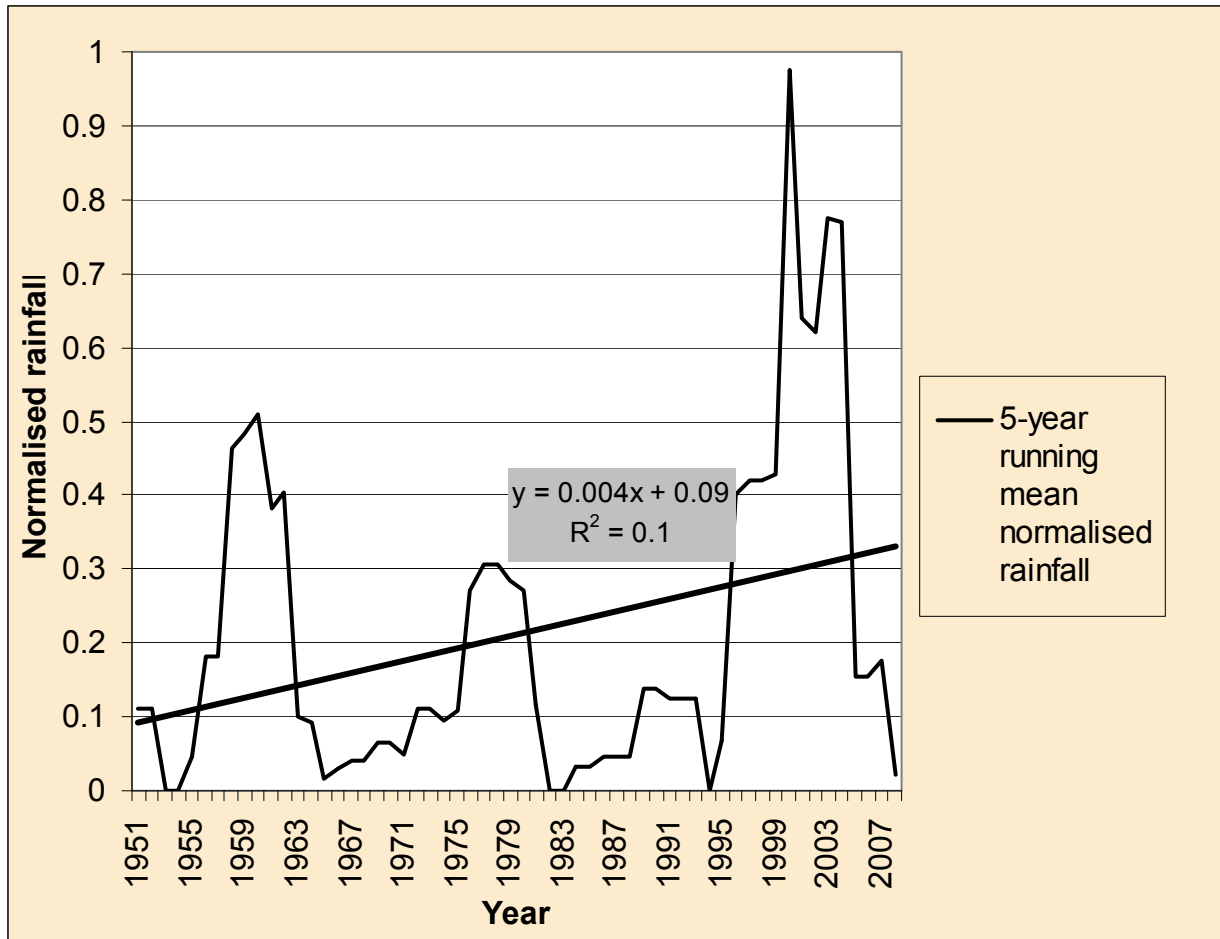


Fig. 5.10: Normalized 5-year average rainfall contributed by tropical systems from the SWIO for all six stations (Fig. 3.1) and a linear trend line fitted to the curve

To minimize the effect of the quasi-bidecadal oscillation, the 20-year moving average (Fig. 5.11) calculated from the average of the normalized rainfall data for the six stations is used for trend analysis by way of the Mann-Kendall test.

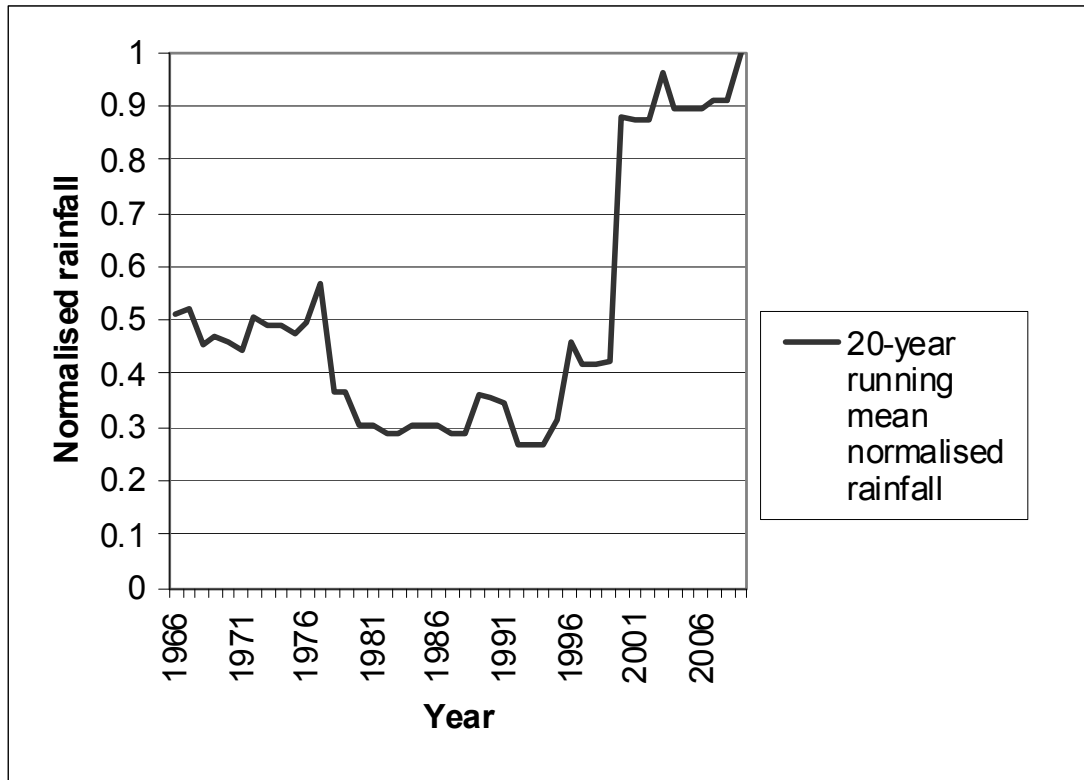


Fig. 5.11: Normalized 20-year average rainfall for all stations

Table 5.3 shows the results of the Mann-Kendall test (Mann, 1945) performed for the 20-year running average rainfall contributed by tropical systems from the SWIO at the stations as indicated.

Table 5.3: Results of the Mann-Kendall test for trend at the stations as indicated

Station	τ	p
Musina	0.23	0.031
Nelspruit	0.07	0.550
Vilanora	0.15	0.169
Makoholi	0.44	0.000
Entabeni	0.35	0.001

The resulting Tau (τ) value can range between -1 and 1 with a larger positive value indicating a stronger upward trend. The p-value indicates the significance of the trend. When subtracted from 1, p gives the confidence associated with the trend. A positive trend exists in the rainfall produced by tropical systems over the stations, but at only three of the stations are these trends significant above the 95% confidence level.

These upward trends are not visible in the total yearly rainfall data over that period. At Entabeni, for example, the Tau-value for annual total rainfall is -0.406 with a p-value of 0.0003, indicating a negative trend in total yearly rainfall rather than the positive trend indicated for rainfall contributed by tropical systems from the SWIO. However, this is not the

case when 5-day heavy rainfall events are considered. At Entabeni, 16 occurrences of rainfall exceeding 400 mm in a 5-day period took place since 1923 (Fig. 5.12). Of these 16 events, 11 occurred in the 42 years after 1966 while only 6 occurred in the 42 years prior to 1966. Tropical systems from the SWIO were responsible for 8 out of the 12 events (about 55%) since 1948. From rainfall data analysis (interpretation of daily rainfall maps for these events – as described earlier) it seems possible that 4 out of the 6 events in the 42 years prior to 1966 were caused by well-defined tropical systems from the SWIO. This means that tropical systems from the SWIO were responsible for 4 events during which rainfall at Entabeni exceeded 400 mm in five days during the 42 years up to 1966, and 6 of the events in the 42 years since 1966 showing an increase when the two periods of equal length are compared. This doesn't necessarily mean an increase in the total of tropical systems influencing the region as it could be the result of specific tracks followed or intensity of systems during these two periods.

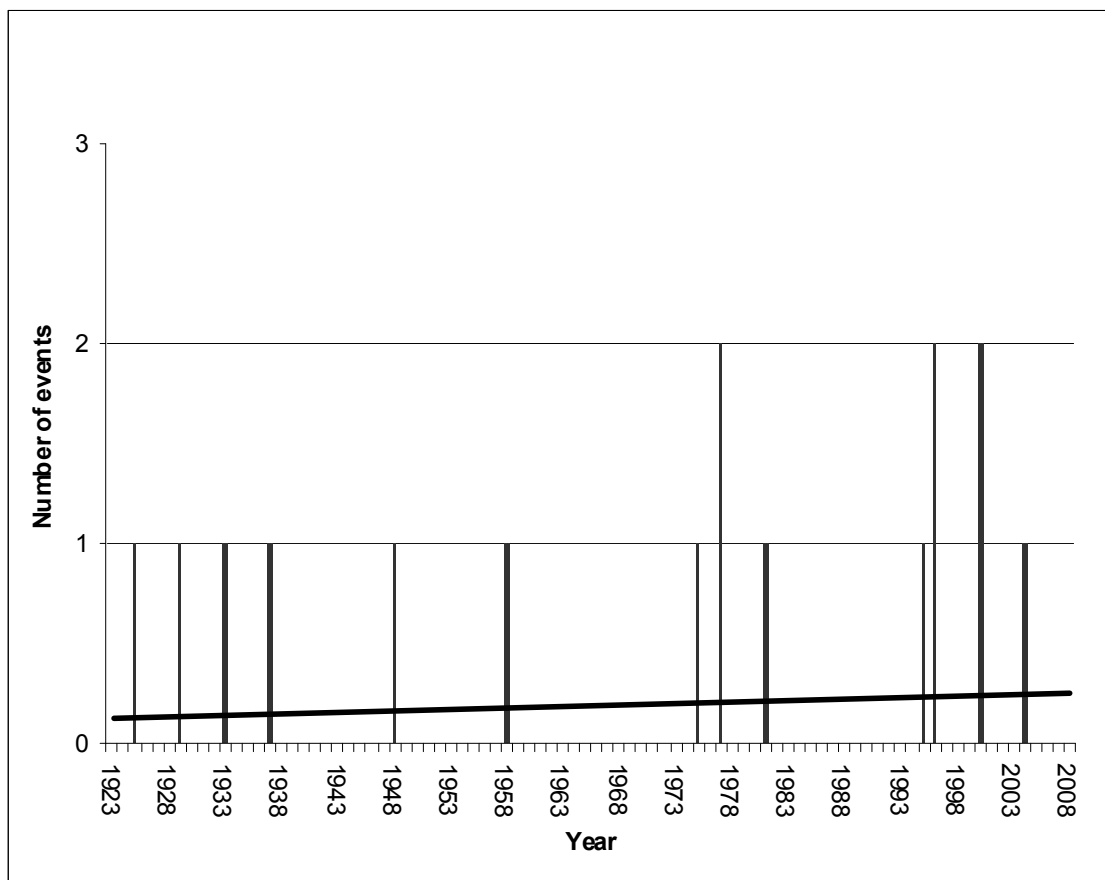


Fig. 5.12: Time series of the yearly amount of events where more than 400 mm was measured within five days at Entabeni since 1923

These findings, together with the divergent results when comparing the trend analysis results when the contribution by tropical systems to rainfall versus the total annual rainfall is considered, suggest that the climatic regime over the last few decades over this region has shifted slightly in favour of tropical systems from the SWIO over the Limpopo River Basin.

Chapter 6

Factors Contributing to Intensity and Extent of Precipitation Associated with Tropical Systems from the SWIO over the Limpopo River Basin

6.1 Introduction

Tropical systems from the SWIO have been associated with several large outbreaks of heavy rainfall over the northeastern parts of South Africa, Mozambique and Zimbabwe over the semi-arid Limpopo River Basin (Malherbe et al., 2012). Heavy rainfall associated with these systems is further enhanced by the orographic effect of the escarpment. In this section, several landfall cases are evaluated focussing specifically on the dynamics of the systems as well as the rainfall produced over the Limpopo River Basin.

Six events of landfalling tropical systems causing rainfall over the Limpopo River Basin have been considered. In terms of total average rainfall contributed to the region (specifically, the six stations used in the previous reports of which five are in northeastern South Africa and one is in Zimbabwe), they can be ranked as follows:

- 1) 3-8 February 2000 (Case 1)
- 2) 21-28 February 2000 (Case 2)
- 3) 7-13 February 1996 (Case 3)
- 4) 17-20 February 1995 (Case 4)
- 5) 3-7 March 2003 (Case 5)
- 6) 2-6 March 1999 (Case 6)

Note that cases of tropical cyclones entering further south and influencing only parts of Mpumalanga and KwaZulu-Natal, but not the Limpopo River Basin, were not considered.

The rainfall product used is a satellite-based rainfall estimate described in section 3.1. For atmospheric data, NCEP Reanalysis data downscaled to a 50 km spatial resolution using CCAM as per section 3.2, are used. Fields calculated/obtained from the resulting 50 km data and used include:

- Divergence at 250 hPa
- 850 hPa moisture convergence
- Total column precipitable water content (PWC)
- Total convective available potential energy (CAPE)

6.2 Discussion

6.2.1 3-8 February 2000 (Case 1)

This system was responsible for very extensive rainfall over large parts of the Limpopo River Basin with westward penetration of heavy rain into the eastern parts of Botswana and central parts of the North West Province. Fig. 6.1 shows the rainfall experienced on the 7th.

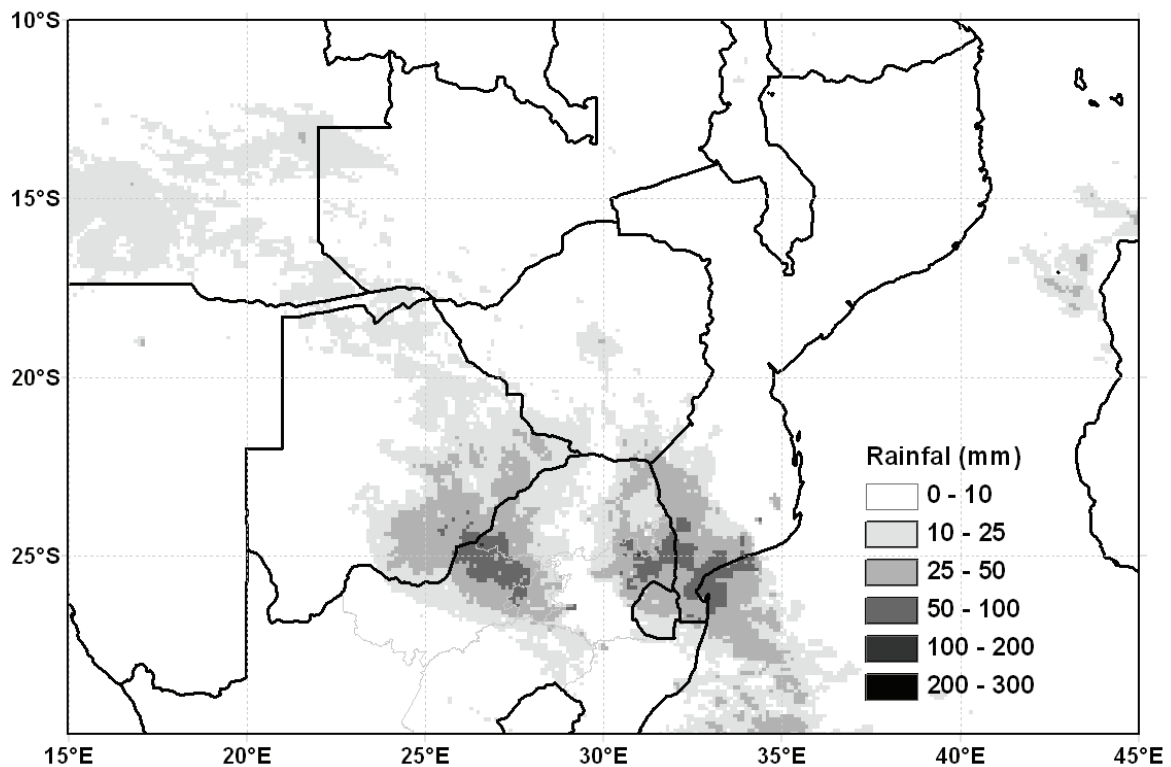


Fig. 6.1: Total rainfall (mm) over southern Africa on 7 February 2000

The tropical depression developed to the east of Mozambique in the Mozambique Channel and tracked westward into subcontinent. Landfall and interior positions of the system are shown in Fig. 6.2, indicating the 700, 500 and 250 hPa heights for the 3rd and 8th of February. The system was steered westward to the north of the lower and mid-tropospheric ridge lying to the north and east of a trough reflected throughout the considered levels to the southwest of the subcontinent as seen in Fig. 6.2. This movement further occurred within the zonally elongated ridge in the upper air, also to the northeast of a trough in the southwest. From Fig. 6.3 it can be seen that lower level convergence of moisture and upward motion at 500 hPa was present over large parts of the South Africa and Botswana with highest values over eastern Botswana closer to the low as well as further southwest towards central North West Province. The low-level flow was from the northeast over South Africa, from where warm moist air invaded the eastern parts of the country. This further resulted in high PWC as well as high CAPE values (Fig. 6.4) over a large region where precipitation occurred. Upper air divergence (also Fig. 6.4) further enhanced conditions specifically over these regions. It therefore seems that all the elements considered favoured widespread and sometimes heavy rainfall over a large part of South Africa and Botswana.

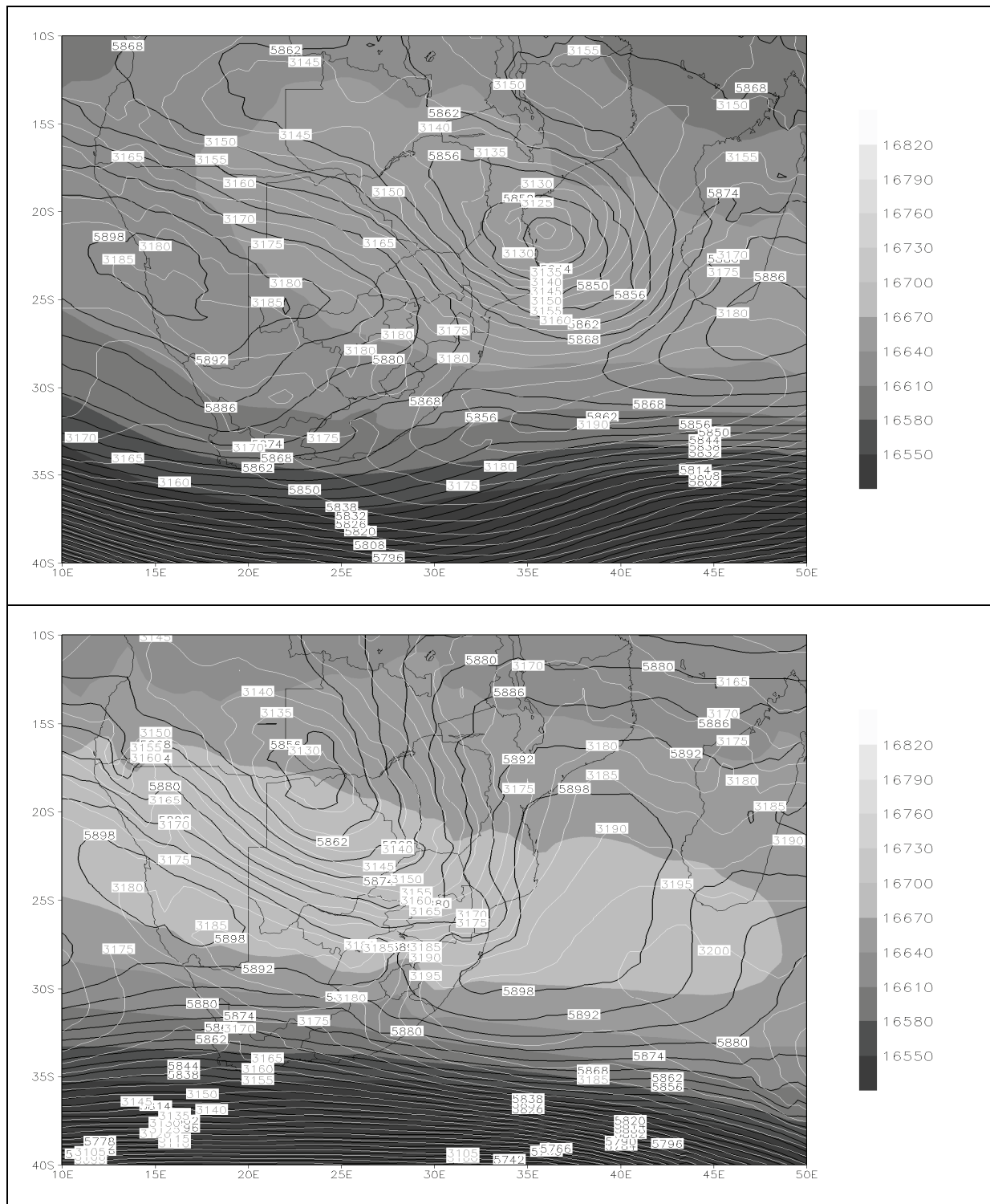


Fig. 6.2: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of a tropical depression on 3 February (top) and 7 February 2000 (bottom)

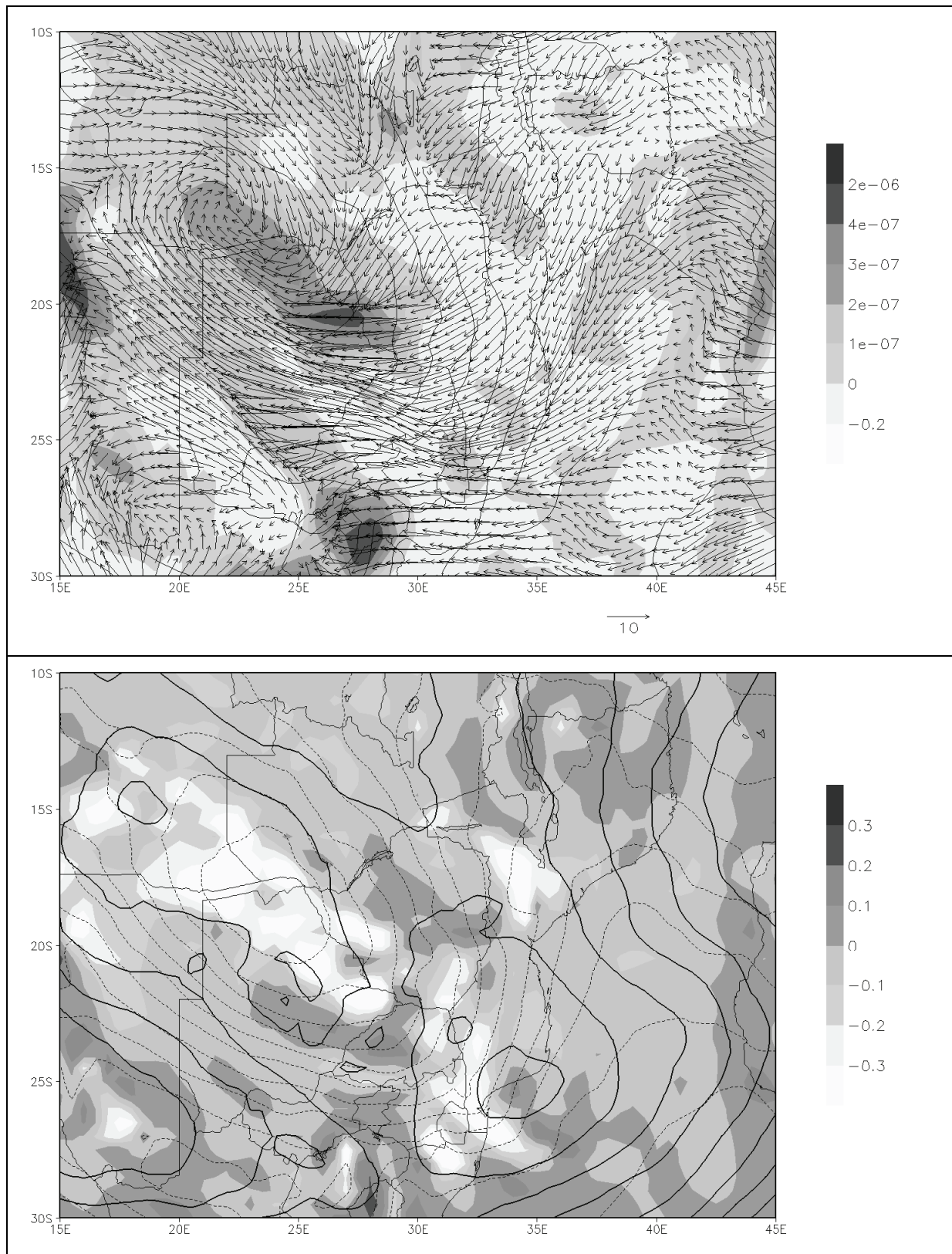


Fig. 6.3: Moisture convergence and wind vectors on 7 February 2000 at 850 hPa (top) as well as vertical velocity (ω – Pa/s) at 500 hPa. The 700 hPa heights are indicated on both maps while the thick solid contours on the bottom map are the 500 hPa heights

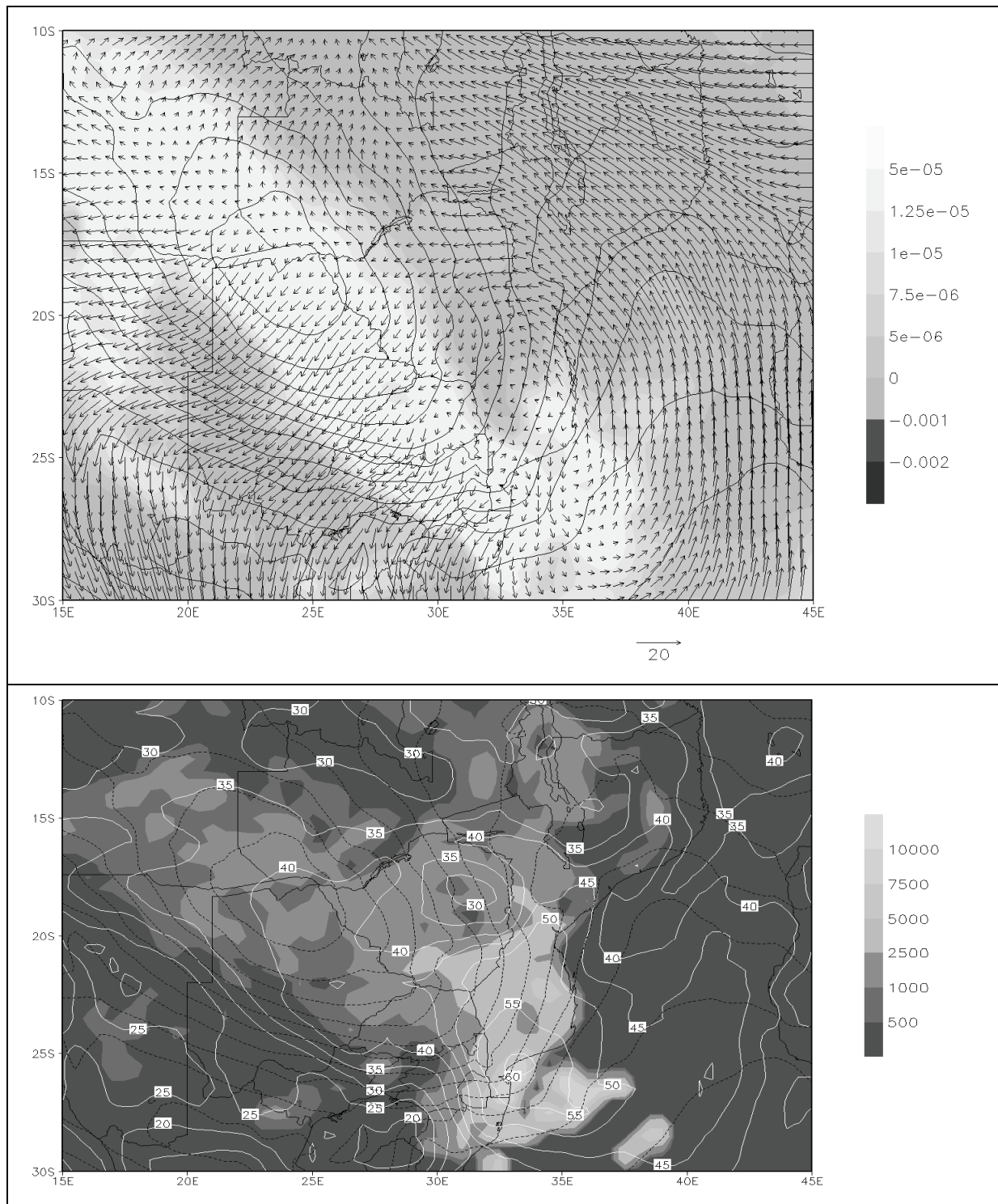


Fig. 6.4: Wind vectors and divergence at 250 hPa (top) for 7 February 2000 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated

6.2.2 21-28 February 2000 (Case 2)

Tropical cyclone Eline was also responsible for widespread, heavy rainfall over the northeastern parts of South Africa in February 2000. However, the extent of rainfall over the interior was smaller than with the tropical depression (Case 1) earlier in the same month. Fig. 6.5 shows the rainfall distribution on the 24th of February 2000.

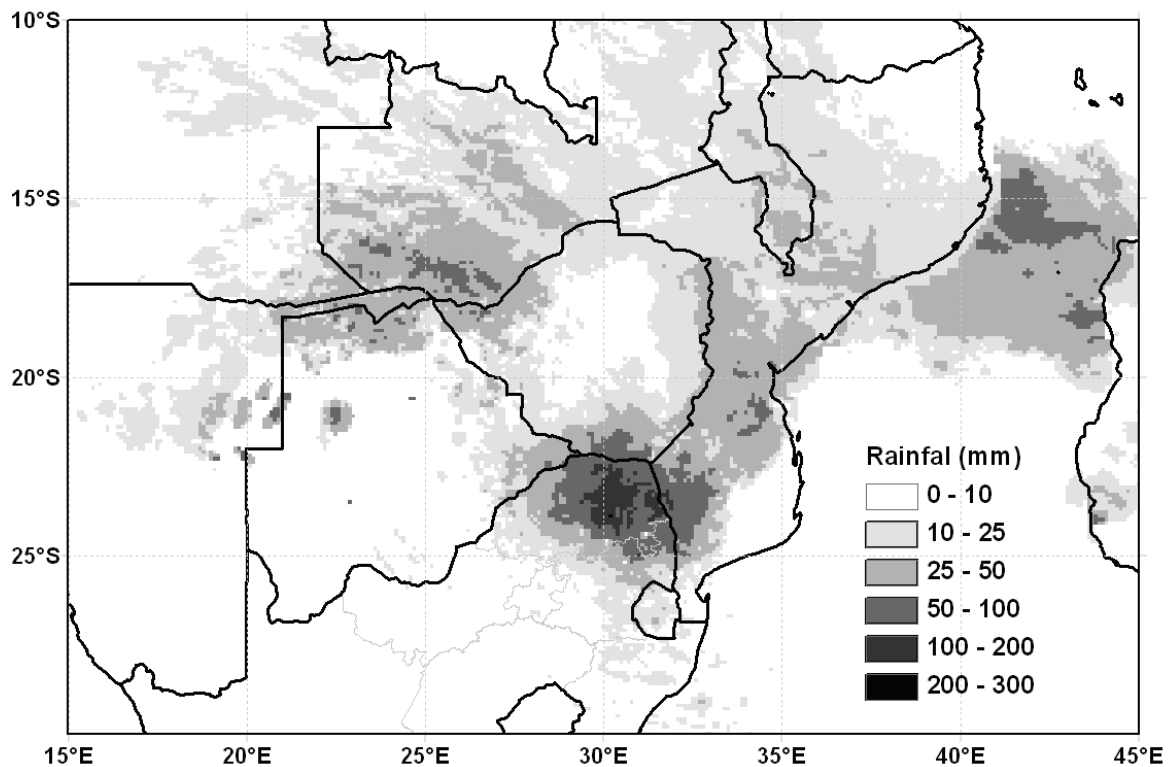


Fig. 6.5: Total rainfall (mm) over southern Africa on 24 February 2000

The movement of tropical cyclone Eline was very similar to the movement described in Case 1 with the geopotential height patterns (Fig. 6.6) also showing the vortex steered westward to the north of a ridge in the lower and middle layers of the atmosphere on the northern edge of an elongated high in the upper air. On the 24th of February, even though the centre of the low was roughly in the same position than the system in Case 1, the trough associated with it was much more east-west orientated, having an effect on the areas where mid-level upward movement was focussed and the flow of low level moisture.

Conditions were also very favourable for heavy precipitation on the 24th. Moisture convergence and upward movement at 500 hPa (Fig. 6.7) was located over much the same areas as during Case 1. However, the flow over the northern parts of South Africa was somewhat more from the south (Fig. 6.7) on this day, still resulting in high PWC amounts over large parts but much lower CAPE values over most of the eastern parts of South Africa (Fig. 6.8). Upper air divergence (Fig. 6.8) and mid-level upward movement (Fig. 6.7) was clearly located somewhat further to the north than during Case 1, together with much lower CAPE values to the south, keeping the rainfall more confined to the north even though precipitation amounts were high.

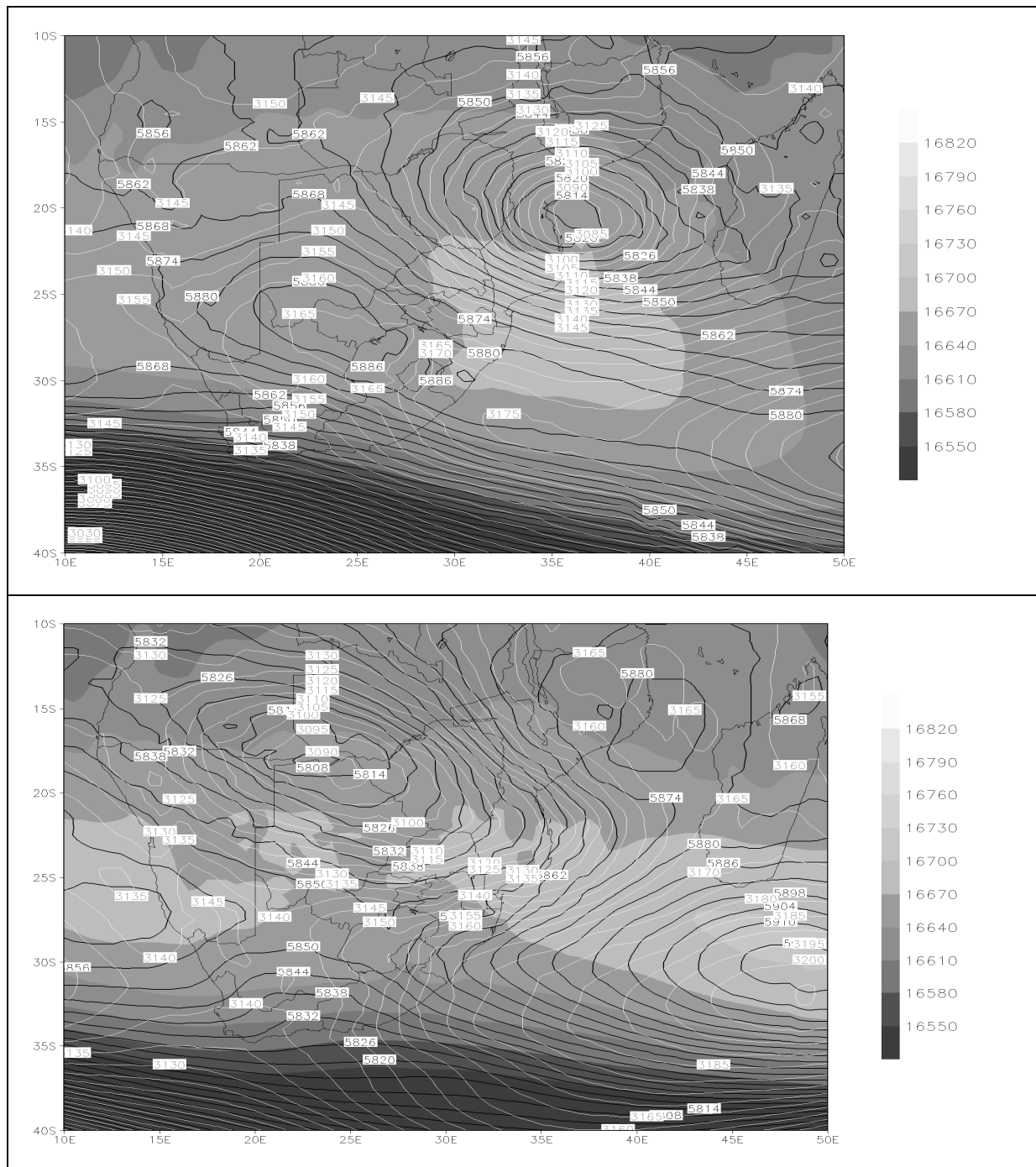


Fig. 6.6: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of tropical cyclone Eline on 19 February (top) and 24 February 2000 (bottom)

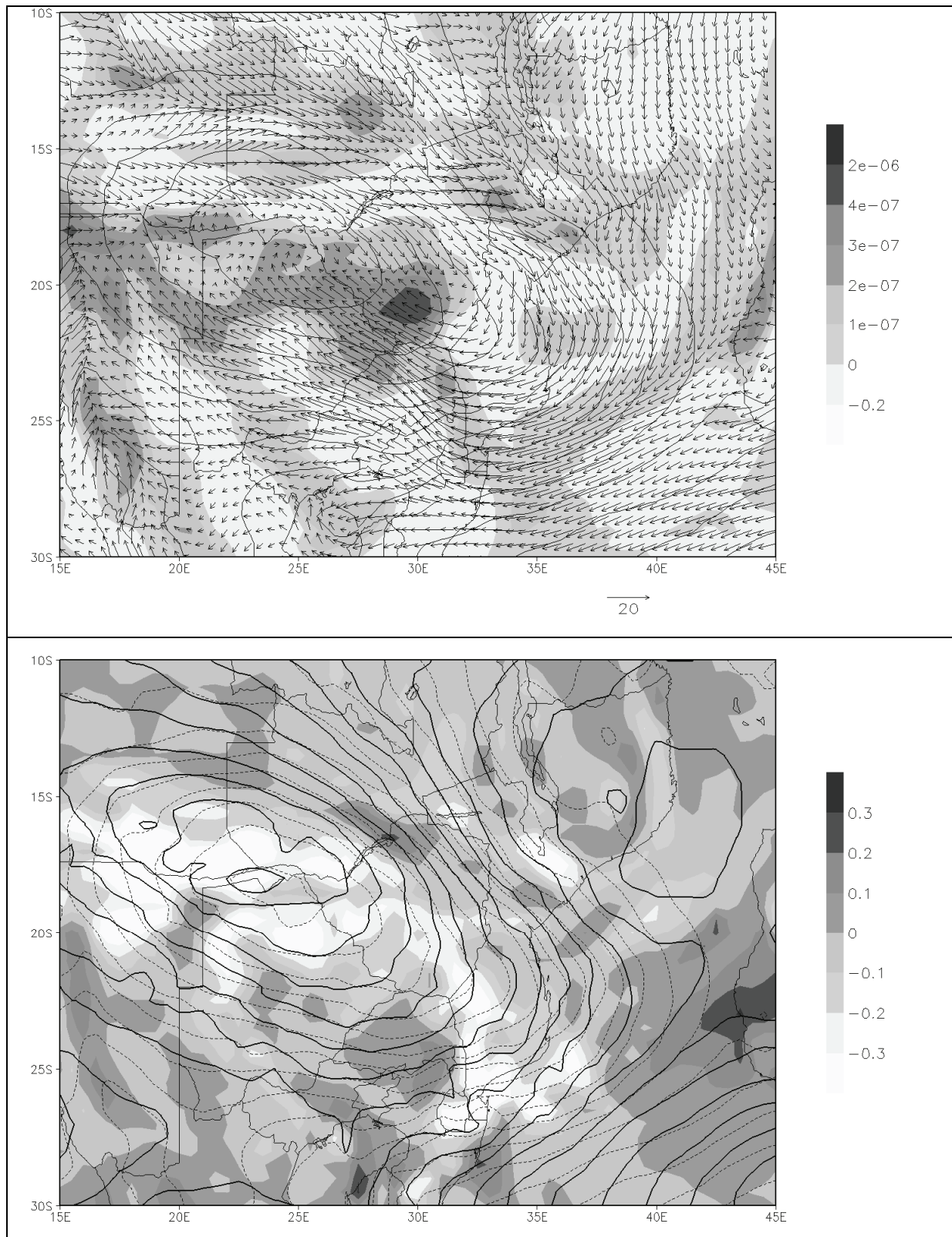


Fig. 6.7: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 24 February 2000. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)

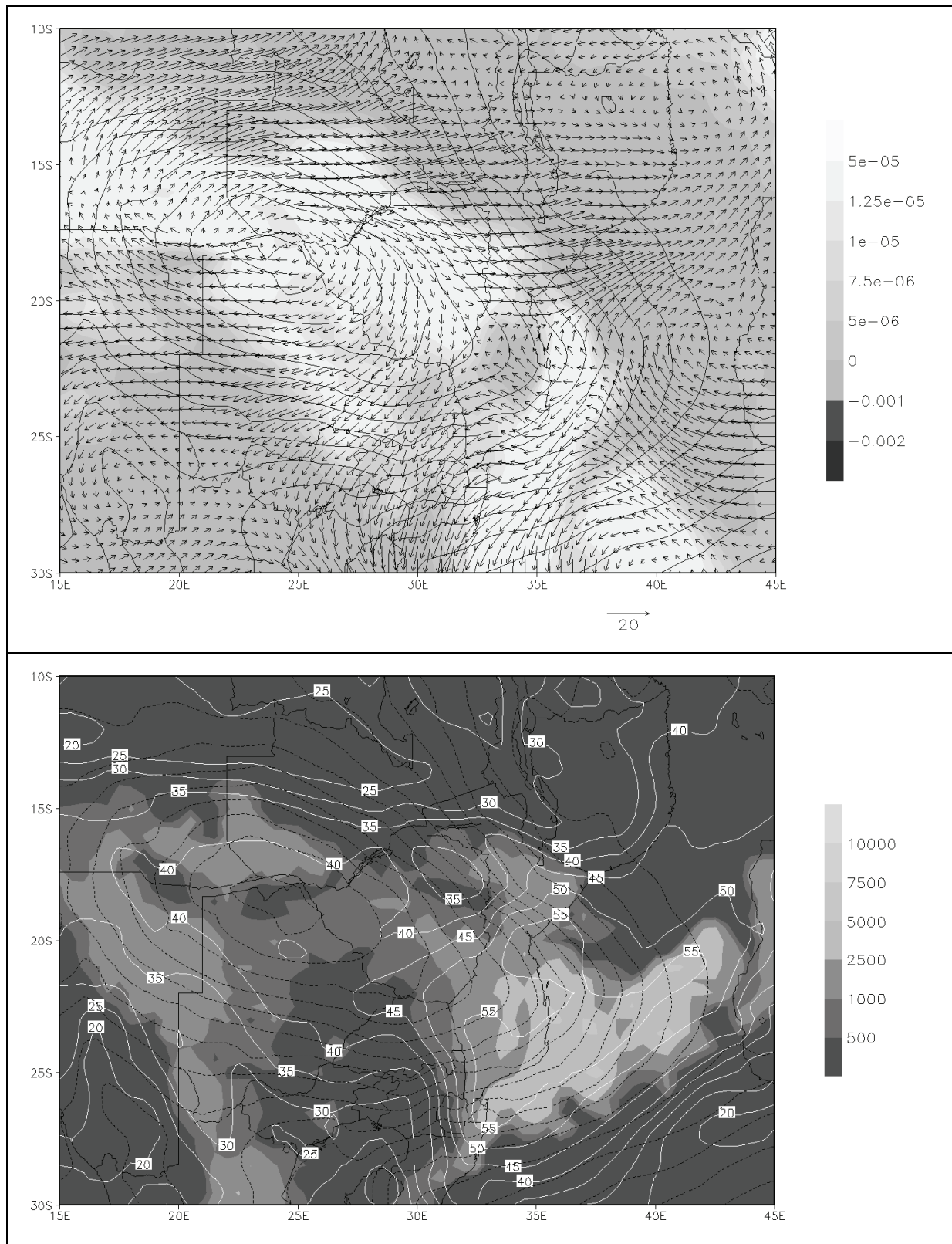


Fig. 6.8: Wind vectors and divergence at 250 hPa (top) for 24 February 2000 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated

6.2.3 7-13 February 1996 (Case 3)

A tropical depression that moved into the subcontinent in February 1996 was responsible for widespread rainfall over the northeastern parts of the country, even before becoming part of a tropical temperate trough as a west wind trough moved in from the west by the 11th. On the 10th a wide area over the northeastern parts of South Africa received rain with heavy falls especially along the escarpment (Fig. 6.9).

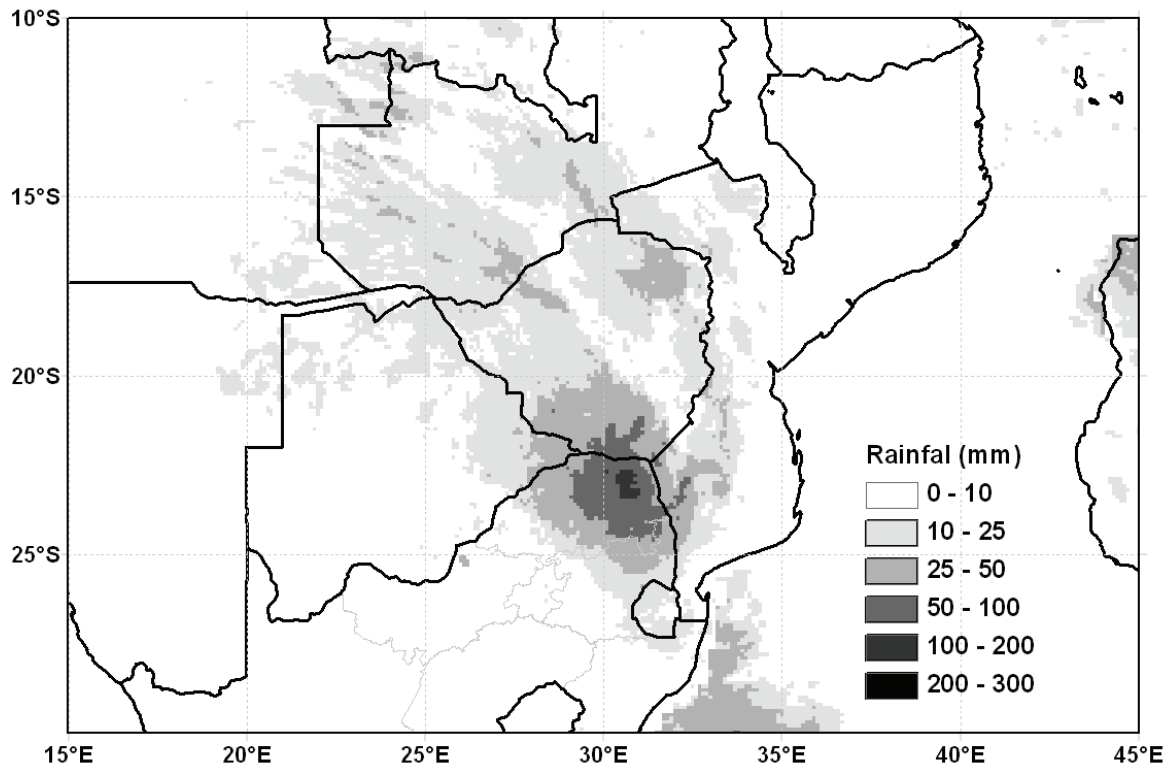


Fig. 6.9: Total rainfall (mm) over southern Africa on 10 February 1996

The system entered much further north than the previous two cases, and continued a westward track toward south/central Zimbabwe (Fig. 6.10). A strong influx of air from the northeast was associated with the system while convergence occurred over large parts of northeastern South Africa (Fig. 6.10). This resulted in high PWC (exceeding 40 mm over the northeastern parts) and high CAPE values (Fig. 6.11) over the same region (higher than during Eline – Case 2). Upper air divergence (Fig. 6.12) was less than during Cases 1 and 2. This is probably related to the fact that this system was much less mature/organized than the previous two when entering the subcontinent. It should also be noted that the high falls around the escarpment were probably enhanced by the high PWC with the flow from the east in the lower layers having its origins over the northern parts of the Mozambique Channel. This onshore flow was also relatively strong due to a steep pressure gradient between the low over Zimbabwe and the Indian Ocean anticyclone.

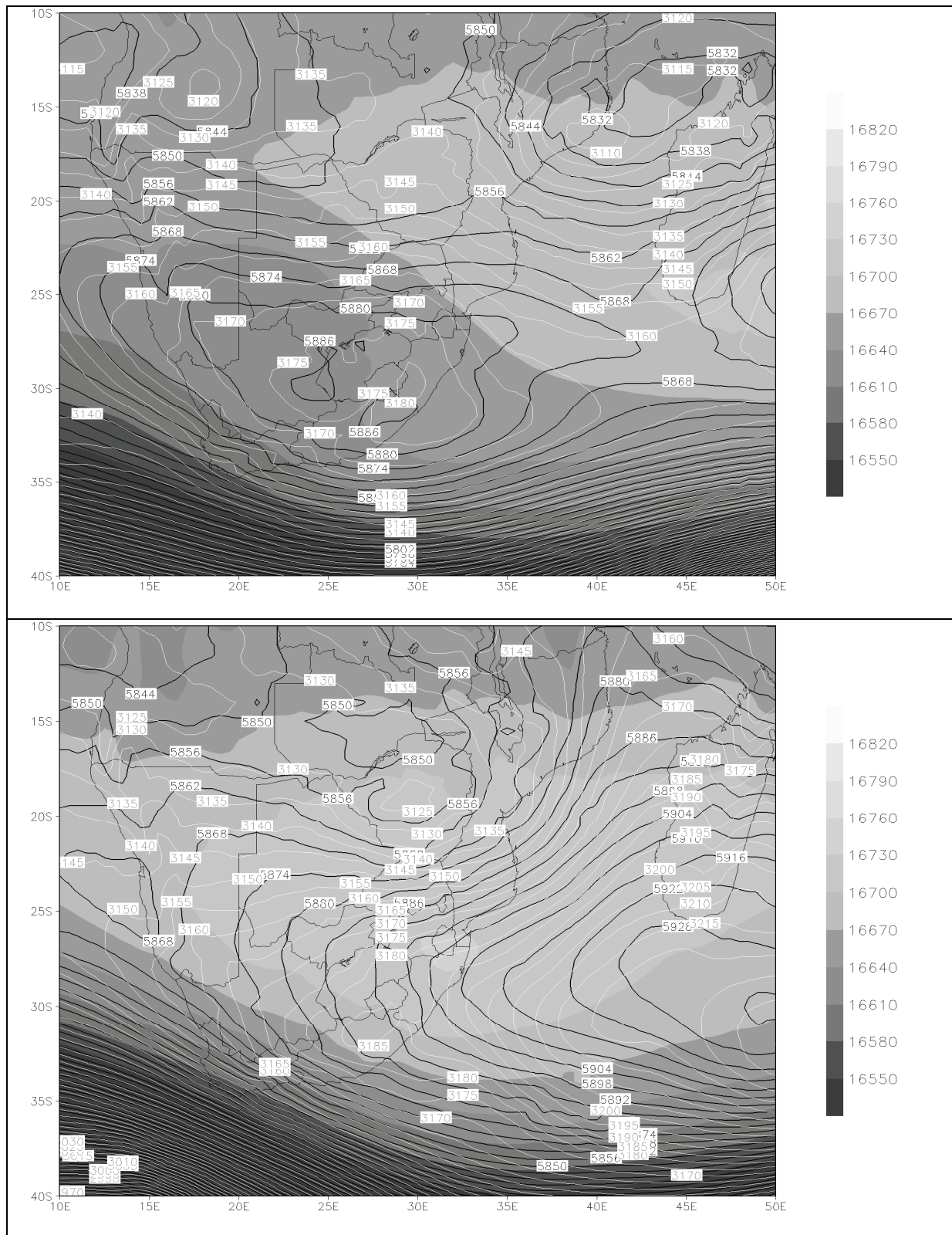


Fig. 6.10: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of a tropical depression on 7 February (top) and 10 February 1996 (bottom)

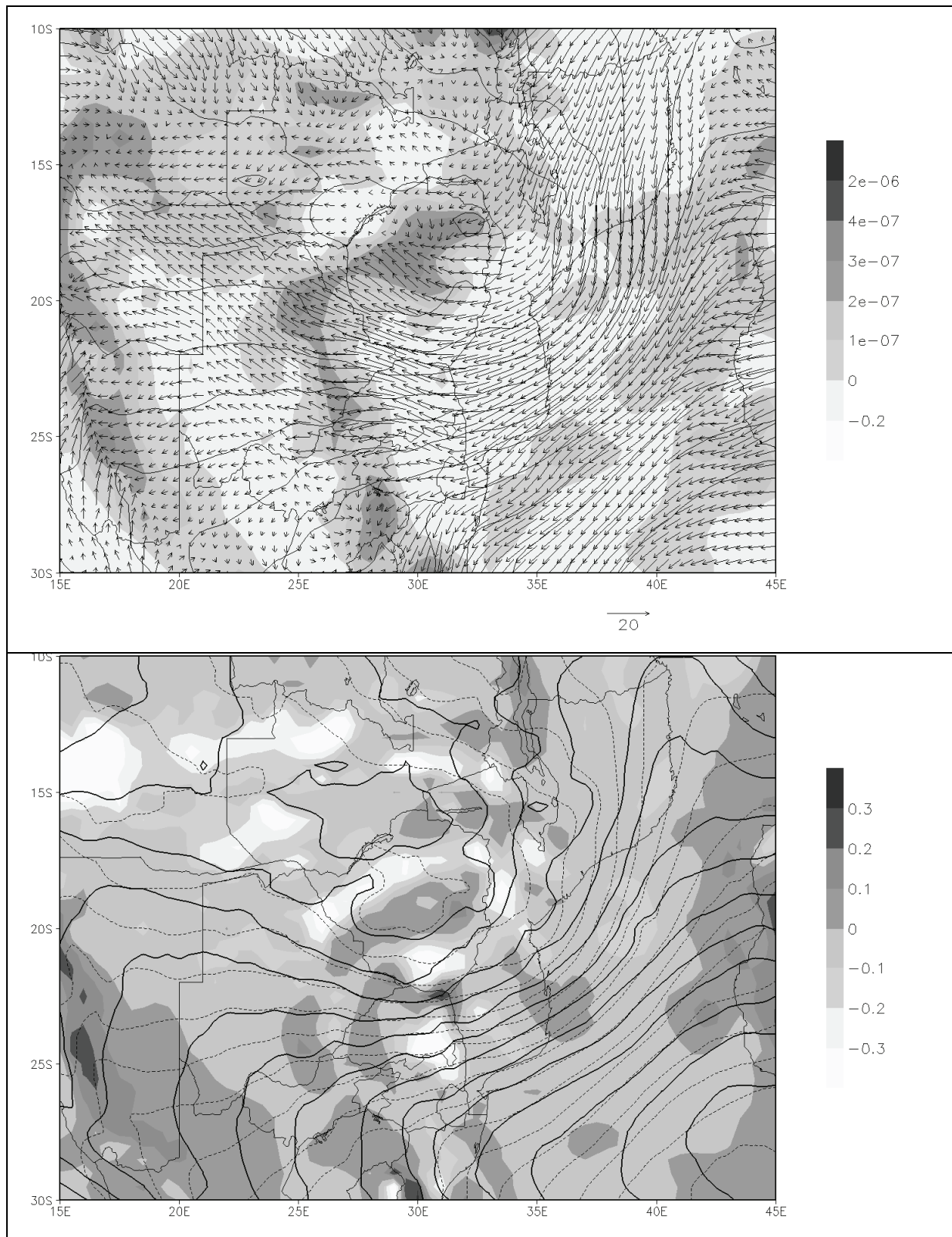


Fig. 6.11: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 10 February 1996. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)

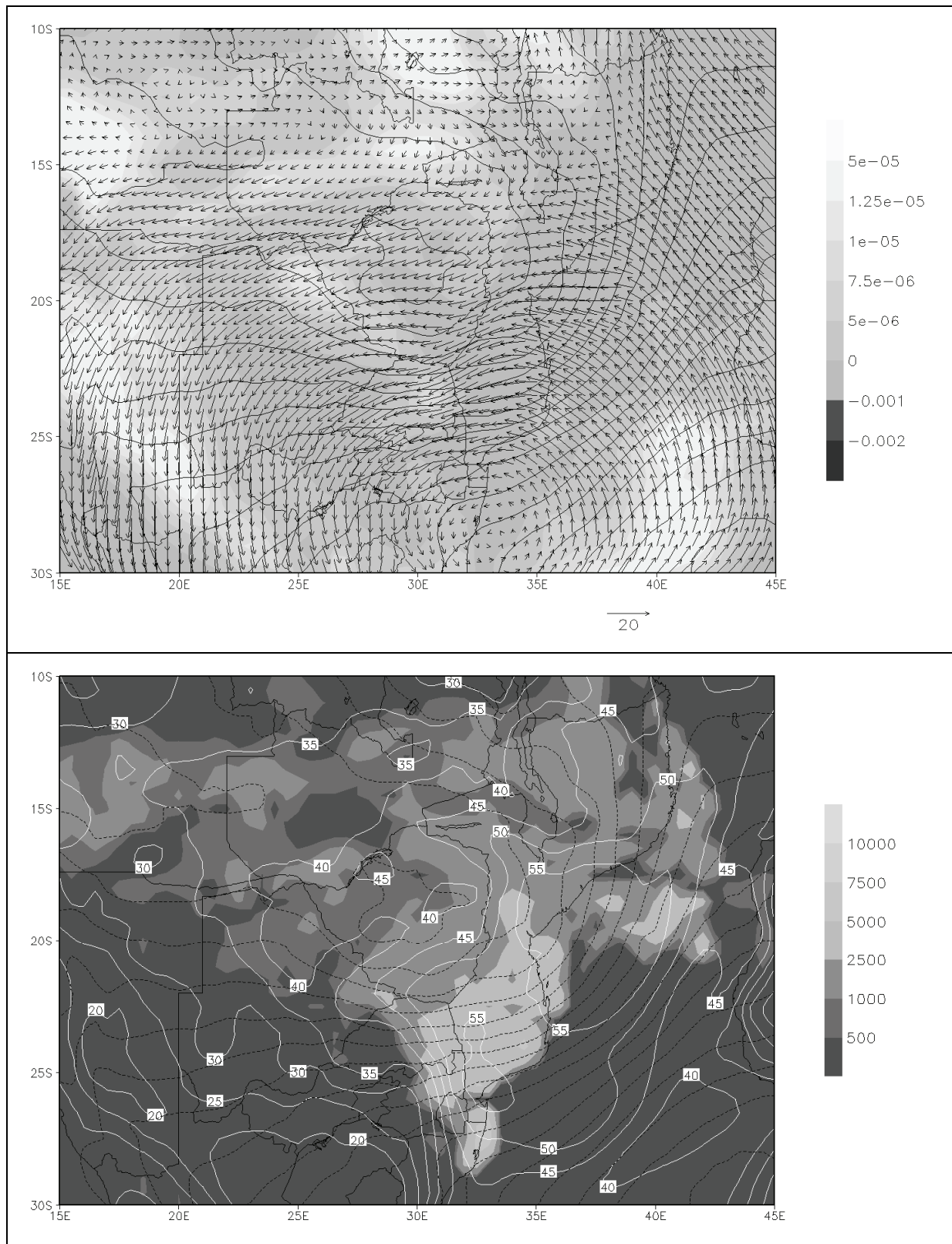


Fig. 6.12: Wind vectors and divergence at 250 hPa (top) for 10 February 1996 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated

6.2.4 17-20 February 1995 (Case 4)

In February 1995 a tropical depression moved into the subcontinent causing widespread rain over much of the northeastern parts of the country. Even though this was a relatively weak system, the track it followed over the subcontinent encouraged widespread rain over parts of South Africa and Botswana. Fig. 6.13 shows the rainfall on the 20th.

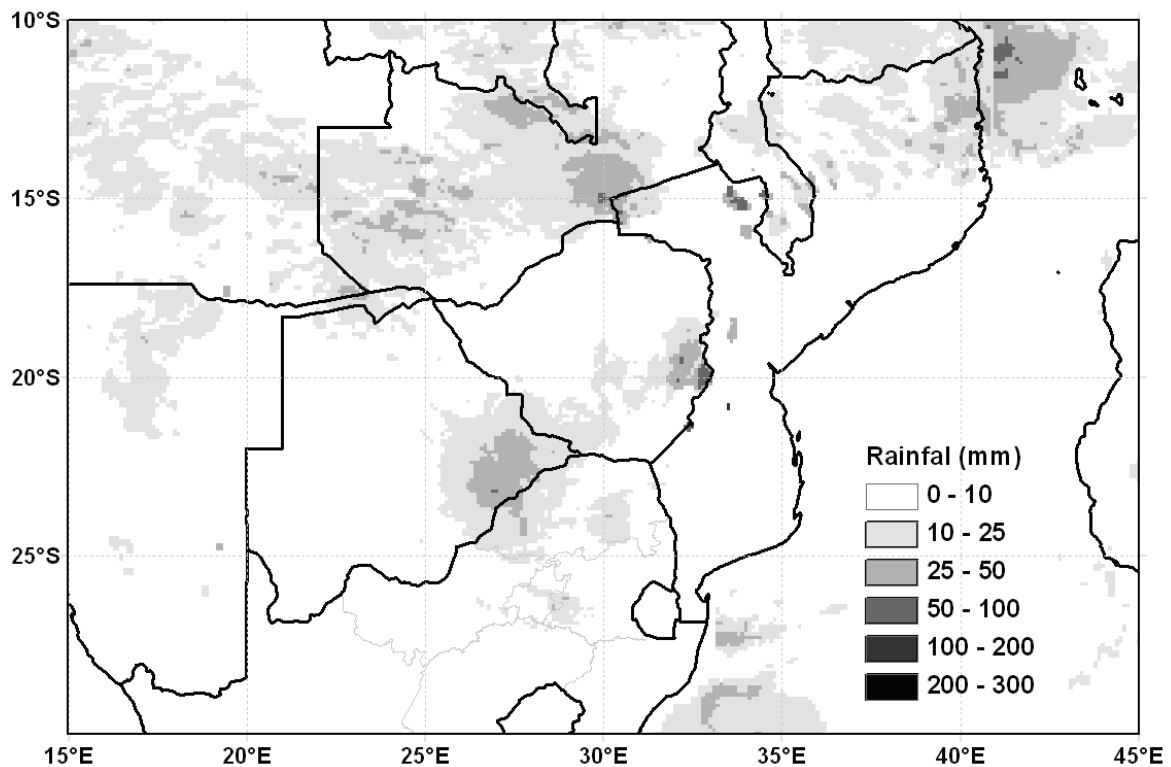


Fig. 6.13: Total rainfall (mm) over southern Africa on 20 February 1995

The system also entered far to the north as in Case 3, but continued a west-southwestward track into the eastern parts of Botswana (Fig. 6.14). Movement was also on the northern edges of the mid-tropospheric subtropical ridge, within the elongated upper high. While surface moisture convergence and strength of the surface flow (Fig. 6.15) as well as upper air divergence (Fig. 6.16) was much weaker than with Cases 1 and 2, the movement of the system into eastern Botswana favoured upward movement at 500 hPa (Fig. 6.15) over large parts of northeastern South Africa as well as the influx of moist air into the area causing high PWC – in excess of 40 mm over large areas (Fig. 6.16). Even though the surface flow associated with this event (Fig. 6.15) was much weaker than with Cases 1-3, the direction was more from the north, further giving rise to high CAPE values and further enhancing the development of thunderstorms.

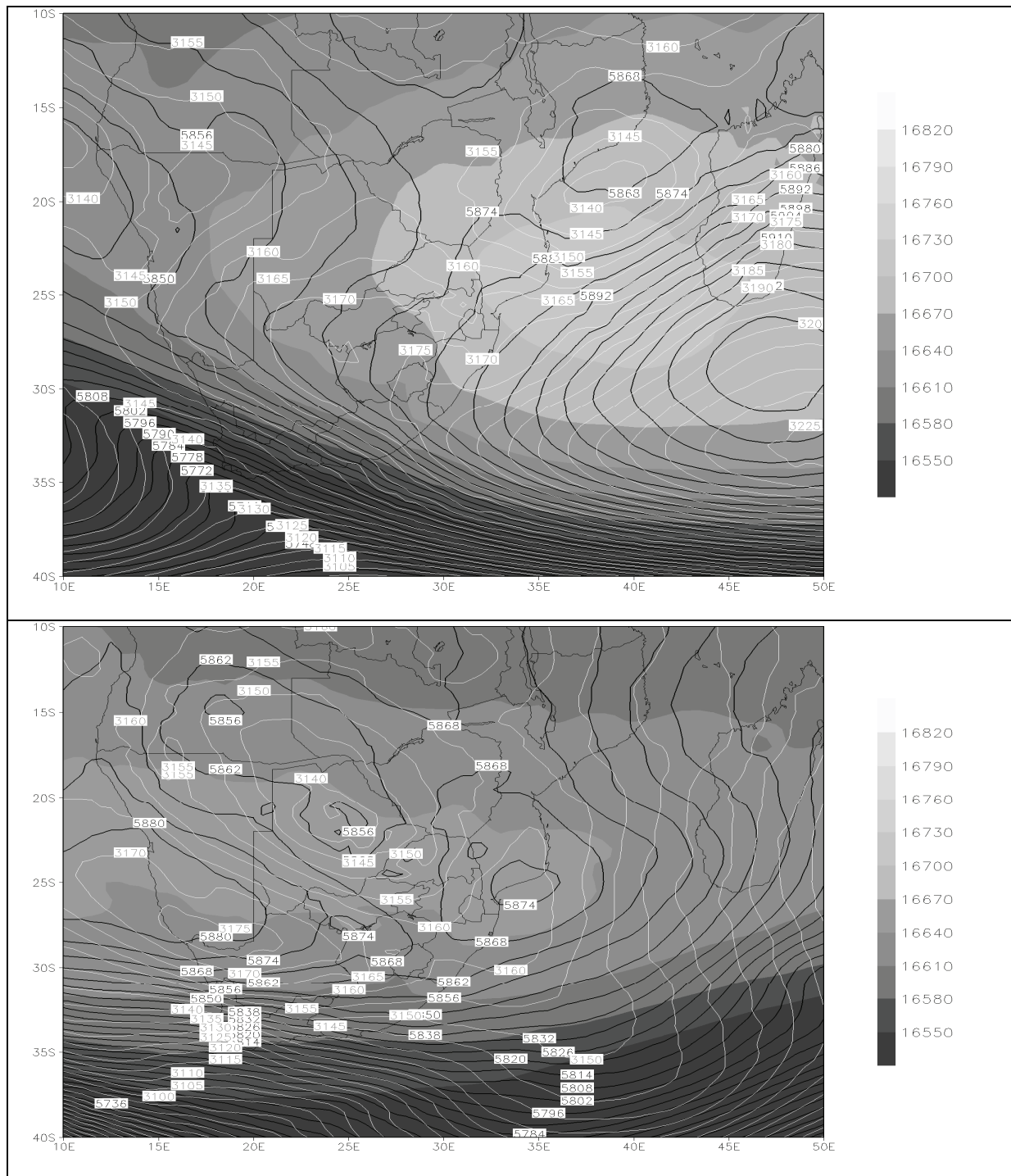


Fig. 6.14: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of tropical of a tropical depression on 16 February (top) and 20 February 1995 (bottom)

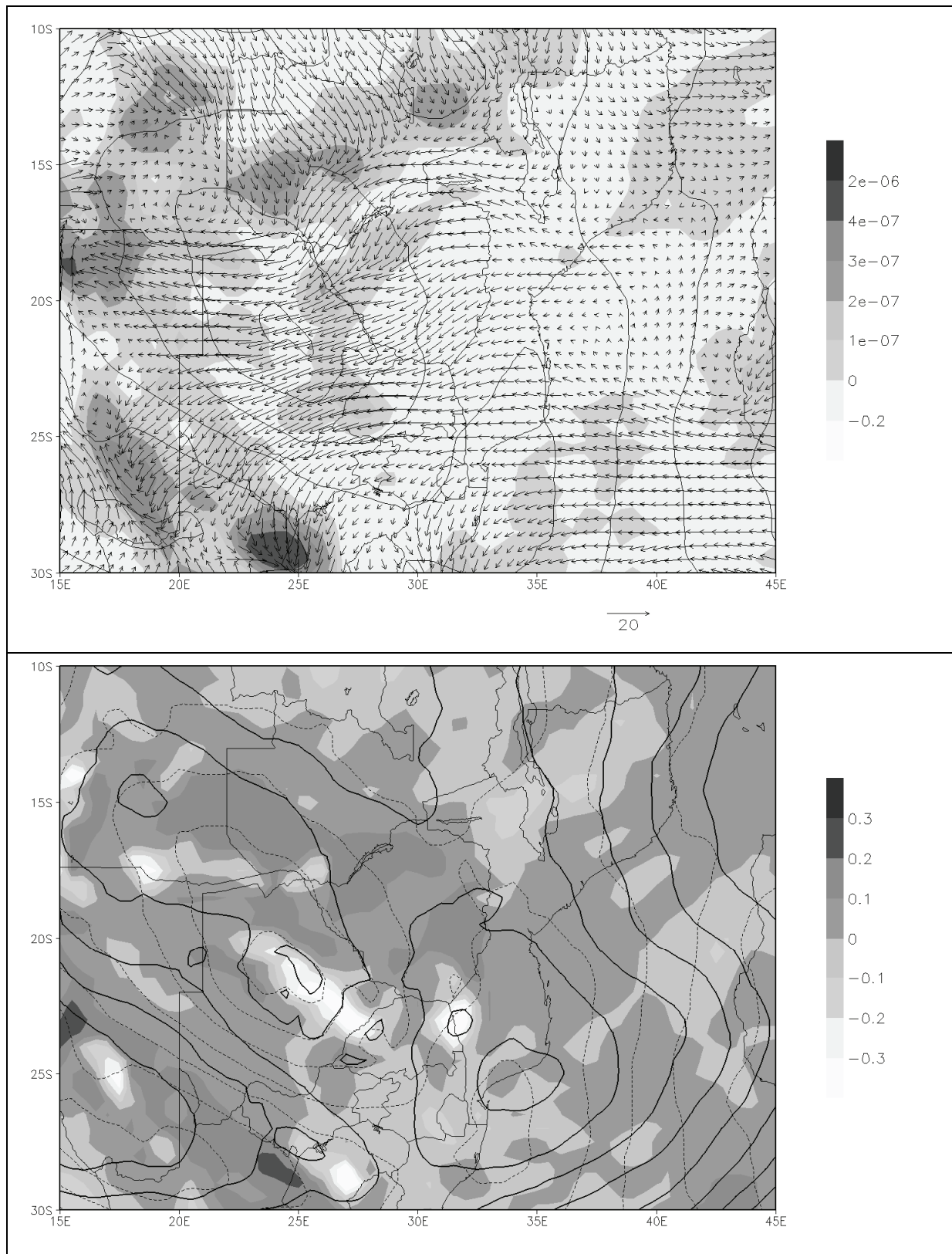


Fig. 6.15: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 20 February 1995. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)

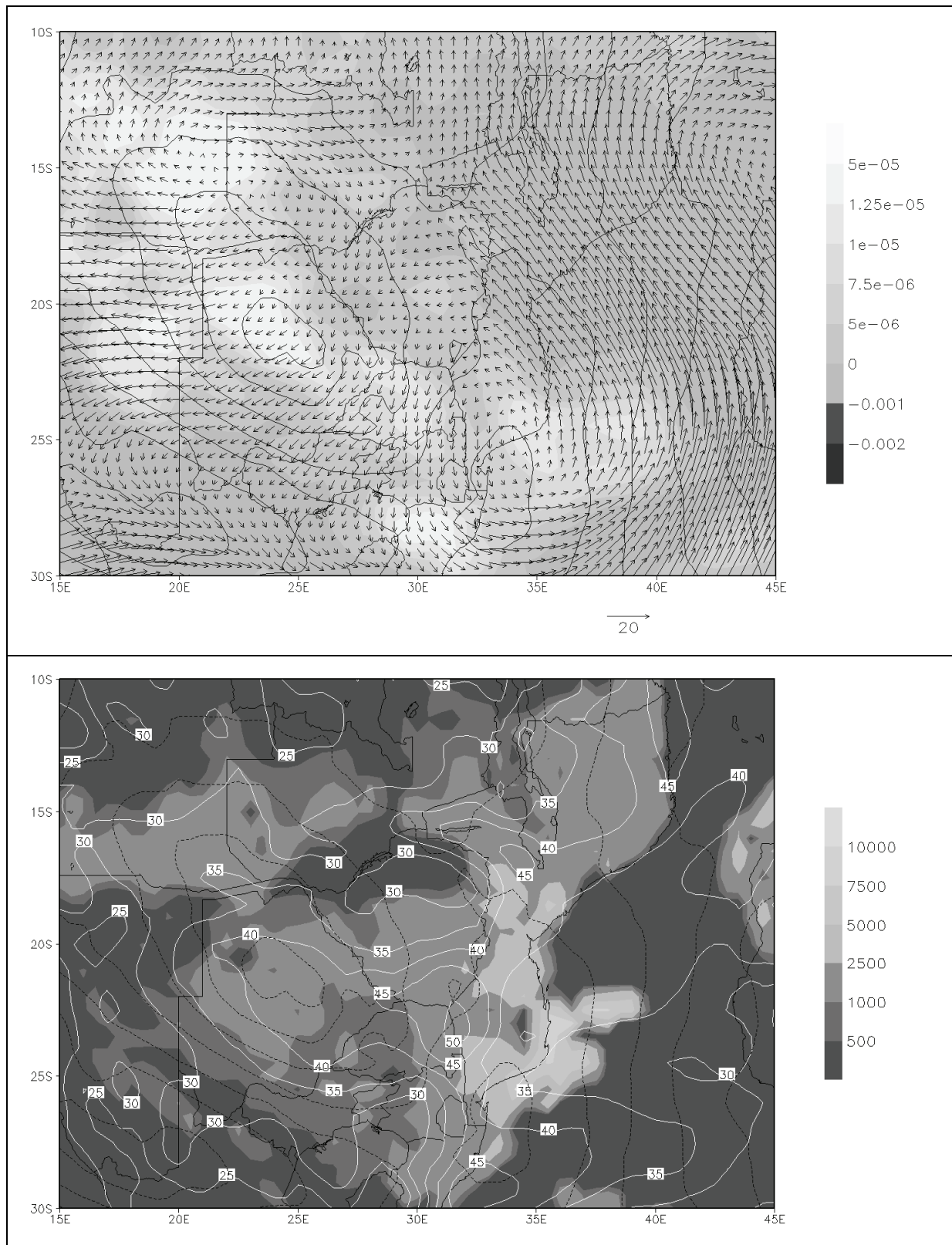


Fig. 6.16: Wind vectors and divergence at 250 hPa (top) for 20 February 1995 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated

6.2.5 3-7 March 2003 (Case 5)

Tropical cyclone Japhet in March 2003 was responsible for heavy rain over much of the areas to the north and east of South Africa. Rainfall over South Africa was, however, low, attributable to the inland track of the system as well as lower level flow over the subcontinent. On the 5th of March, heavy rain fell over southern Zimbabwe and southern Mozambique, with lighter falls over the far northeastern parts of South Africa (Fig. 6.17).

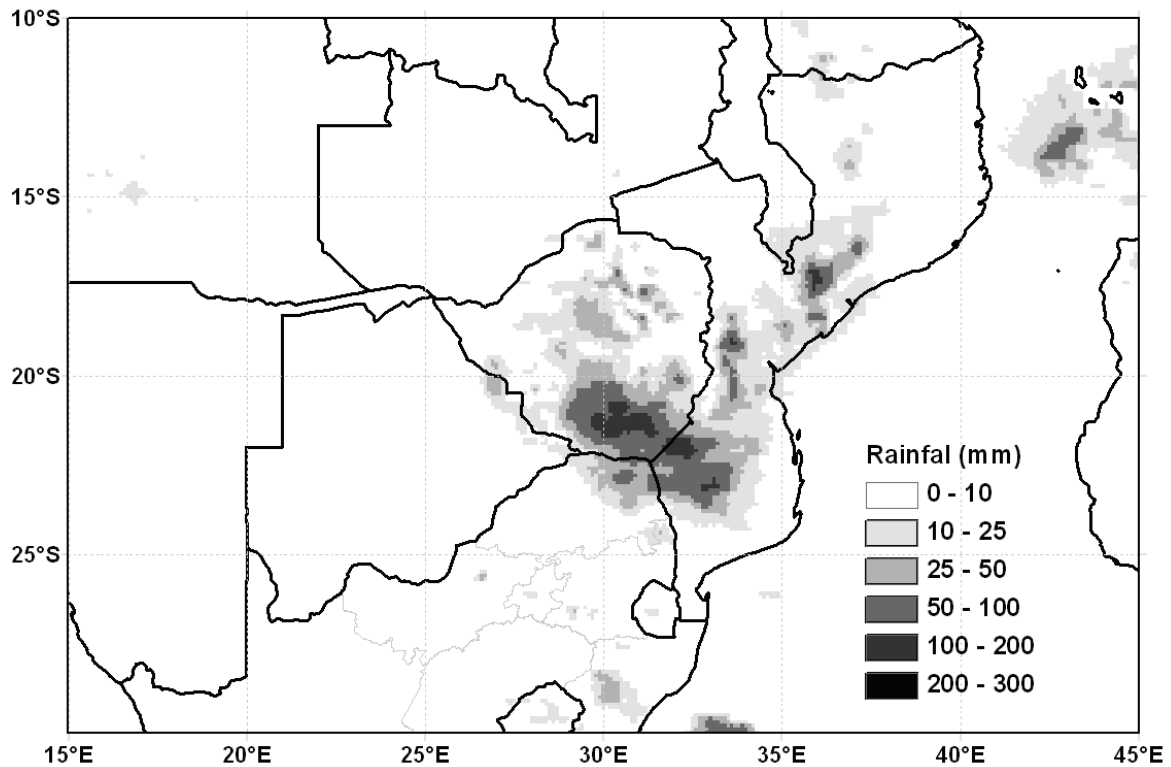


Fig. 6.17: Total rainfall (mm) over southern Africa on 5 March 2003

Even though the system made landfall relatively far to the south (23°S), it moved in a northwesterly direction towards central Zimbabwe on the northeastern side of the high pressure system located over the central to western parts of South Africa whilst also moving away from a developing west wind trough developing towards the southeast (Fig. 6.18).

Figs. 6.19 and 6.20 show that all the elements mentioned earlier that are associated with heavy rainfall after landfall of these systems (upper air divergence, lower level moisture convergence, middle level upward movement, high PWC amounts and high CAPE values) were present but confined to the far northeastern parts of South Africa and the other areas where heavy rain was recorded. Even though there was moisture convergence over the central parts of South Africa, the high PWC amounts (exceeding 40 mm) were all confined to the immediate area surrounding the system in the northeast. Due to the high pressure system towards the west, the surface flow over much of South Africa was much drier than with Cases 1 to 4. The position of the system towards the northeast with the ridge towards the west over South Africa was also not conducive to upward movement at 500 hPa over South Africa (Fig. 6.19).

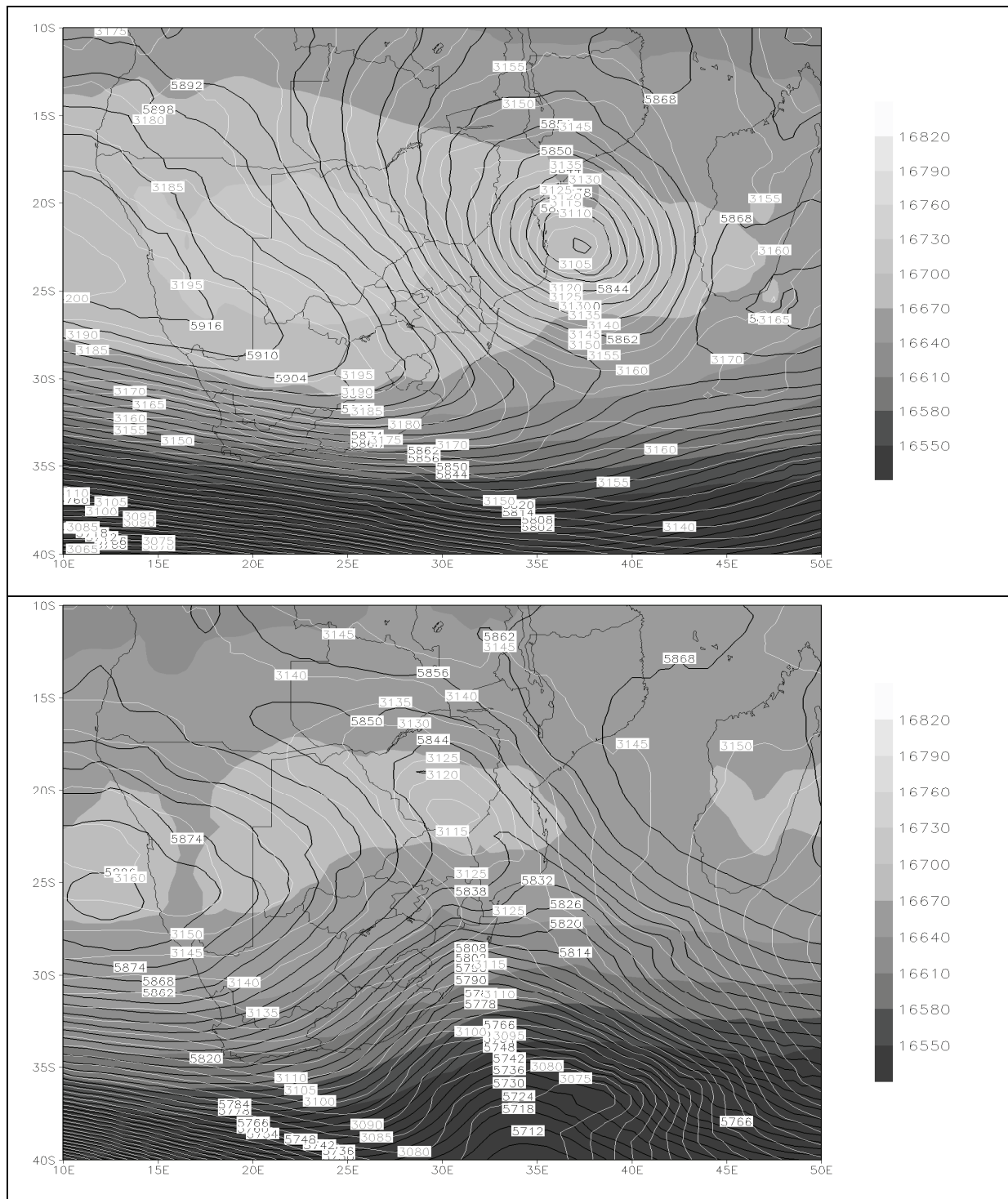


Fig. 6.18: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of tropical cyclone Japhet on 2 March (top) and 5 March 2003 (bottom)

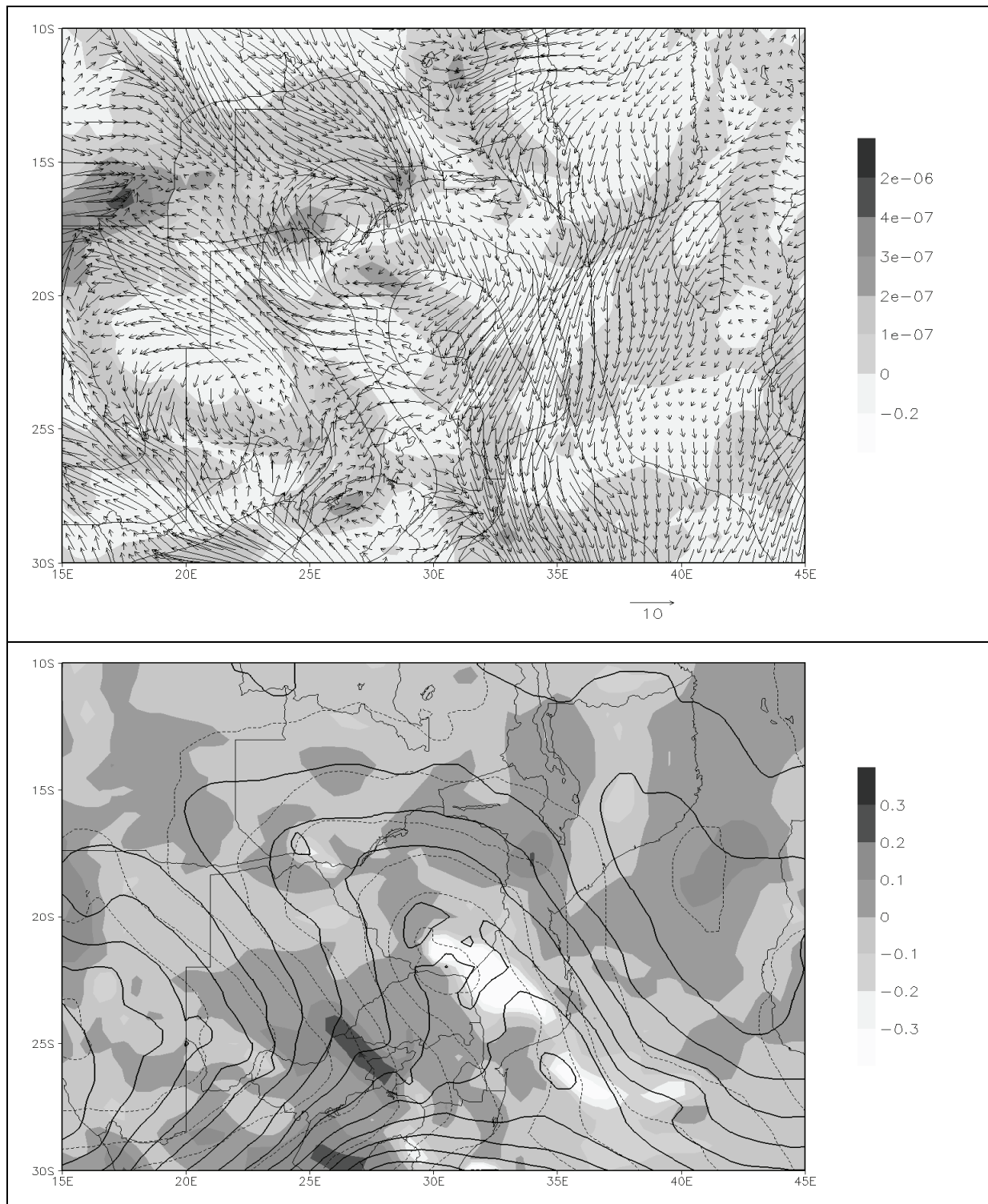


Fig. 6.19: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 5 March 2003. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)

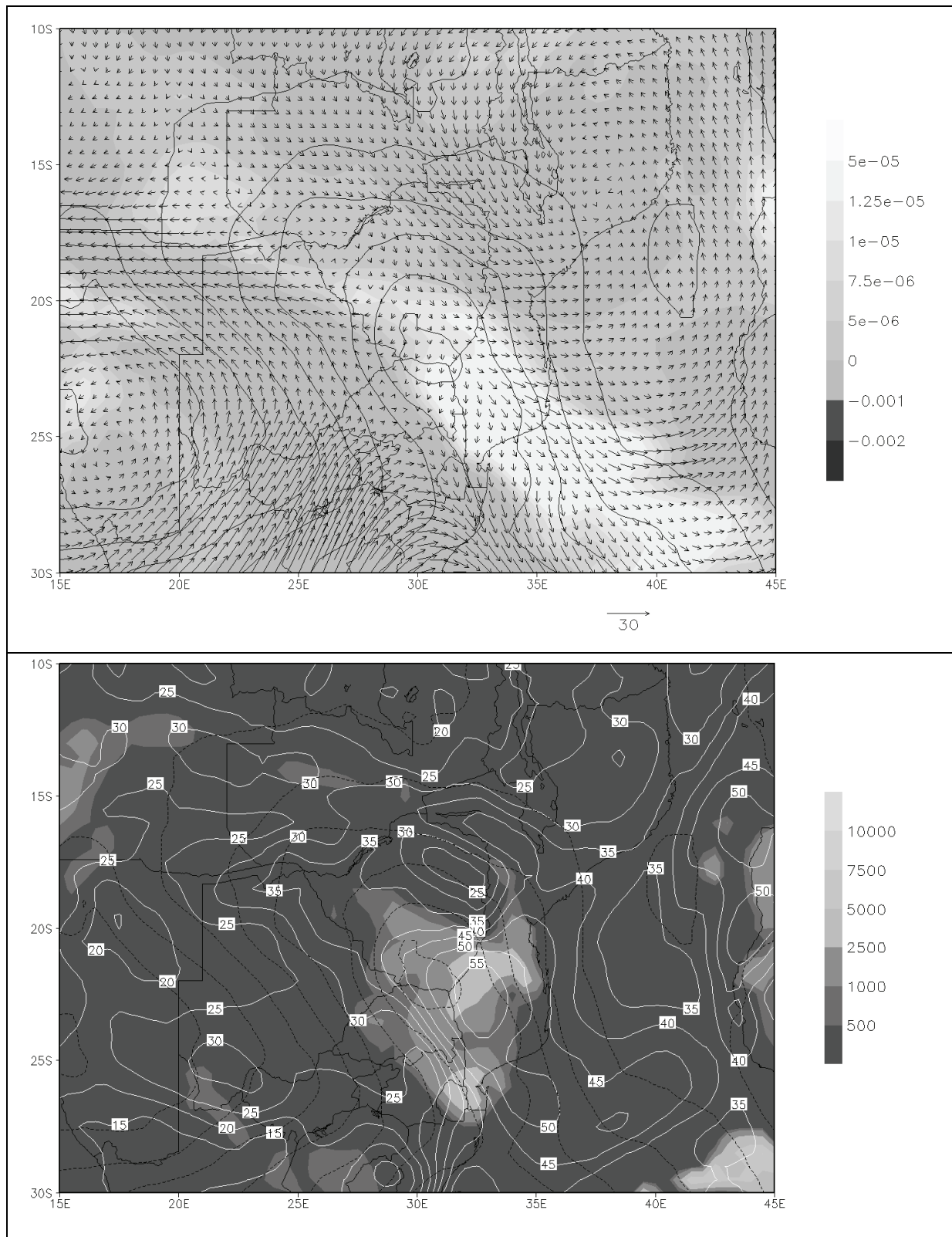


Fig. 6.20: Wind vectors and divergence at 250 hPa (top) for 5 March 2003 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated

6.2.6 2-6 March 1999 (Case 6)

A tropical depression made landfall and moved into the subcontinent between 2 and 6 March 1999. As with Cases 1 to 4, the system was steered westward on the northern side of the lower and middle level subtropical ridge, existing to the northeast of a trough over the southwestern parts of the subcontinent. On the 5th of March showers occurred over some areas in the far northeast of South Africa (Fig. 6.21).

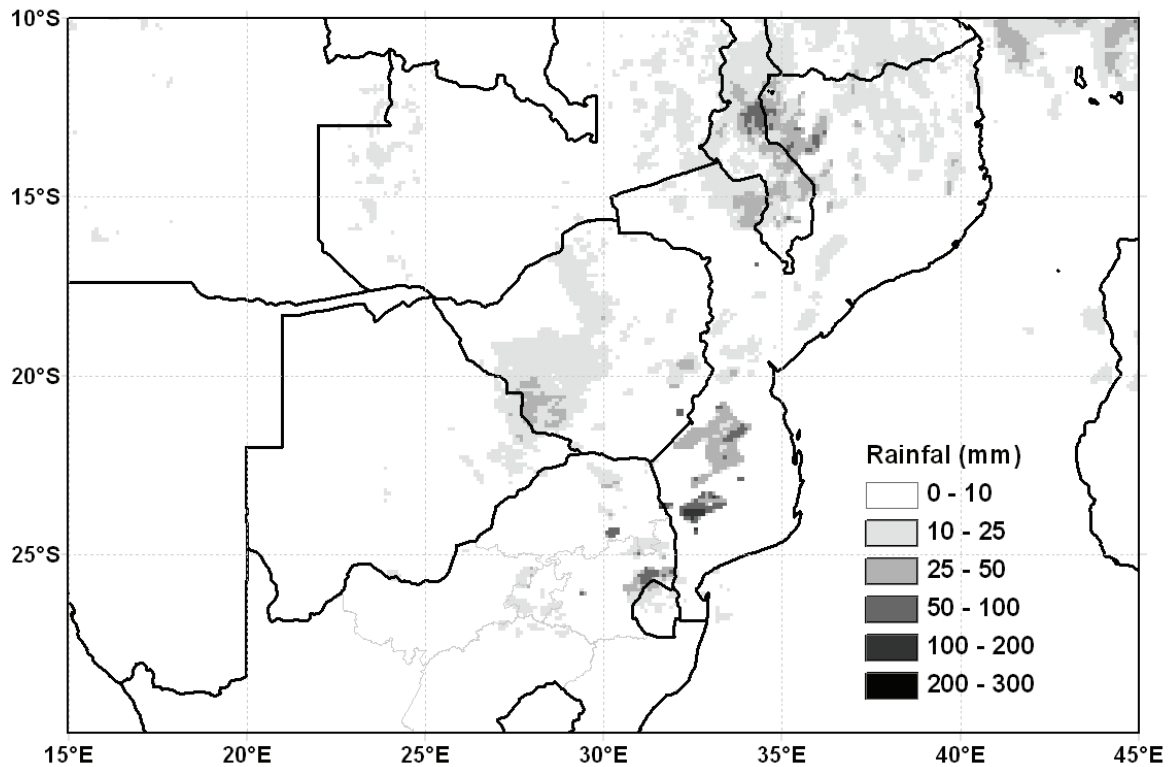


Fig. 6.21: Total rainfall (mm) over southern Africa on 5 March 1999

The system made landfall around 20°S, but due to the strength and orientation of the subtropical ridge towards the southwest, was steered in a northwesterly direction towards northern Zimbabwe (Fig. 6.22). The surface flow, even though relatively strong (Fig. 6.23), was from a more southerly origin than during the other events (Cases 1-5) and the surface flow between the low towards the north and the high towards the southeast was divergent over most of the Limpopo River Basin with surface moisture convergence happening only far to the west over the central parts of South Africa. The surface flow from a more southern origin resulted in lower PWC amounts (only exceeding 40 mm in the extreme east) over most of the region than during the other events while the distribution of high CAPE values was also more restricted than with the other systems (Fig. 6.24). Finally, the position of the low to the north did not contribute to upward motion at the 500 hPa level over most of the Limpopo River Basin (Fig. 6.23) while the 250 hPa flow only diverged over the far northeastern areas (Fig. 6.24).

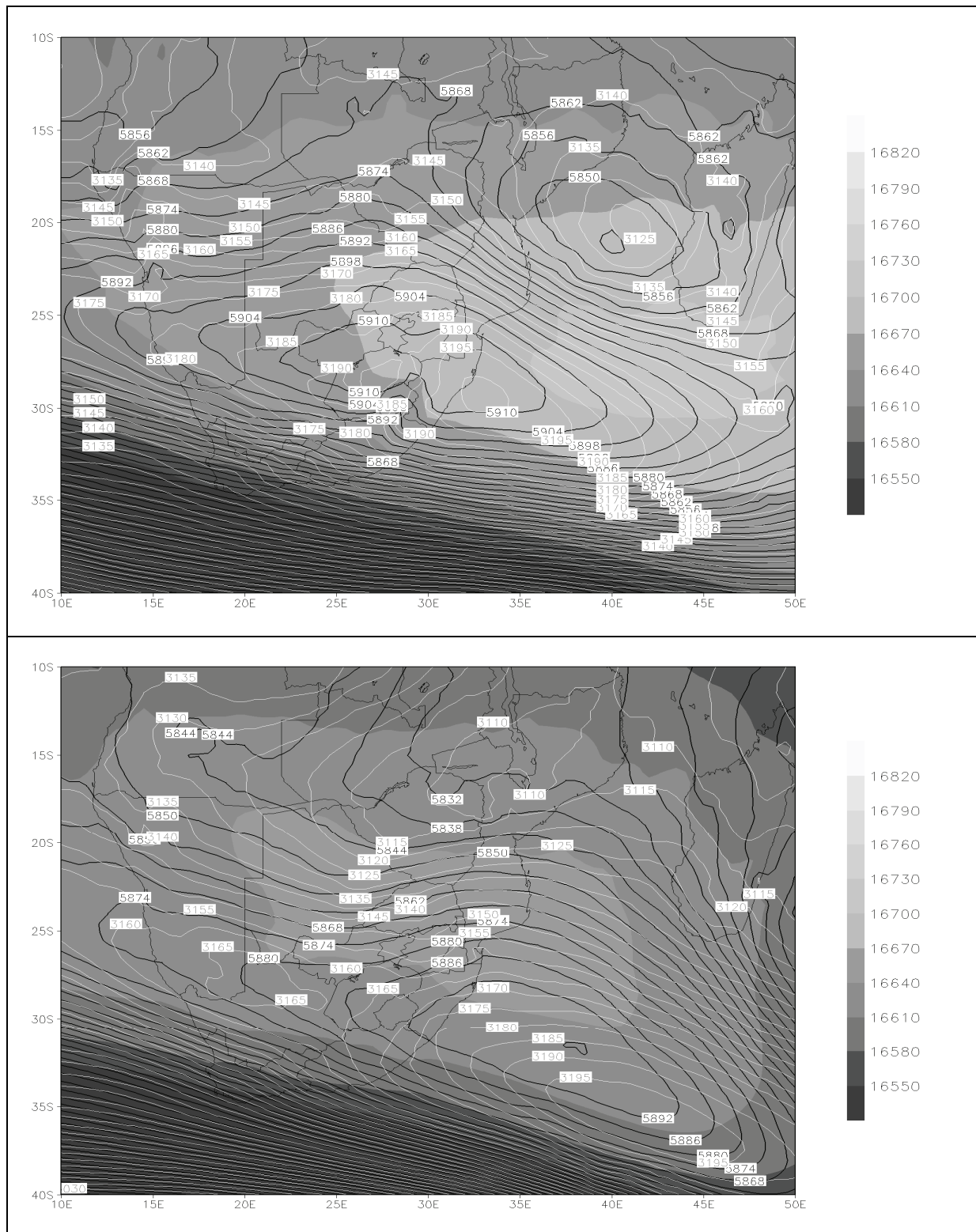


Fig. 6.22: Geopotential height at 700 hPa (white contours), 500 hPa (black contours) and 250 hPa (shades) during the landfall of a tropical depression on 3 March (top) and 6 March 1999 (bottom)

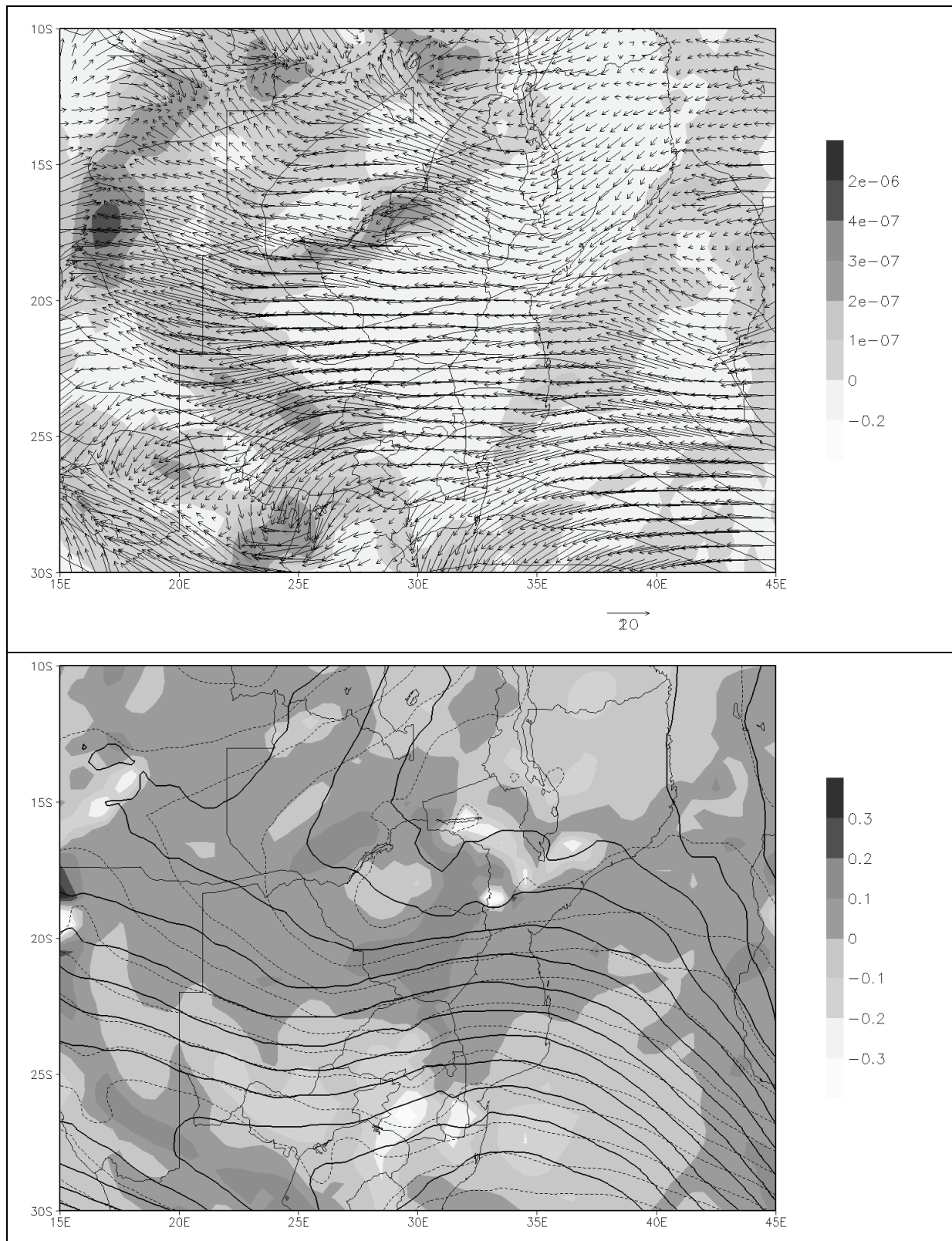


Fig. 6.23: Moisture convergence and wind vectors at 850 hPa (top) and vertical velocity at 500 hPa (bottom) on 6 March 1999. Solid contours are for the 700 hPa level (top) and for the 500 hPa level (bottom)

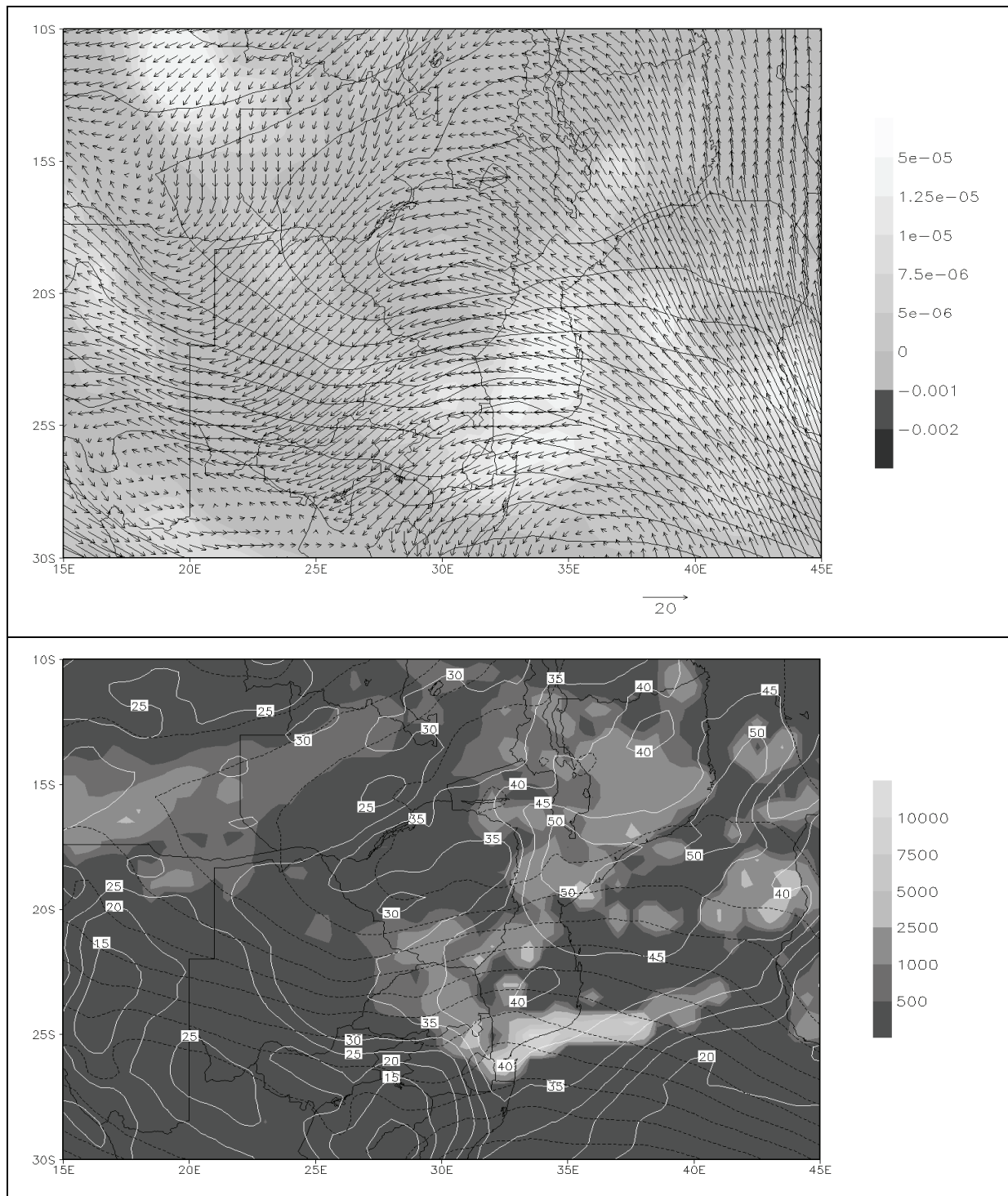


Fig. 6.24: Wind vectors and divergence at 250 hPa (top) for 6 March 1999 as well as total precipitable water content (PWC) in mm (contours, bottom image) and convective available potential energy (CAPE) in J/kg (shades, bottom image). The 700 hPa geopotential heights are also indicated

6.3 Summary

To summarize, from the six cases of landfalling tropical systems considered, a larger rainfall contribution over the Limpopo River Basin can be related to the following:

- Intensity/maturity of the system is related to the organization of convergence and divergence fields as well as extent of precipitation directly associated with the system.
- Surface-flow origin – associated with the system when located over the interior, a flow from the northeast is more conducive to rainfall than a flow from the southeast – related to the moisture and temperature properties of the air mass.
- Surface flow convergence over the Limpopo River Basin.
- Strength of surface onshore flow – a stronger flow is associated with larger and more extensive surface (moisture) advection and moisture convergence and also enhances the orographic effect of the escarpment.
- At 500 hPa, vertical upward motion associated with the system as it moves into the interior can be located somewhat further towards the north depending on the orientation of the trough associated with the system over the interior. A more zonal pattern associated with the trough associated with the inland system over the interior will keep precipitation further north.
- The track after landfall – more zonal track or a track into Botswana as opposed to a northwestward track (or a southwestward track towards southern Mozambique, e.g. Domoina).

Fig. 6.25 shows the tracks of four of the tropical systems that made landfall since the 1960s. Two of these systems (Emily and Eline – squares) were responsible for widespread rain and flooding over the Limpopo River Basin, while the other two (circles) caused rainfall limited only to the east. The nature of the westward track over the interior of southern Africa therefore seems to be related to the quasi bi-decadal oscillation as mentioned in Ch. 5. Atmospheric and/or surface conditions leading to increased rainfall over the Limpopo River Basin seem to alter the preferred tracks of tropical systems from the SWIO. Other properties (such as the convergence/divergence fields, etc.) associated with these systems probably also vary according to this multi-year variation.

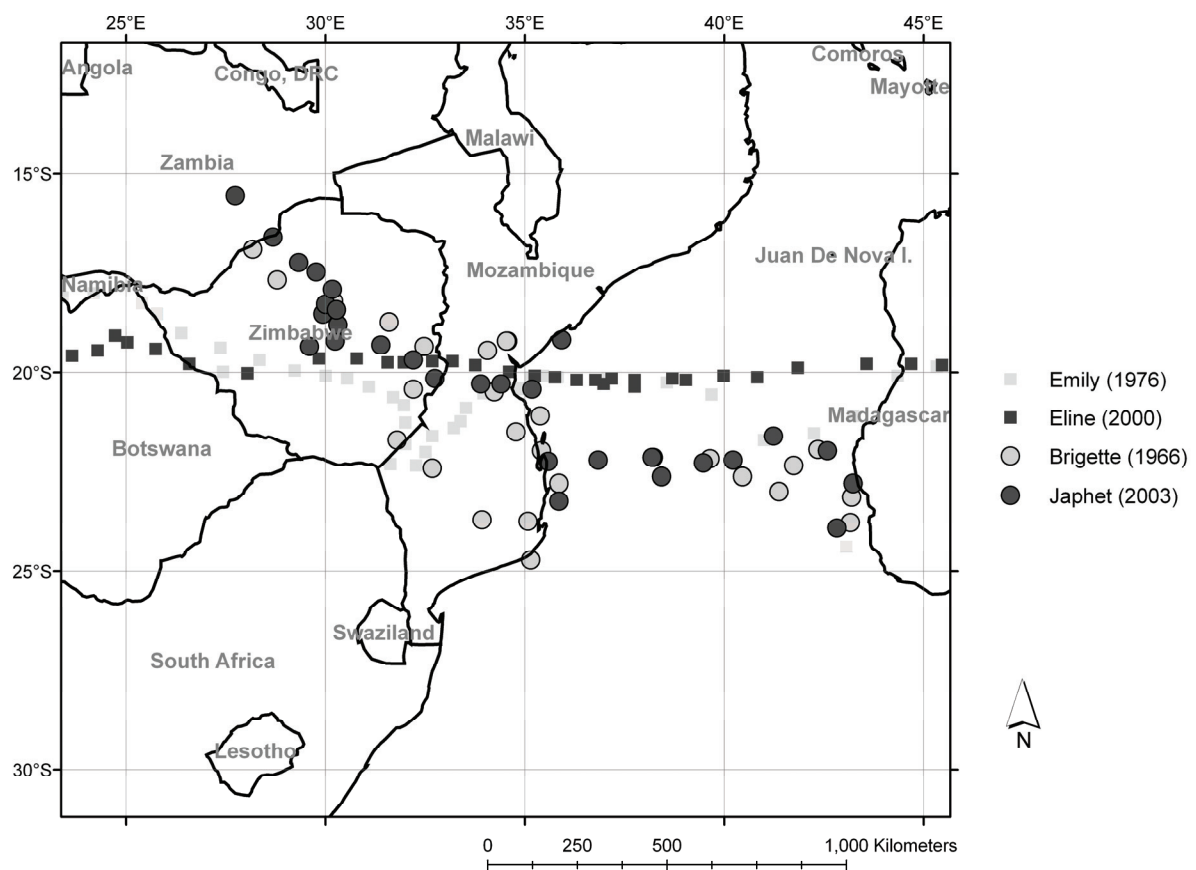


Fig. 6.25: Zonal and less zonal tracks followed by tropical systems from the SWIO

Chapter 7

Case Studies of Steering Mechanisms of Landfalling versus Non-landfalling Tropical Systems

7.1 Introduction

Tropical systems from the SWIO contribute a significant portion of multi-day heavy rainfall events over the eastern interior of southern Africa. The westward movement of these systems tends to occur at latitudes where some of the large southern African river basins are found (e.g. Dyson and van Heerden, 2002) with devastating effects downstream. Between 1995 and 2000, these types of systems were responsible for large amounts of rainfall over much of the Limpopo River Basin region (Crimp and Mason, 1999; Reason and Keibel, 2004; Padgett, 2007). Several more systems have also occurred during the years surrounding this period, causing rainfall over the eastern parts of the Limpopo River Basin (e.g. Klinman and Reason, 2008).

Tropical cyclones such as Eline (2000) and Favio (2007) can have devastating effects causing flooding, damage to infrastructure and loss of lives (Reason and Keibel, 2004; Klinman and Reason, 2008). Tropical cyclonic systems exhibit variable tracks but it has been suggested that if the skill of the inter-annual variability of the tracks that are forecast and frequency of these systems can be improved, then it may aid in predicting the risk of a system causing landfall (Vitart et al., 2003; Reason and Keibel, 2004).

Environmental influences observed on tropical cyclone behaviour such as track and intensity happen on various time scales ranging from the climatological, decadal, down to the diurnal scale (Gray, 1988). Within the background of environmental influences on various time and spatial scales, the environmental steering flow (basic current), defined as the average wind in the middle layers of the atmosphere, as well as an interaction with the earth's vorticity field causing a deviation from the steering flow (Holland, 1984), have been shown to be the main determinants of the movement of tropical cyclones. Vertical average values between various lower and middle tropospheric pressure levels (Holland, 1982; Vitart et al., 2003) and for specific levels (Harr and Russel, 1991; Dong and Neumann, 1986) have been used to define the steering flow. In the southern hemisphere, tropical cyclones moving in a west/southwest direction within the environmental steering flow are generally less sensitive to perturbations in this basic current than tropical cyclones moving south or east while a stronger basic current also causes the tropical cyclone movement to exhibit less fluctuations. The role of potential vorticity in the upper troposphere has been shown to also have an influence on the motion of tropical cyclones with hurricanes moving away from areas of strong positive (negative) potential vorticity in the northern (southern) hemisphere, also explained by the movement of tropical cyclones along the upper air ridges where inertial and static stability is weak, promoting organized convection associated with a tropical cyclone (Becker et al., 2010). Further contributions to the movement of tropical cyclones are made by vorticity advection and diabatic heating (e.g. Chan, 2005), potential vorticity tendency (Chan et al., 2002) and SSTs (Mandal et al., 2007).

Landfall of tropical cyclones from the SWIO over southern Africa has been related locally to the zonal steering current and sea surface temperatures in the SWIO (Vitart et al., 2003; Reason and Keibel, 2004; Klinman and Reason, 2008).

The displacement of several tropical cyclonic systems over the Mozambique Channel is considered in relation to the following fields: 500-700 hPa steering flow, 700 hPa

geopotential heights, 250 hPa temperatures, 250 hPa potential vorticity (tendency) and 250 hPa flow.

7.2 Methodology

Cases of nine tropical systems are evaluated. The elements relevant to steering of the systems considered are:

- a) The environmental steering flow and 700 hPa heights
- b) 250 hPa temperature
- c) Upper air potential vorticity and winds

The steering flow is calculated from the 700 and 500 hPa wind fields, as these heights have been pointed out as being important for steering in the SWIO specifically (Jury and Pathack, 1991) and fall within the envelope generally accepted as the steering flow (Chan, 2005). Upper air potential vorticity is calculated from the following equation:

$$PV = -g (\delta\theta/\delta p) (\zeta + f)$$

where Potential Vorticity (PV) is the product of gravity (g), the vertical potential temperature gradient with respect to pressure and vertical vorticity with:

$(\delta\theta/\delta p)$	Pressure change with height
ζ	Vertical vorticity
f	Coriolis
$\theta = T(1000/P)^{0.287}$	Potential temperature

The other elements (height, wind vector and temperature) are all used as available from NCEP Reanalysis.

The cases are grouped according to track:

Landfall – zonal: The first four cases are representative of systems tracking westward that eventually made landfall and moved zonally into the interior of southern Africa. These types of systems are responsible for widespread rain over the Limpopo River Basin.

Landfall – northwestward tracking: This is for a system that also made landfall, but did not follow a westward track – instead following a northwestward track limiting the contribution to rainfall over the Limpopo River Basin.

Non-landfall – eastward tracking: A system that moved towards the east away from southern Africa, not contributing to rainfall over the Limpopo River Basin.

Non-landfall – southward tracking: The last three systems considered generally moved in a southerly direction over Madagascar or the Mozambique Channel, not contributing to rainfall over the Limpopo River Basin.

A summary of the findings based on individual events is obtained also by considering average values over a set of 26 tropical cyclone events ranging from 1989 to 2008, downscaled from NCEP Reanalysis data by CCAM as per section 2.1. These events are listed in Table 7.1.

Table 7.1: Tropical cyclone events used in the dataset showing the months they occurred, days of data, year and if they made landfall or not

Month	Day	Year	Landfall
January	10-19	1989	No
January	1-10	1990	No
January	20-30	1995	No
January	9-18	1996	Yes
January	1-11	1997	No
January	19-31	1997	No
January	22-31	2003	No
February	1-10	1994	No
February	11-20	1994	No
February	13-21	1995	Yes
February	1-10	1996	Yes
February	24-28	1997	No
February	10-19	1999	No
February	1-10	2000	Yes
February	17-26	2000	Yes
February	1-10	2006	No
February	19-28	2007	Yes
February	14-23	2008	No
March	3-12	1995	No
March	1-10	1997	yes
March	1-10	1999	Yes
March	5-14	2000	Yes
March	4-13	2001	No
March	1-10	2003	Yes
March	4-13	2004	No
March	3-12	2008	No

The data (1028 input synoptic fields per variable) are subdivided into 4 categories in order to enhance collective analysis.

The data for the periods during which landfalling and other systems moved over the Mozambique Channel are identified separately (2 classes).

The data for the entire sequences of landfalling and non-landfalling systems are also subdivided (another 2 classes).

Differences are calculated according to the 4 classes for steering flow, upper air geopotential heights and upper air wind.

7.3 Results and Discussion

7.3.1 Landfalling systems – westward tracking

The zonal nature of the tracks of these systems westward over the subcontinent is responsible for them causing widespread and heavy rain over the Limpopo River Basin.

Steering flow: A strong steering flow from the east is present at the 700-500 hPa level in a continuous belt stretching over the entire area of interest mostly between 15 and 30°S for all these systems (e.g. Fig. 7.1).

700 hPa heights: Throughout the event (prior to, and after landfall) a ridge at 700 hPa is located over the far southern parts of the subcontinent and extends towards an anticyclone located to the southeast of southern Africa over the Indian Ocean (e.g. Fig. 7.1). With each event, the ridge over the southern parts of southern Africa is located towards the northeast of a trough located towards the southwestern parts of South Africa.

250 hPa temperatures: Even before landfall, and during the event, a zonally elongated area of higher temperatures (a warm ridge) at this level extends from the SWIO over the subcontinent (e.g. Fig. 7.2). Movement of the system occurs in the ridge towards the west. Temperature fields are orientated zonally rather than meridional.

250 hPa wind vectors and potential vorticity: In all cases, a northwesterly flow is evident over the southwestern parts of the subcontinent in association with the trough to the southwest, while an anticyclonic curvature can be seen over the far eastern parts of southern Africa and more strongly towards the east or southeast of the subcontinent (e.g. Fig. 7.3). The flow around this anticyclone is also conducive to steering of the system towards the west.

Potential vorticity (PV) over the subcontinent remains small with higher values towards the south. The PV becomes less negative in the direction into which the system moves. Negative PV values (associated with building troughs) become stronger sometimes towards the southeast of the subcontinent as the system moves further to the west.

7.3.2 Landfalling system – northwestward tracking

This system made landfall but was tracking northwestward from the southern Mozambique Channel towards the northern parts of Zimbabwe, limiting the extent of rainfall over the Limpopo River Basin.

Steering flow: Steering flow was weak compared to the systems mentioned above. At the relevant level, the flow around the cyclone is cyclonic with a stronger flow from the southeast on the western side of the system where the 700 hPa anticyclone is ridging in to the southwest of the system from the subcontinent. This flow from the southeast seems to be a main contributor to the movement of the system at this level as it travels northwestward around the edges of the high towards the west over the subcontinent.

700 hPa heights: The system moves around the eastern edge of a ridge located over the subcontinent. Unlike the above-mentioned cases of zonal tracking landfalling systems, a ridge to the south of the system is absent, hence the lack of a clear westward steering flow.

250 hPa temperatures: The movement of this system also happened within the warm ridge at 250 hPa, stretching west-northwestwards into the eastern parts of the subcontinent.

250 hPa wind vectors and potential vorticity: Unlike the systems tracking zonally into the subcontinent, there is no indication of the existence/strengthening over and to the southeast of the system. However, with the trough building towards the southeast of the system, the flow

at this level is becoming more southerly. This could also be related to the northwestward displacement of the system.

The PV becomes strongly negative where an upper air trough builds to the east of the 700 hPa ridge located over the subcontinent. This drastic decrease in PV can also be related to the northwestward movement of the system, in the absence of a strong steering flow at the 700-500 hPa levels mentioned earlier, due to vorticity interaction.

7.3.3 Eastward-moving system

This system moved from Madagascar in an easterly direction becoming part of a west-wind trough.

Steering flow: The flow towards the southwest of the system initially seems to be strong easterly, but weakens drastically through the event and by the end is rather strong westerly. The weak ridge towards the south of the system disappears completely and becomes displaced by the trough in the westerlies. This westerly flow suggests an eastward displacement rather than a movement to the west.

700 hPa heights: Again, the 700 hPa high pressure system is stronger over the subcontinent towards the west of the system, with a very weak ridge towards the south of the system. This weak ridge and a high pressure system towards the southeast weaken further and become displaced by a trough, which is responsible for the collapse of the easterly flow which existed earlier.

250 hPa temperatures: Higher temperatures at this level overlay the system with no indication of a zonally elongated warmer region stretching towards the west from the system as with the landfalling westward tracking systems.

250 hPa wind vectors and potential vorticity: At this level the wind field gives a good indication of a reason for the system to move eastwards. Towards the west of the system, an anticyclonic flow is evident with the flow being generally westerly across the system, together with the high pressure system collapsing to the southeast and the 700-500 hPa steering flow also becoming westerly and being responsible for the displacement towards the east.

Large negative PV values towards the west of the system are also an indication of the developing trough in the westerlies, within which the system travels eastward. The magnitude of negative potential vorticity values in the direction of movement of the system diminishes.

7.3.4 Southward-moving system

This system moved towards the south over the SWIO, not making landfall over southern Africa.

Steering flow: Easterly flow becomes disrupted and is usually more strongly orientated from the north (e.g. Fig. 7.4). Where the easterly flow remained throughout the landfalling events, there is a clear displacement in these events by a flow from the west/north.

700 hPa heights: The 700 hPa high is strongly developed towards the east of the system facilitating a northerly flow while there is a weak area in the ridge towards the south of this system. The ridge towards the south/southwest of the system weakens during the event while the high pressure system towards the east stays in place, strengthening the flow from the north while a westerly trough approaches towards the southwest.

250 hPa temperatures: Temperatures towards the west of the system are lower (unlike with westward moving systems). This is also indicative of a (developing) trough over southern Africa or the SWIO (e.g. Fig. 7.5).

250 hPa wind vectors and potential vorticity: Potential vorticity towards the west of the systems is usually strongly negative indicating the development of a trough directly to the west of the system, over southern Africa or even further east (e.g. Fig. 7.6). The wind at this level is usually westerly to northerly over the system, with a cyclonic curvature, unlike the anticyclonic curvature over or towards the east/south of the system with landfalling systems. This also shows the contribution at this level of the air flow to the movement of the system and why the wind vectors up to this level are sometimes considered to be the environmental steering flow.

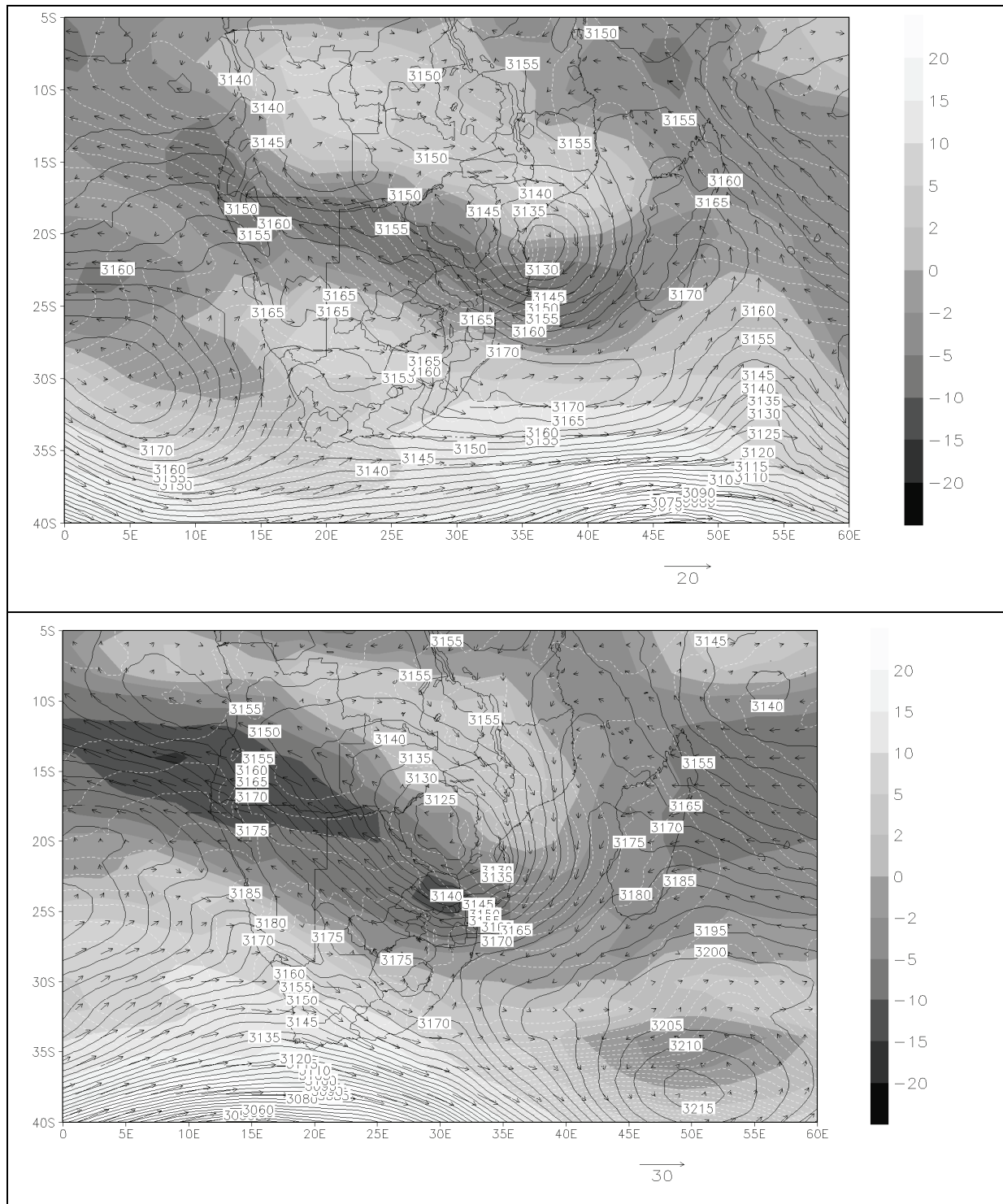


Fig. 7.1: Steering flow (700-500 hPa) vectors (arrows) and zonal component (shading) for a landfalling tropical depression in February 2000 on 4 February (top) and after landfall on 7 February (bottom). Black contours represent the synoptic situation at 700 hPa on the 4th and white contours represent the situation on the 7th

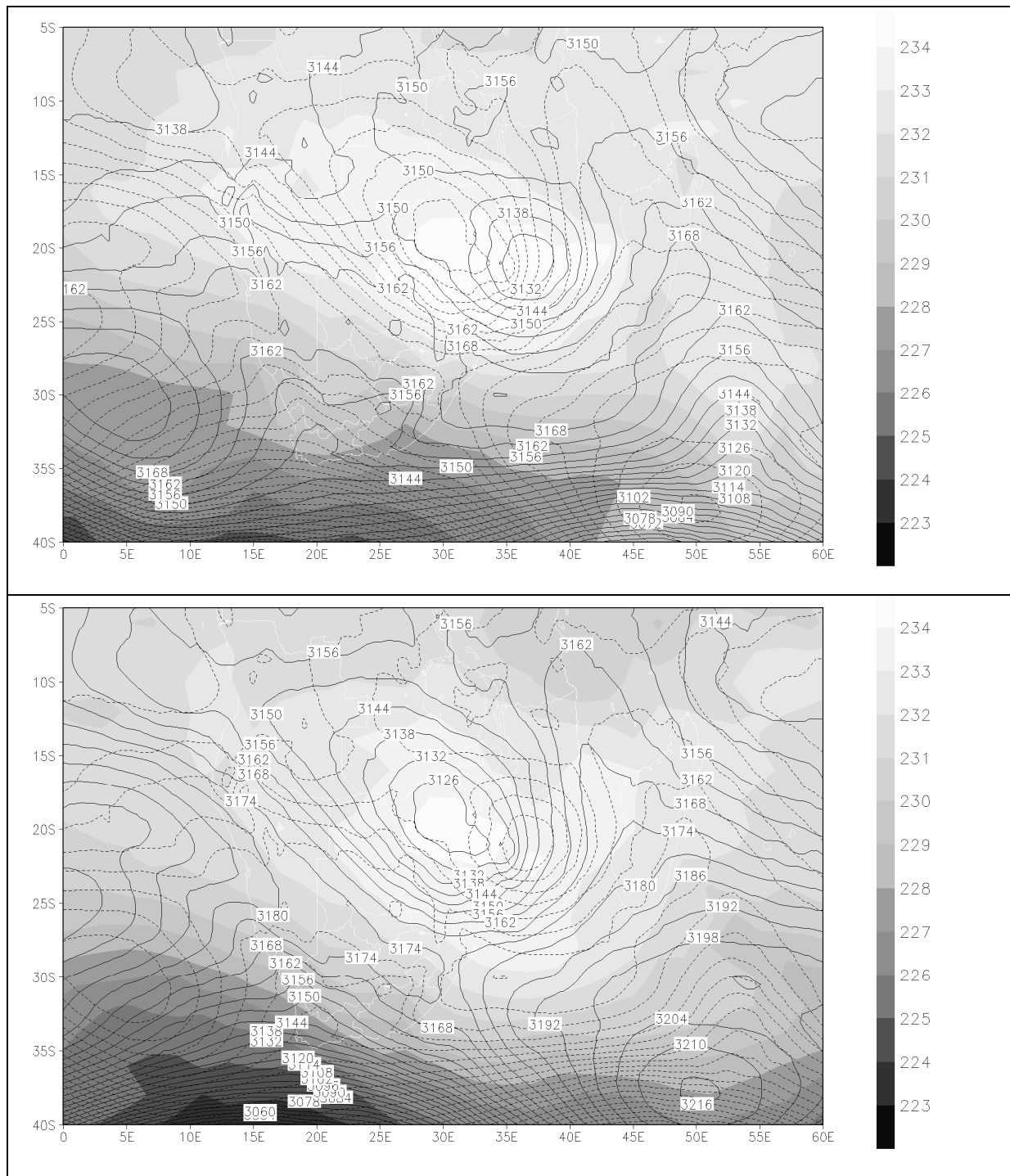


Fig. 7.2: Temperatures at the 250 hPa level during the landfall of a tropical depression in February 2000 on 4 February (top) and after landfall on 7 February (bottom). Solid contours represent the synoptic situation at 700 hPa on the 4th and dashed contours represent the situation on the 7th

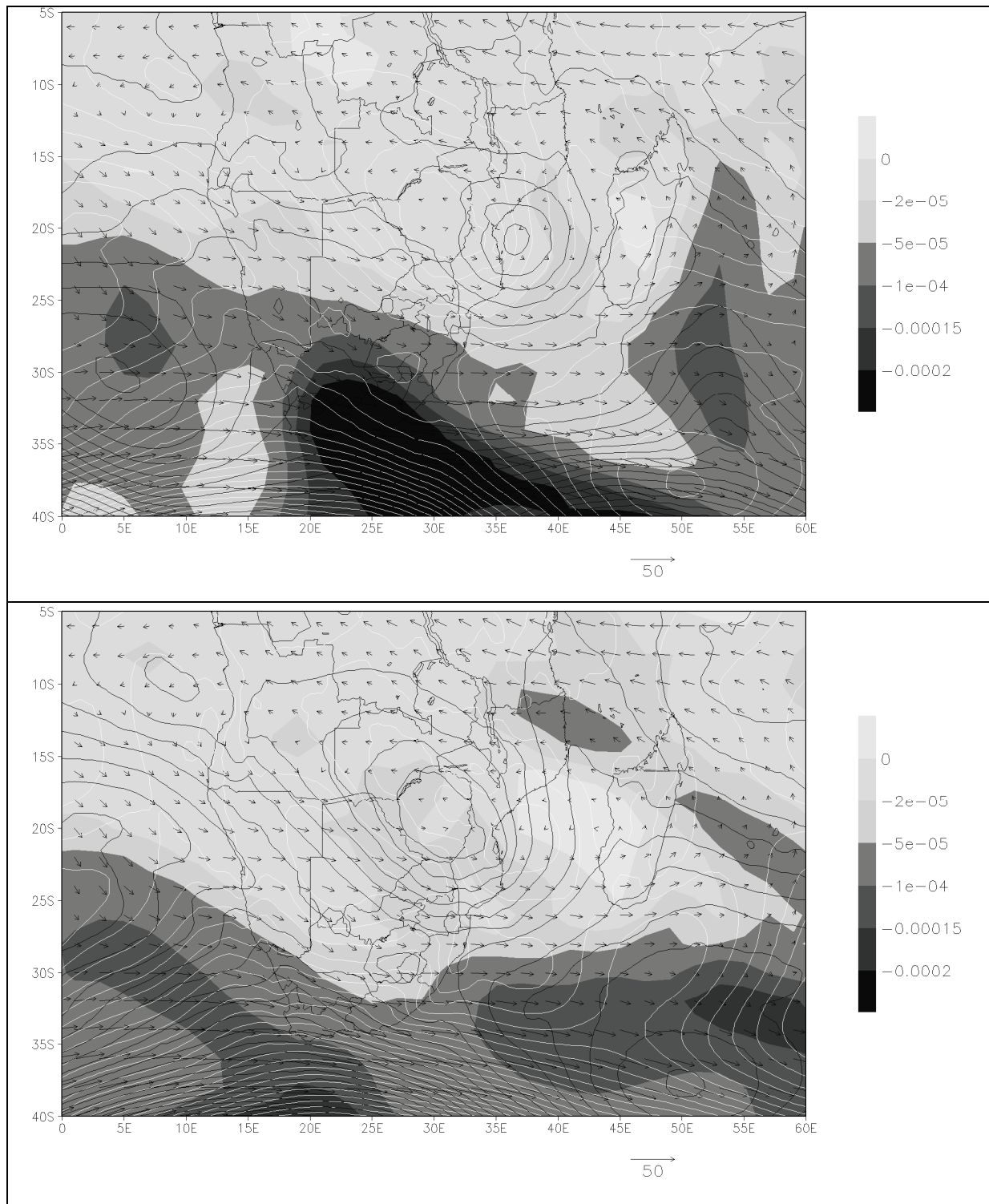


Fig. 7.3: Wind vectors and potential vorticity at 250 hPa during the landfall of a tropical depression in February 2000 on 4 February (top) and after landfall on 7 February (bottom). For each image, black contours represent the synoptic situation at 700 hPa at that specific time and the white contours represent the situation during the other relevant time

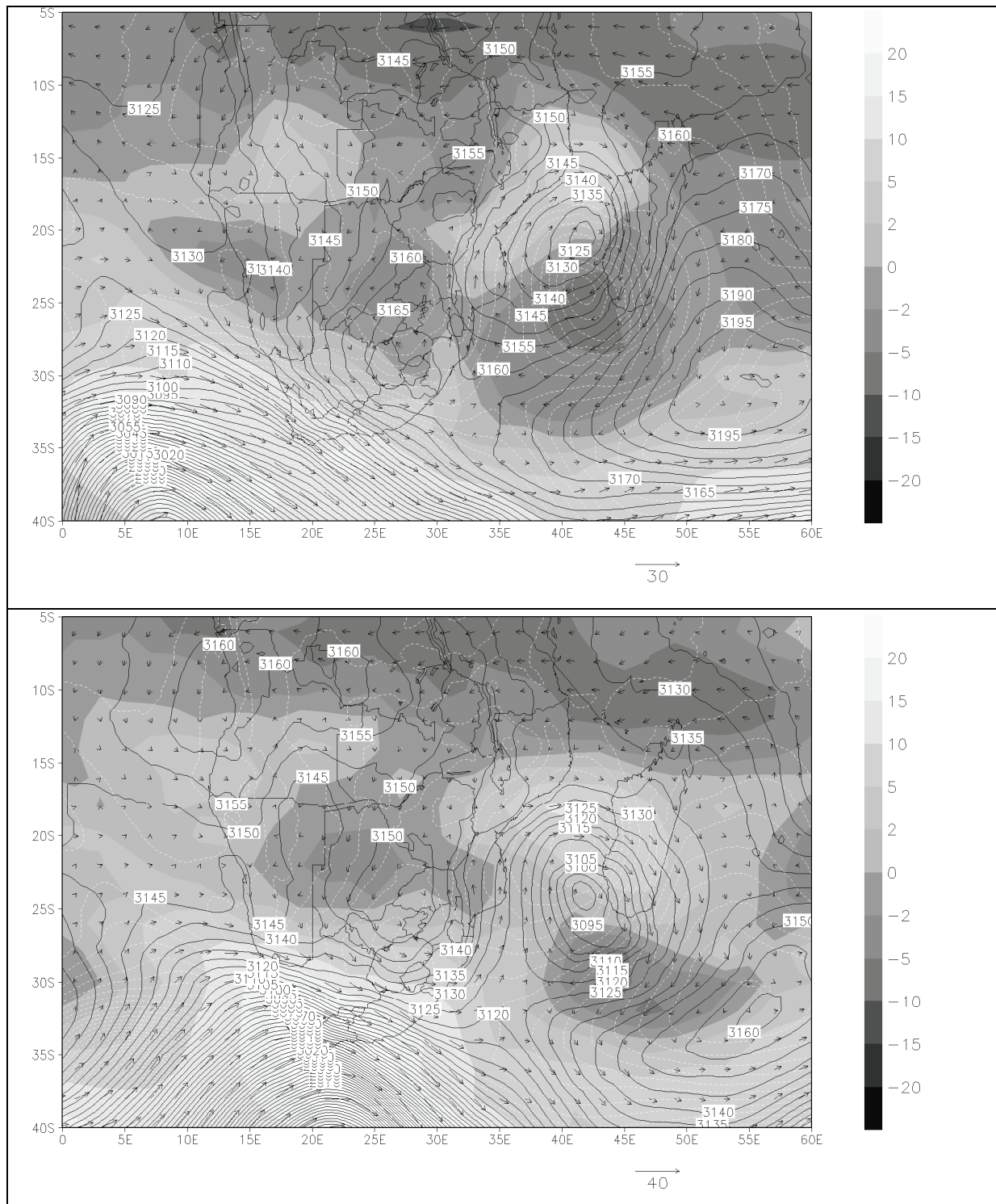


Fig. 7.4: Steering flow (700-500 hPa) vectors (arrows) and zonal component (shading) for a southward moving tropical cyclone (Fabriola) in January 1997 on 7 January (top) and on 9 January (bottom). For each image, black contours represent the synoptic situation at 700 hPa at that specific time and the white contours represent the situation during the other relevant time

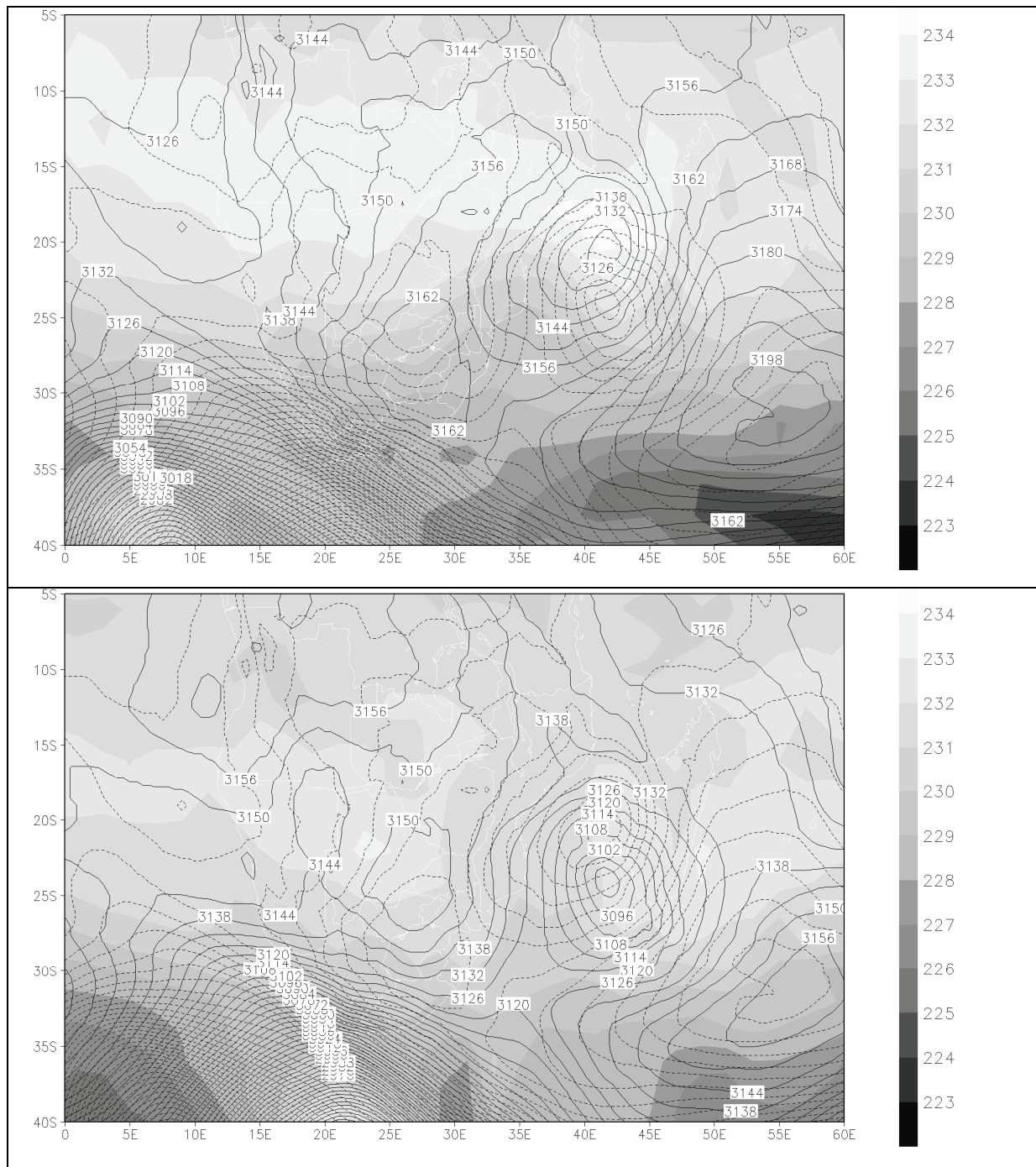


Fig. 7.5: Temperatures at the 250 hPa level during the southward movement of a tropical cyclone (Fabiola) in January 1997 on 7 January (top) and on 9 January (bottom). For each image, solid contours represent the synoptic situation at 700 hPa at that specific time and dashed contours represent the situation during the other relevant time

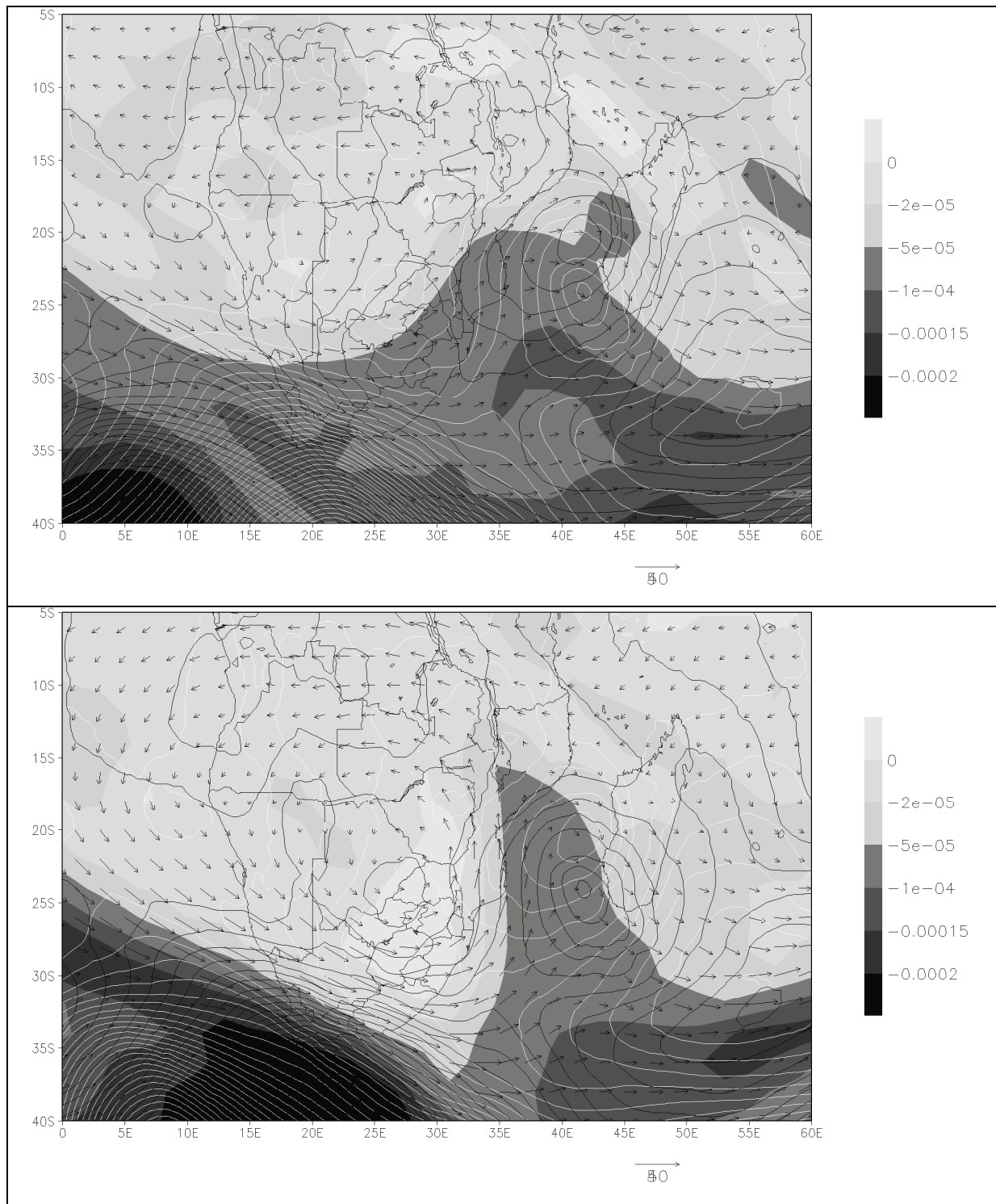


Fig. 7.6: Wind vectors and potential vorticity at 250 hPa during the southward movement of a tropical cyclone (Fabiola) in January 1997 on 7 January (top) and on 9 January (bottom). For each image, black contours represent the synoptic situation at 700 hPa at that specific time and the white contours represent the situation during the other relevant time

7.4 Discussion

Four landfalling systems which tracked zonally into the southern African interior are considered. These landfalling systems are steered towards and into the subcontinent in the 500-700 hPa easterly flow found to the north of a 700 hPa ridge located from the southern subcontinent towards the east with westward movement also supported by an upper air anticyclonic gyre located towards the east/southeast of the continent, over or to the south of the system. The westward movement further happens within the warm ridge located towards the northeast of an upper air trough over the southwestern parts of southern Africa. Another system tracking northwestward into the subcontinent reveals possible upper air PV interaction, responsible – together with an approaching anticyclone to the west – for steering the system away towards the northwest from a developing trough to the southeast over the southern SWIO. On the contrary, cyclones moving away from the subcontinent (the four considered here) experience a collapse in the easterly steering current with a weakening of the 700 hPa ridge to the south and a trough in the upper air developing towards the west of the system, cutting off further westward movement towards the subcontinent by also contributing to a westerly steering current over the system. Table 7.2 summarizes the two opposing sequences.

Table 7.2: Characteristic steering mechanism differences between landfalling and non-landfalling tropical systems in the SWIO-southern Africa region

Synoptic Feature	Landfall	Non-landfall
700 hPa heights and 700-500 hPa steering	A trough is situated over the southwestern parts of southern Africa with a ridge in the lower levels towards the north and east of the trough in the southwest	A trough in the westerlies is located over southern Africa, the ridge towards the east is much weaker to the south of the system while a high pressure system is located towards the east/southeast of the system over the SWIO, far east (50-60°E)
	The ridge in the lower levels is continuous from the subcontinent towards the SWIO to the southeast of the subcontinent	The ridge towards the south of the system is weak with a stronger high pressure system towards the east
250 hPa heights and steering	An anticyclonic gyre is situated over the eastern parts of southern Africa or towards the east over the SWIO	The wind at this level is westerly over large areas towards the east of the subcontinent, indicating a (developing) trough to the east of the subcontinent
	The system is located within or to the north of the anticyclonic gyre, favouring westward movement	The system is located in the cyclonic curvature of upper air winds, towards the east of the developing upper trough located to the east of the subcontinent
Temperatures at 250 hPa	Temperatures at 250 hPa over the system and into the subcontinent are higher than the surrounding temperature. The system is located towards the eastern edge of this region before landfall	Temperatures towards the subcontinent are lower in the developing trough than over the system and surrounding areas
Potential vorticity at 250 hPa	Strongly negative PV values are restricted towards the south or southeast, not hindering the westward movement into the subcontinent	Strongly negative PV values are located towards the west or southwest of the system, indicative of a developing upper air trough hindering the westward movement of the system

Synoptic Feature	Landfall	Non-landfall
Tracking of system	The system moves into the subcontinent being steered westward by the easterly steering flow at the 500 to 700 hPa levels and the upper anticyclonic circulation	The system moves east/south with collapse of ridge and easterly flow at 500-700 hPa and upper air cyclonic movement resulting from upper westerly trough developing

Considering the four classes mentioned earlier, based on 1028 time steps spanning 26 tropical system events, the findings in Table 7.2 are summarized as follows:

The 700 hPa ridge: Fig. 7.7 emphasizes the importance of a ridge over the southern parts of the subcontinent and into the southern parts of the SWIO, towards the south of a system over the SWIO, in steering it towards and over the subcontinent, as opposed to the non-landfalling events.

Stronger easterly steering flow with landfall: Fig. 7.8 shows clearly in considering averages that a much stronger easterly flow is present at the 700 hPa to 500 hPa levels before and throughout the landfall events causing widespread rain over the Limpopo River Basin as opposed to other events. The westerly deviations towards the southeast are present towards the southeast of the ridging high pressure area, located south of and responsible for the easterly steering current. Fig. 7.9 shows that temperatures over large parts of the subcontinent are lower prior to and during landfall than during non-landfall events. This can be attributed to the stronger steering flow present (also before landfall and westward movement) during landfalling events responsible for cooler (wetter) conditions over the subcontinent.

The importance of an upper air anticyclonic gyre towards the east of southern Africa: Figs. 7.10 and 7.11 shows the average differences of the 250 hPa heights and wind vectors respectively, indicating the tendency for an upper air high pressure system anomaly over and towards the east of the subcontinent during landfall, in contrast to the upper air westerly trough developing over or to the east of the subcontinent when tropical systems are steered away from it. These differences therefore also show the negative effect of a trough developing between the subcontinent and the system to the east on further westward movement and landfall.

A warm ridge extending into the subcontinent in the upper air: The relatively high upper air temperatures, representing the warm ridge, extend into the continent before landfall and intensify during the landfall events, as opposed to lower temperatures over the same areas during non-landfall events, as seen in the 500-250 hPa thickness (Fig. 7.12).

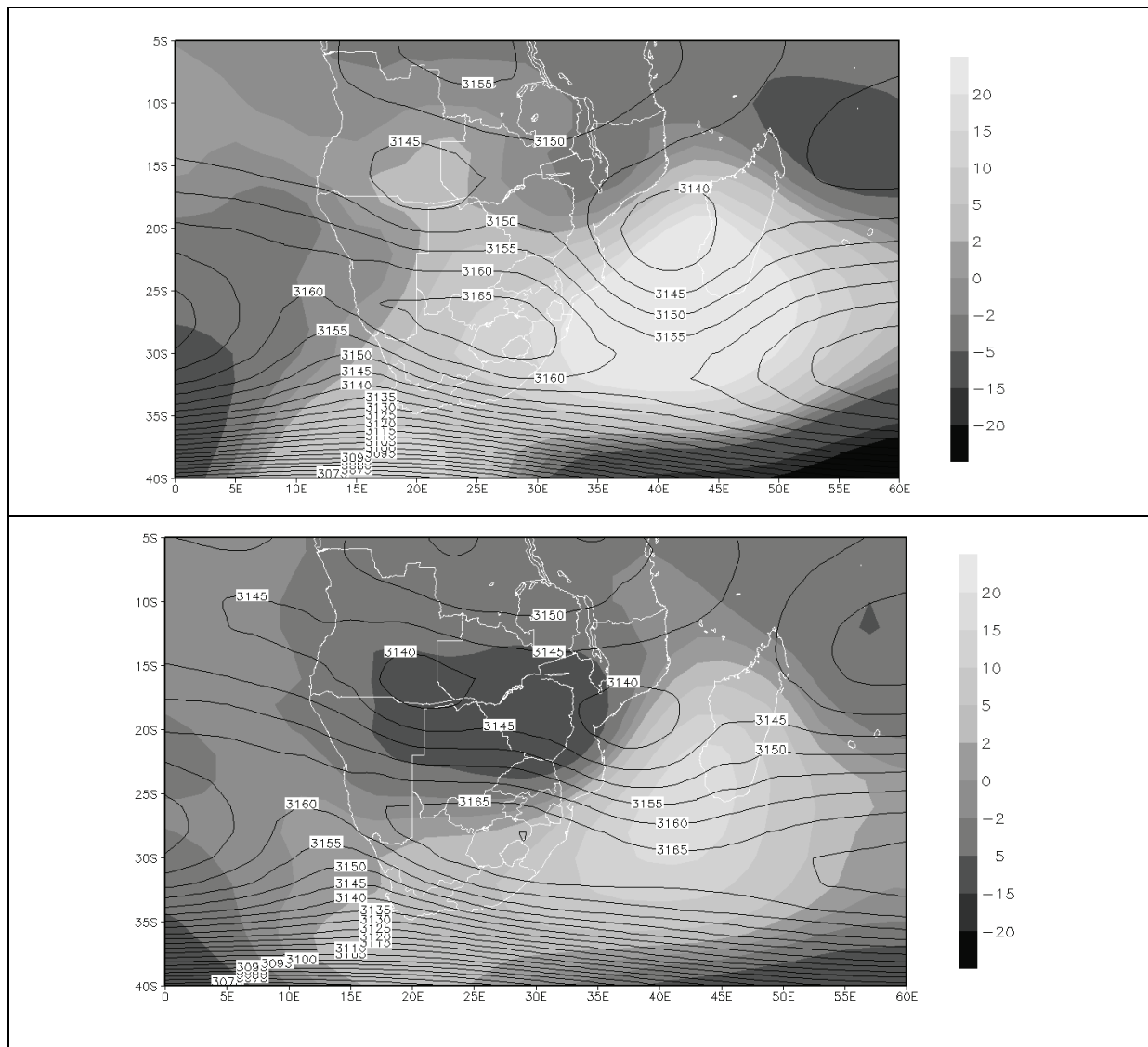


Fig. 7.7: 700 hPa height average (contour) and differences (landfall - non-landfall) for all cases while the tropical system is still in the Mozambique Channel (top) and for the entire sequence (bottom). Average values are shown in contours while differences are indicated in shading

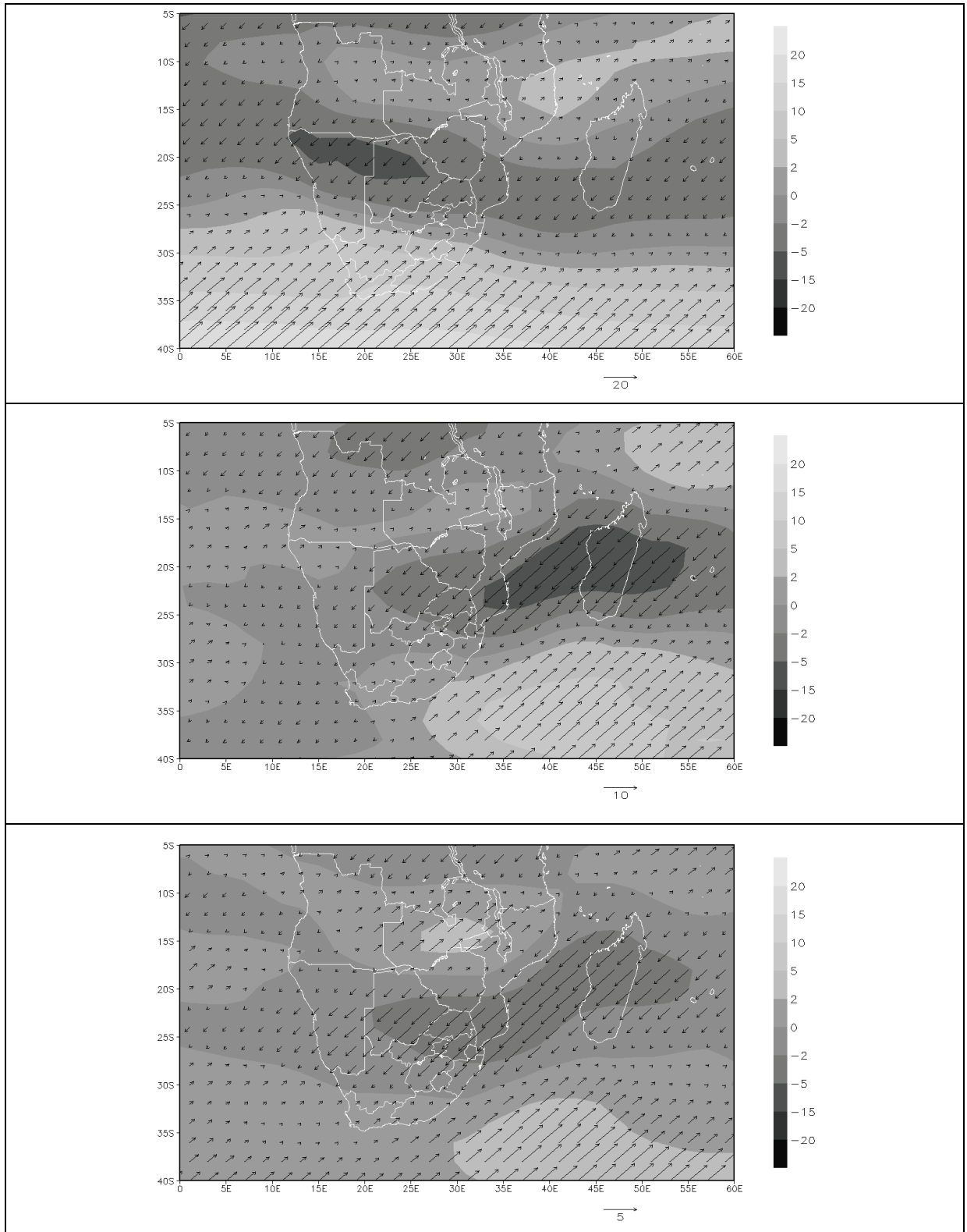


Fig. 7.8: 700-500 hPa average wind vectors (top) for all cases (landfall and other), difference (landfall - other) in wind vectors for all cases while still in the Mozambique Channel (middle) and differences (landfall - other) in wind vectors for the entire sequences (bottom). Wind vectors are represented by arrows while the difference in the zonal wind component is represented by shading

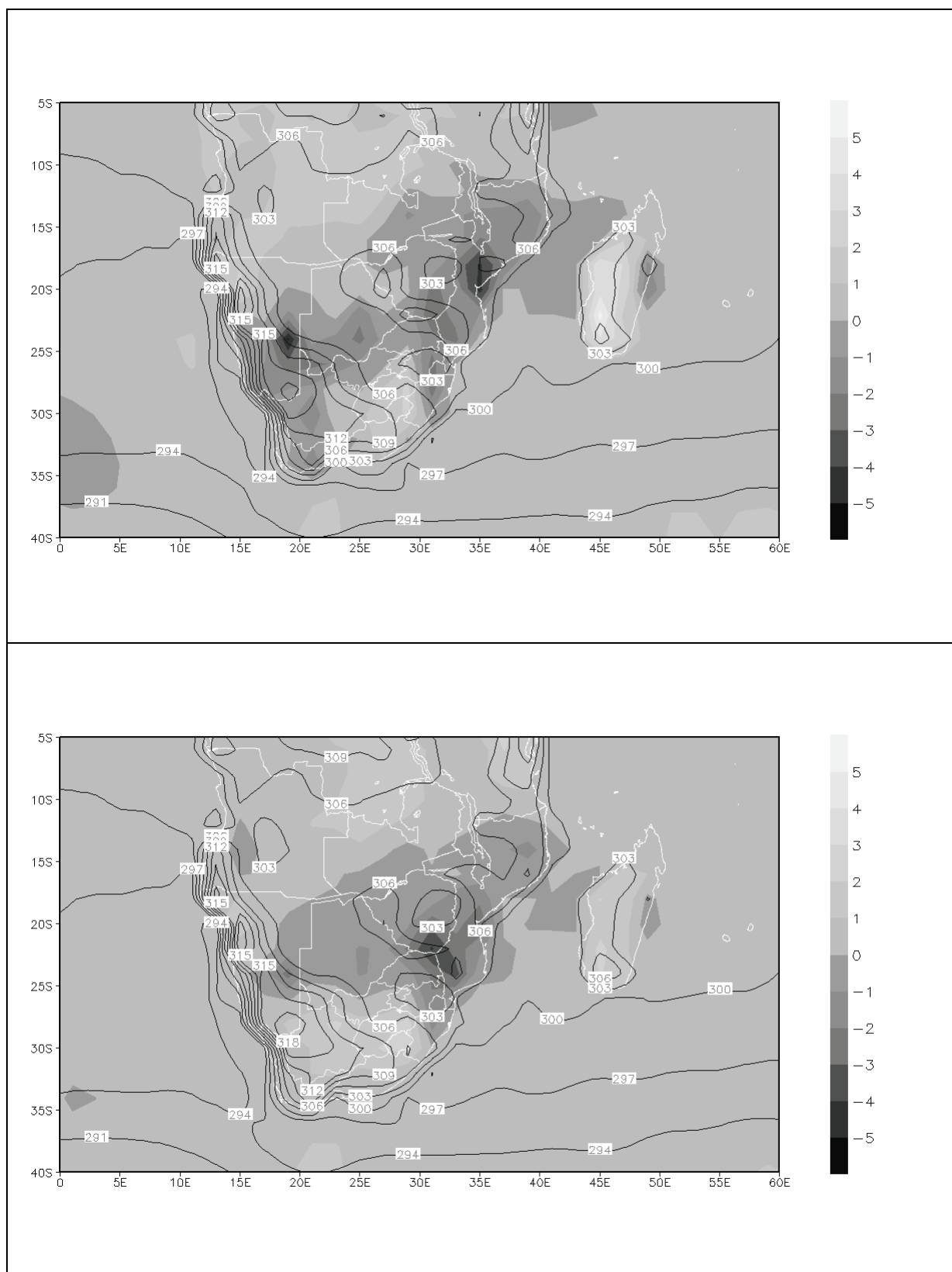


Fig. 7.9: Surface temperature averages and differences (landfall - non-landfall) for all cases while the tropical system is still in the Mozambique Channel (top) and for the entire sequence (bottom). Average values are represented by contours while differences are represented by shading

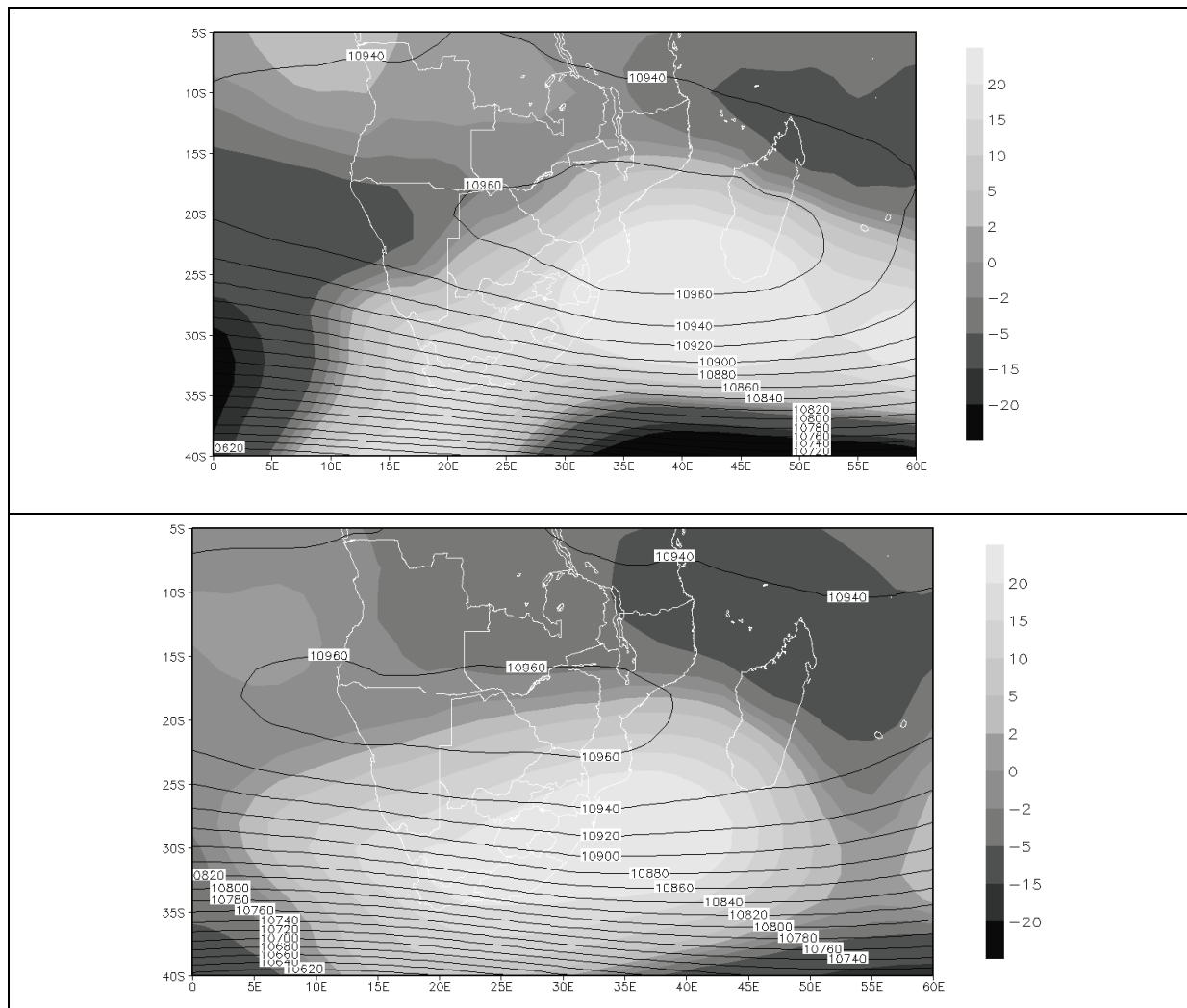


Fig. 7.10: 250 hPa height average and differences (landfall - non-landfall) for all cases while the tropical system is still in the Mozambique Channel (top) and for the entire sequence (bottom). Average values are shown in contours while differences are indicated in shading

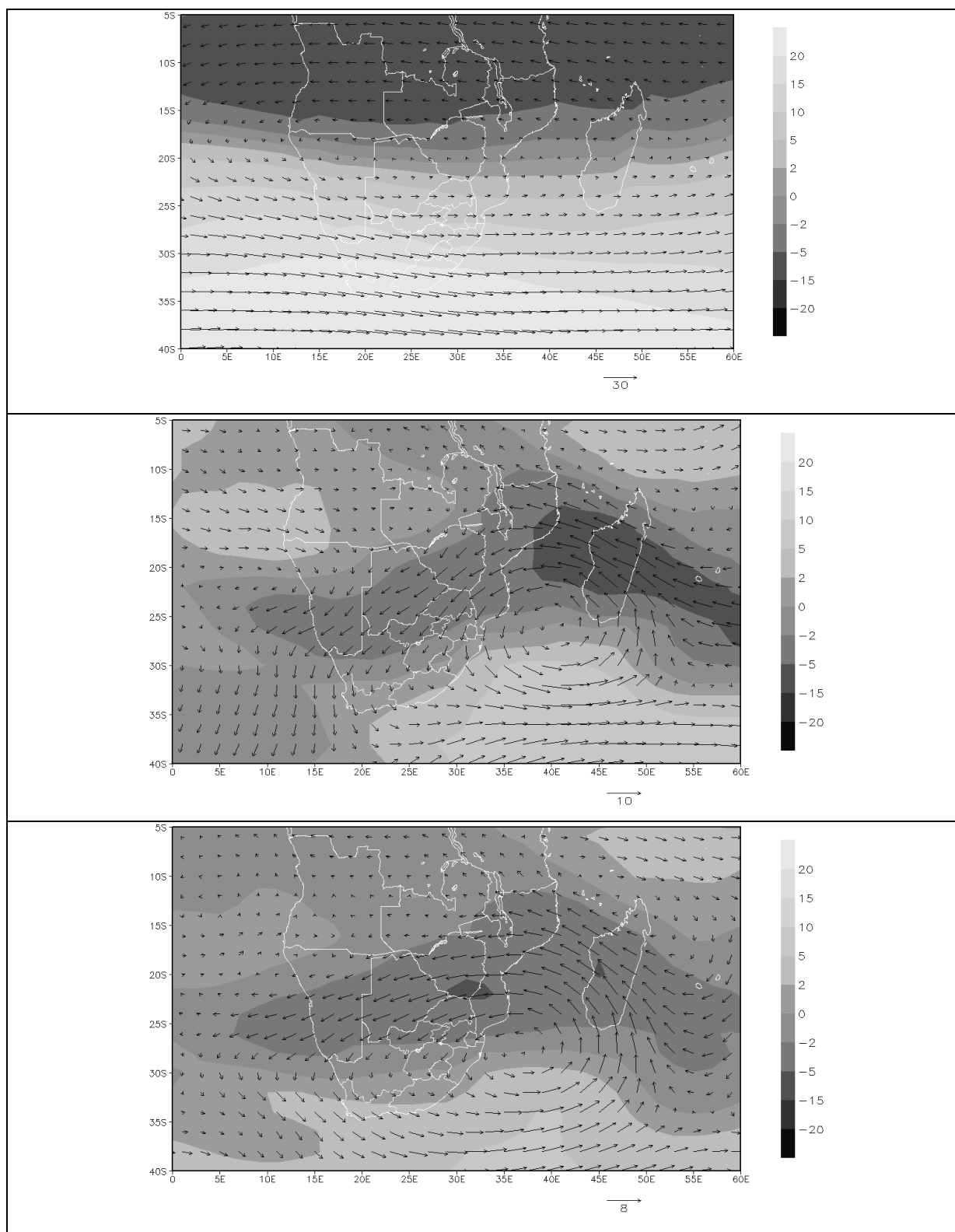


Fig. 7.11: 250 hPa average wind vectors (top) for all cases (landfall and other), difference (landfall - other) in wind vectors for all cases while still in the Mozambique Channel (middle) and differences (landfall - other) in wind vectors for the entire sequences (bottom). Wind vectors are represented by arrows while the difference in the zonal wind component is represented by shading

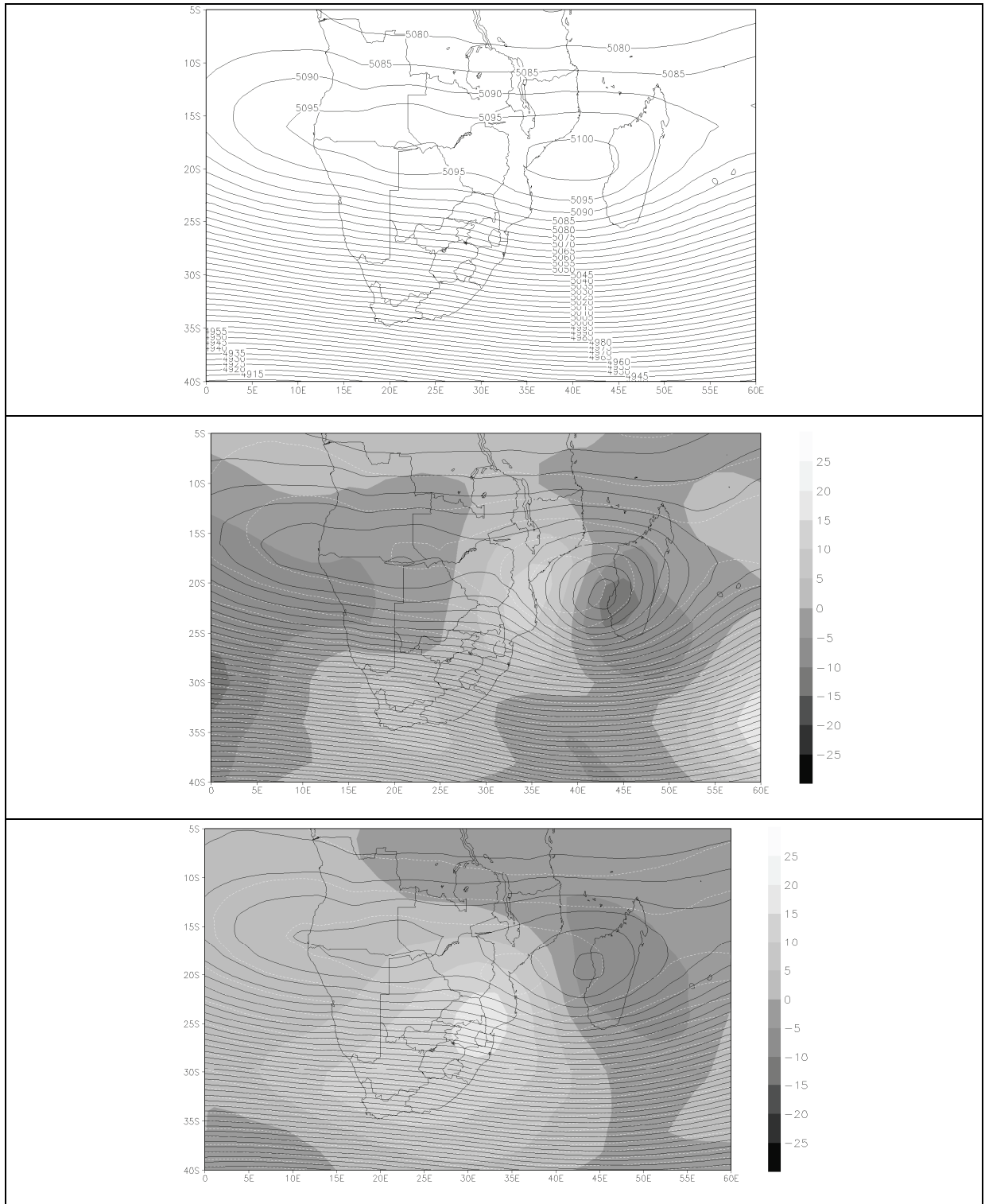


Fig. 7.12: 500-250 hPa average thickness (top) for all cases (landfall and other), difference (landfall - other) in thickness for all cases while still in the Mozambique Channel (middle) and differences (landfall - other) in thickness for the entire sequences (bottom). Difference values are represented by shading while the average values for the landfall and other cases are shown in black and white contours respectively

Chapter 8

Synoptic Climatology of Landfalling Tropical Systems

8.1 Introduction

This chapter further builds on the findings presented in Ch. 7 using a Self-Organizing Map (SOM) approach and also considering variation in the vertical. SOM analysis forms part of the Artificial Neural Network-based downscaling approach and is able to capture the non-linearity of the provided dataset (Takatsuka, 2001; Kohonen, 2001; Hewitson and Crane, 2002). An advantage of SOMs is that they are adaptable and can be a very useful tool for analyzing high-dimensional datasets such as atmospheric data (Takatsuka, 2001; Hewitson and Crane, 2002). A two-dimensional array is created that allows the operator to visualize the relationship between the nodes as well as identifying synoptic changes over time (Hewitson and Crane, 2002). Unsupervised learning algorithms are used to form cluster centres that represent the dataset (Luengo et al., 2004). The number of classes which the SOM represents is chosen by the user through experimenting with the data and background knowledge of the data (Luengo et al., 2004). The higher number of classes will allow for finer detail to be represented, while fewer will result in more generalization (Hewitson and Crane, 2002).

This non-linear classification approach can be used in various fields which include the investigation of extreme climate events in the Balkans in terms of wintertime precipitation (Cavazos, 2000). In this study SOM analysis is used to identify the spatial and temporal atmospheric structure and analyze climate anomalies. SOMs are created in this project to objectively determine synoptic conditions that may occur during landfall of tropical cyclones or to substantiate any type of oscillatory pattern within the synoptic dynamics of the region. SOM algorithms are used to classify the data after which the synoptic conditions favouring landfall of tropical cyclone events can be determined objectively for several variables throughout the troposphere.

8.2 Methodology

The dataset used is the CCAM simulation based on 6-hourly NCEP Reanalysis data as described in Ch. 3 for the events listed in Table 7.1 with a focus on AOI 1 and AOI 2 (Fig. 3.1). The dataset used contains 26 tropical cyclone events ranging from 1989 to 2008 with each event having approximately 10 days of data.

From the larger set of data simulated from NCEP Reanalysis, only the 850 hPa, 700 hPa, 600 hPa, 500 hPa, 250 hPa and 100 hPa levels are considered due to time and data capacity limits. In the case of relative humidity, geopotential heights and temperature from 1000 hPa to 700 hPa as well as surface temperature, only the 12:00 data are used to avoid large changes as a result of surface heating and resultant atmospheric reaction.

The basic processes of creating a SOM include map initialization, training, evaluation of map quantification error and map visualization (Kohonen et al., 1996). Initially, a random collection of nodes are chosen within the high dimensional data space known as reference vectors (Kohonen, 2001; Hewitson and Crane, 2002; Luengo et al., 2004). The input variables presented to the SOM are now compared to the reference vectors and it modifies these vectors to the best matching node updating this reference map (Cavazos, 2000; Hewitson and Crane, 2002; Luengo et al., 2004). To calculate and compare the nodes the smallest Euclidean distance is mostly used (Cavazos, 2000; Hewitson and Crane, 2002; Luengo et al., 2004). This process is often referred to an unsupervised learning process, as it

finds the clusters entered and determines patterns in the data (Kohonen, 2001; Cavazos, 2000). Through an iteration process the reference vectors are adapted one by one until the difference between the dataset no longer exists (Hewitson and Crane, 2002; Luengo et al., 2004). This iteration process accounts for the non-linear characteristics of the data (Hewitson and Crane, 2002). Neighbourhood Kernel functions are also used in this process and can either be Gaussian or ‘bubble’, where the latter uses a neighbourhood set of points around a specific node (Kohonen, 2001). The final output is thus the representation of the cluster of the data and shows the relationship of the data in the specified classes (Luengo et al., 2004). Statistically the accuracy of this method is optimal when the number of iterating steps is at least 500 times greater than the number of classes or nodes (Cavazos, 2000).

The statistical software package SOM_PAK 3.1 is used to create a 5x5 SOM for each of the atmospheric variables at the seven different pressure levels. The data are standardized prior to SOM analysis.

Analysis for this particular section is done for the case where a tropical cyclone occurred within the Mozambique Channel and from where it then either made landfall or did not make landfall. Favourability of synoptic states represented by the SOM nodes for landfall is calculated by considering the percentage of landfall and non-landfall that occurs within each node. For the analysis the highest three landfalling classes and highest three non-landfalling classes are used, although the other classes showing landfall and non-landfall are also considered for further clarity. An index system is also created to determine at which levels the analyzed atmospheric variables exhibit a higher association with the tendency of tropical cyclone movement. This index is calculated using the following formula:

$$\text{Index} = \sum^{1..25} \text{ABS} [(\% \text{landfall}) - (\% \text{non-landfall})]$$

according to which the absolute value of the difference in association with either landfall or non-landfall is summed over the 25 SOM nodes per level in the troposphere.

8.3 Results and Discussion

8.3.1 Geopotential heights

Levels considered: 1000 hPa, 850 hPa, 700 hPa, 600 hPa, 500 hPa, 250 hPa and 100 hPa.

In the lower troposphere the following, according to the SOM analysis, is associated with **landfalling systems**:

- a strengthening of the ridge to the south of the Mozambique Channel and Madagascar – strong positive anomalies are found over that region

During the non-landfalling events:

- a weakening of the ridge occurs with strong negative anomalies found to the south of Madagascar – where the ridge can be replaced by a trough

This strengthening of the ridge during landfall events can also be seen in the middle to upper atmosphere, namely 700 hPa (Fig. 8.1), 600 hPa, 500 hPa, 250 hPa and 100 hPa.

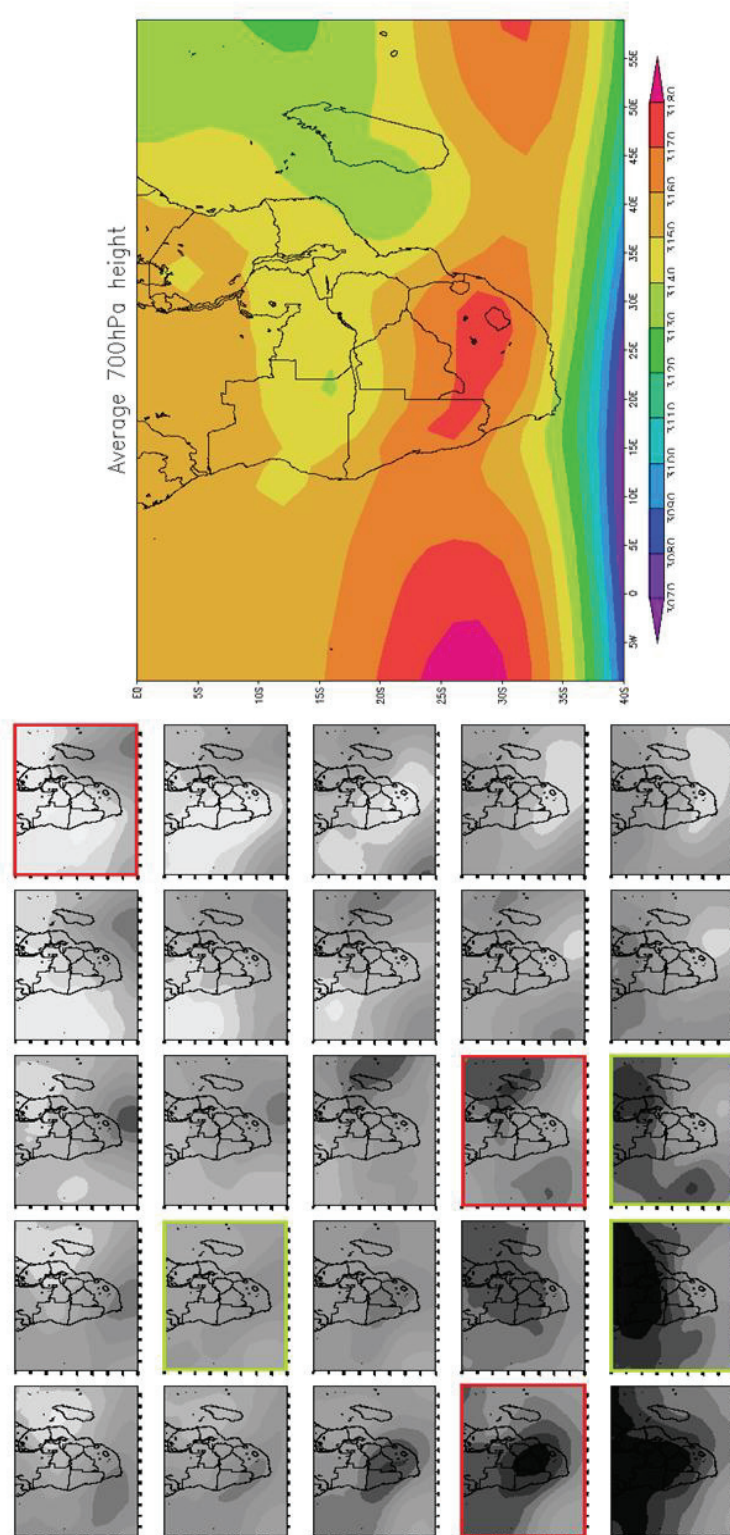


Fig. 8.1: The 700 hPa SOM for geopotential heights with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average value for the 26 tropical cyclone events is shown on the right

In the middle to upper troposphere, **during landfall**:

- strengthening of the ridge towards the south and southeast of southern Africa occurs further to the north (with increasing height)
- positive anomalies occur over the entire domain to the south of Mozambique Channel and Madagascar

- there is a trough to the southwest of the subcontinent which becomes noticeable from 500 hPa and very pronounced at the 250 hPa and the 100 hPa levels

The above-mentioned will be alluded to in Ch. 9.

The strengthening of the ridge to the south of the Mozambique Channel, characterizing landfall of tropical cyclones, allows for the strengthening of the easterly flow over the Mozambique Channel. This was shown by Vitart et al. (2003) to be a requirement of the landfall of tropical cyclones. The presence of this ridge was also argued to be the cause of landfall of tropical cyclone Eline in February 2000 (Reason and Keibel, 2004). Here the ridge to the south of the Mozambique Channel was one of the argued precursor conditions that could have been used to predict the landfall of tropical cyclone Eline, while a preceding trough over southern Africa was another precursor condition. The trough in the upper levels to the southwest of the subcontinent could also play a role in strengthening or maintaining the ridge to the south and southeast of the subcontinent.

8.3.2 Zonal winds

The zonal winds in the lower troposphere, namely 1000 hPa and 850 hPa (not shown), **during landfall:**

- weak values of positive anomalies to the south of Madagascar, thus a slight strengthening of the westerly flow in that region during landfall
- no strong indications are found over the regions where the easterly flow dominates in the mean at the lower levels

From 700 hPa upwards to 250 hPa **during landfall:**

- strengthening of the easterly flow over the Mozambique Channel as also seen in Fig. 8.2 for the 600 hPa level
- strengthening of the westerly flow further to the south

This, along with the easterly flow over the Mozambique Channel, is indicative of the anticyclonic flow pattern of anomalies over the Mozambique Channel region, particularly from 700 hPa to 250 hPa during landfall of tropical cyclones.

These levels (700 hPa to 250 hPa) showing a strengthened easterly flow were also noted by Vitart et al. (2003) to be the most important when considering the environmental steering flow of tropical cyclone movement. An easterly flow was also noted to occur prior to tropical cyclone Eline making landfall (Reason and Keibel, 2004). This enhanced easterly flow is found over a large area, and from about 500 hPa it is found to extend over the subcontinent as well. The enhanced easterly flow is also found to have a displacement to the north with an increasing height.

During non-landfall:

- there exists an anomalously cyclonic pattern in the same region where the anticyclonic pattern dominates towards the southern Mozambique Channel
- throughout the troposphere from 1000 hPa to 100 hPa, a weakening of the easterly flow is found to occur over the Mozambique Channel
- to the south and southeast of Madagascar and the Mozambique Channel negative anomalies are observed resulting from a weakening of the westerly flow over this region

Once again the weakening of the easterly flow over the Mozambique Channel was found in literature to be unfavourable with landfall of tropical cyclones (Vitart et al., 2003).

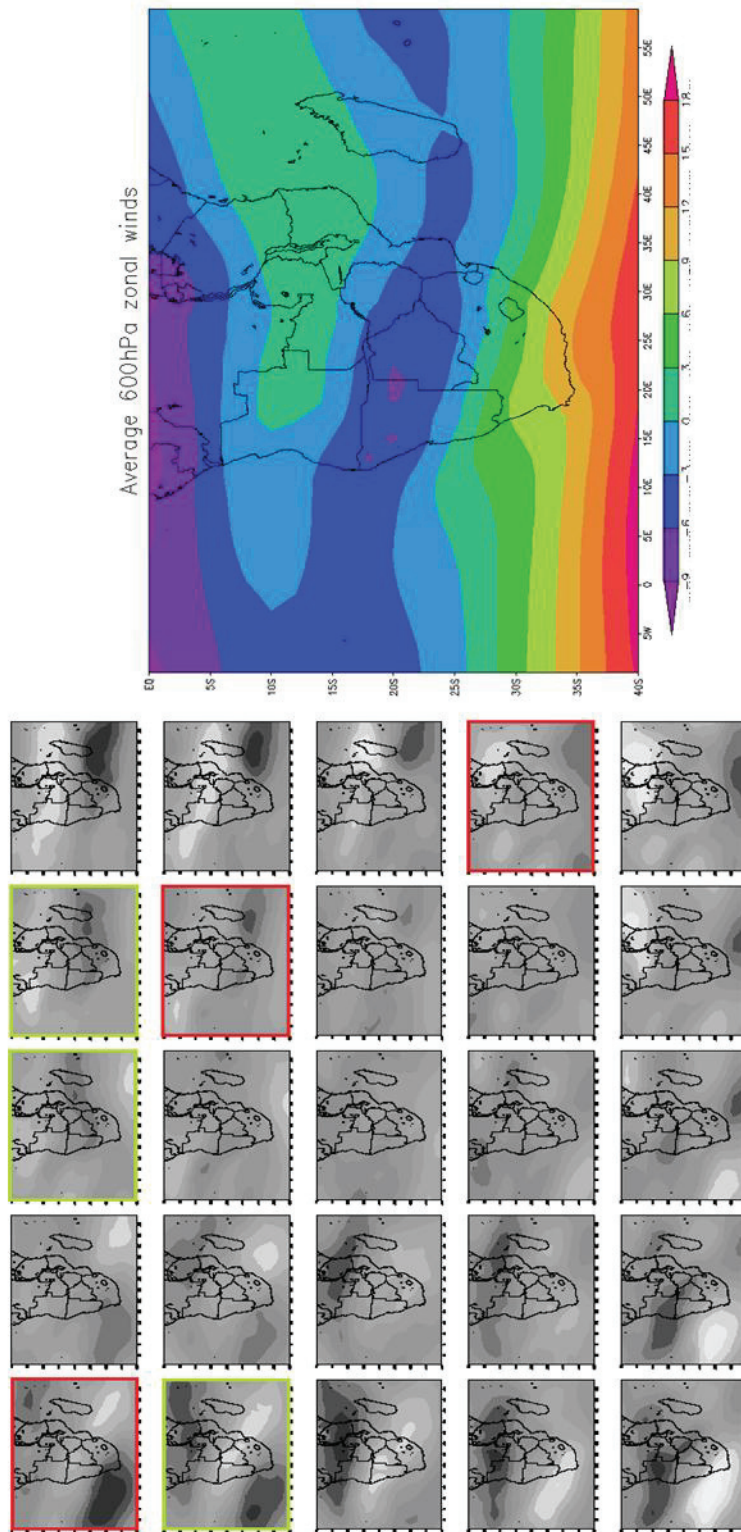


Fig. 8.2: The 700 hPa SOM for 600 hPa zonal wind with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average value for the 26 tropical cyclone events is

8.3.3 Meridional winds

Meridional winds in the lower troposphere (1000 hPa and 850 hPa) show the following **during landfall** as opposed to non-landfall:

- southerly flow deviation over the Mozambique Channel

- a strengthening of northerly flow over southern Africa and down to the south of the subcontinent. This allows for an anticyclonic pattern in anomalies over the eastern parts of the subcontinent as supported by the zonal wind anomaly patterns

This anomalously anticyclonic circulation during landfall versus non-landfall also dominates at the middle (700, 600 (Fig.8.3) and 500 hPa) and upper troposphere (not shown).

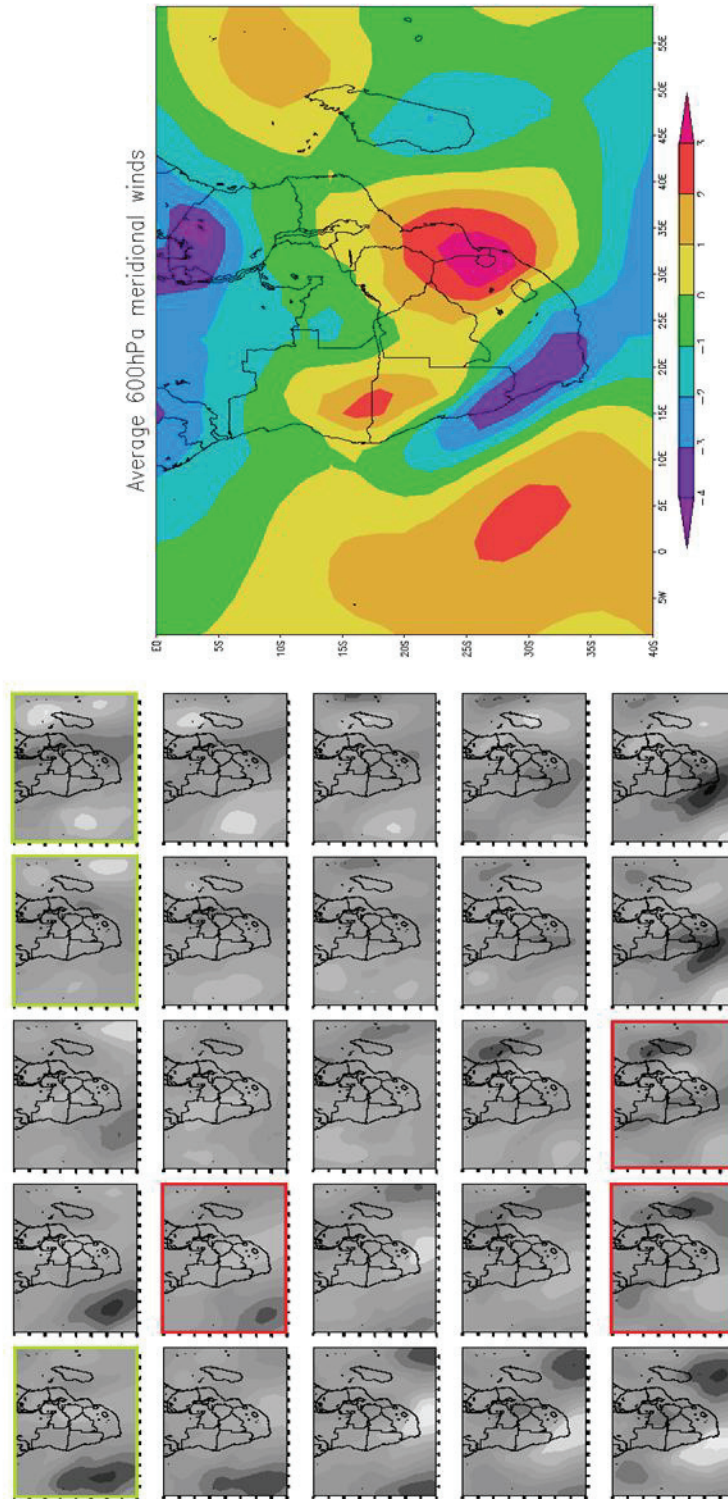


Fig. 8.3: The 600 hPa SOM for meridional winds with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also

8.3.4 Mixing ratio

The mixing ratios in the lower troposphere (not shown) **during the landfalling cases** as opposed to non-landfall, have the following characteristics:

- lower mixing ratio values to the south and southeast of Madagascar
- higher mixing ratio values along the eastern parts of the subcontinent

The opposite is associated with non-landfalling events. This is also found throughout the middle troposphere (700 hPa, 600 hPa and 500 hPa) with the lower mixing ratio values found to the south of Madagascar at the lower levels extending further north with anomalously lower values over the eastern parts of the subcontinent. More noticeable in the middle troposphere is the strong gradient between the higher mixing ratio values towards the east of the subcontinent and lower values over the subcontinent as also seen in Fig. 8.4 for 600 hPa during non-landfall. This allows for an increase in mixing ratio values over this region, where relatively higher mean mixing ratio values on average occur (Fig. 8.4). Over the eastern parts of the subcontinent, however, weak negative anomalies are found, thus causing a strong gradient between much higher mixing ratio values to the east while lower over the eastern parts of the subcontinent.

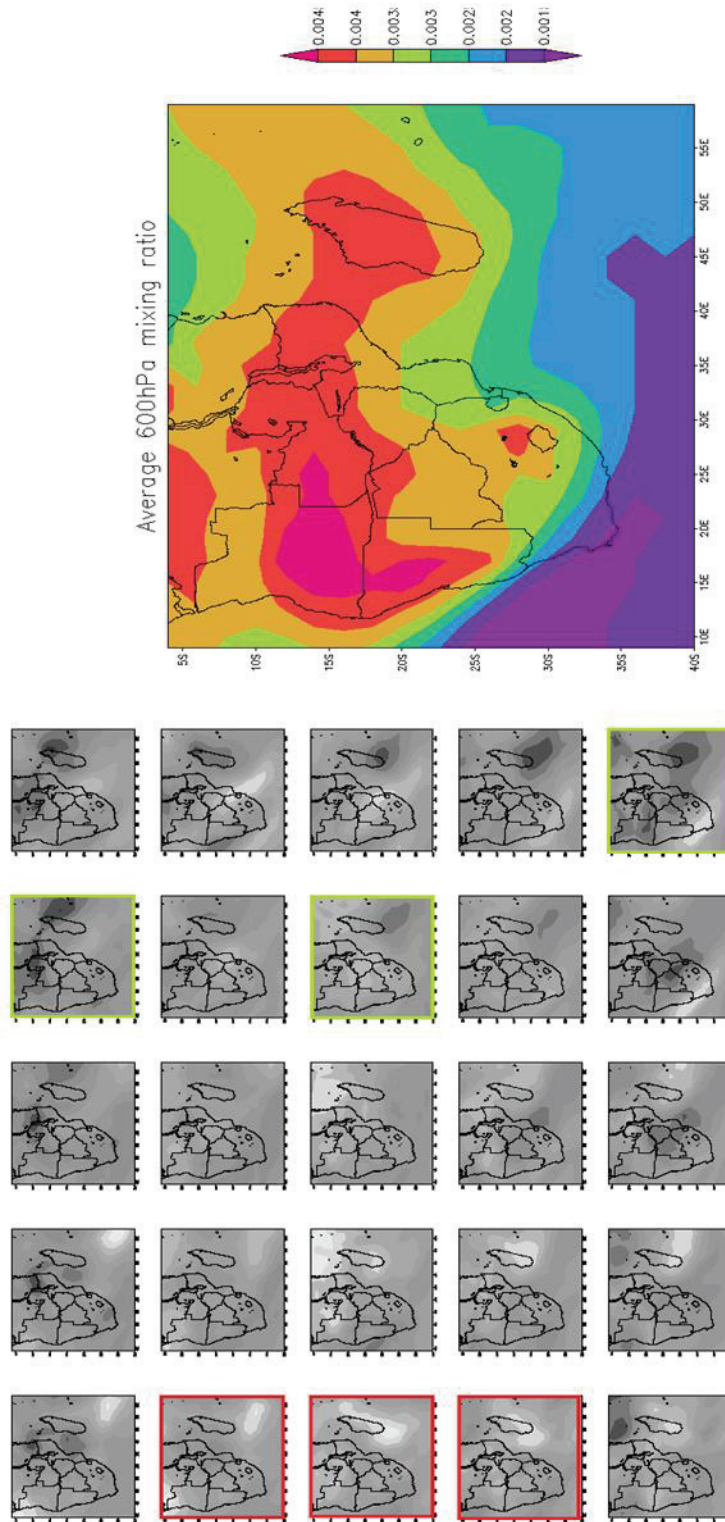


Fig. 8.4: The 600 hPa SOM for mixing ratios with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right)

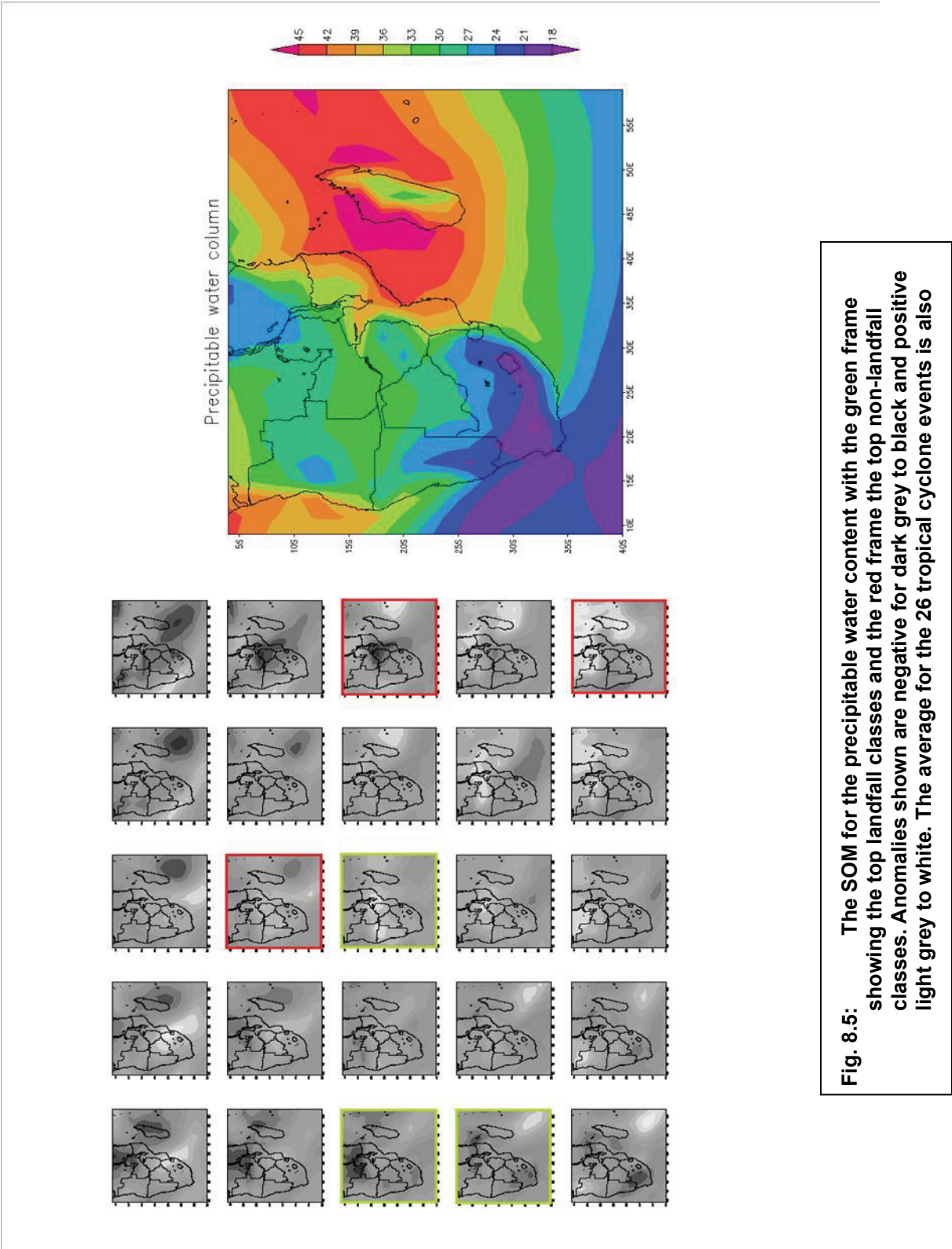
This pattern, favouring landfall and non-landfall in the mixing ratio values, is repeated to at least 250 hPa but becomes unnoticeable at higher altitudes.

In general, the anomalies over the subcontinent and the adjacent SWIO show the mixing ratio to have a more zonal distribution of tropospheric moisture during landfall as opposed to non-landfall. This zonal distribution of moisture, with higher values along the eastern parts of the

subcontinent, could be as a result of the enhanced easterly flow present during landfall of tropical cyclones.

8.3.5 Precipitable water content

The SOM for the column precipitable water content (PWC) can be seen in Fig. 8.5.



The most striking differences noticeable **within the landfalling cases** as opposed to non-landfall are:

- lower PWC towards the north of the southern position of the Inter-Tropical Convergence Zone (ITCZ), over northern Zambia towards the DRC
- higher PWC over the southern parts of the ITCZ, including Zimbabwe and southern Mozambique
- a weaker gradient between higher PWC values towards the SWIO and the lower PWC values located over the subcontinent

These differences between landfalling and non-landfalling events suggest the influence of the strengthened zonal steering flow in enhancing PWC values over the subcontinent and also a southward displacement of the ITCZ over the subcontinent in association with the increased easterly flow advecting moisture from the SWIO over the subcontinent.

The patterns of the vertical average flux variables and their associations with landfall versus non-landfall events are similar to the patterns of the zonal and meridional winds and their associations with landfall versus non-landfall events.

The results for Vertical Average of Zonal and Meridional Temperature Flux are not shown.

8.3.6 Vertical average of zonal and meridional humidity flux

The vertical average of zonal humidity flux results are shown in the SOM in Fig. 8.6. This variable reveals similar results to those of the zonal winds and the vertical average of zonal temperature flux. During the landfall of tropical cyclones, the vertical average of zonal humidity flux has negative anomalies, mainly to the south of the Mozambique Channel, extending over the subcontinent allowing for the enhancement of the mean easterly flux of humidity over that region as seen in the average map in Fig. 8.6. To the north of the Mozambique Channel and the northern parts of the southern African subcontinent, positive anomalies are observed, representing an enhanced westerly flux of humidity present over that region. The enhanced vertical average of easterly flux of humidity over the Mozambique Channel can also be seen as a result of enhanced easterly flow by the zonal winds.

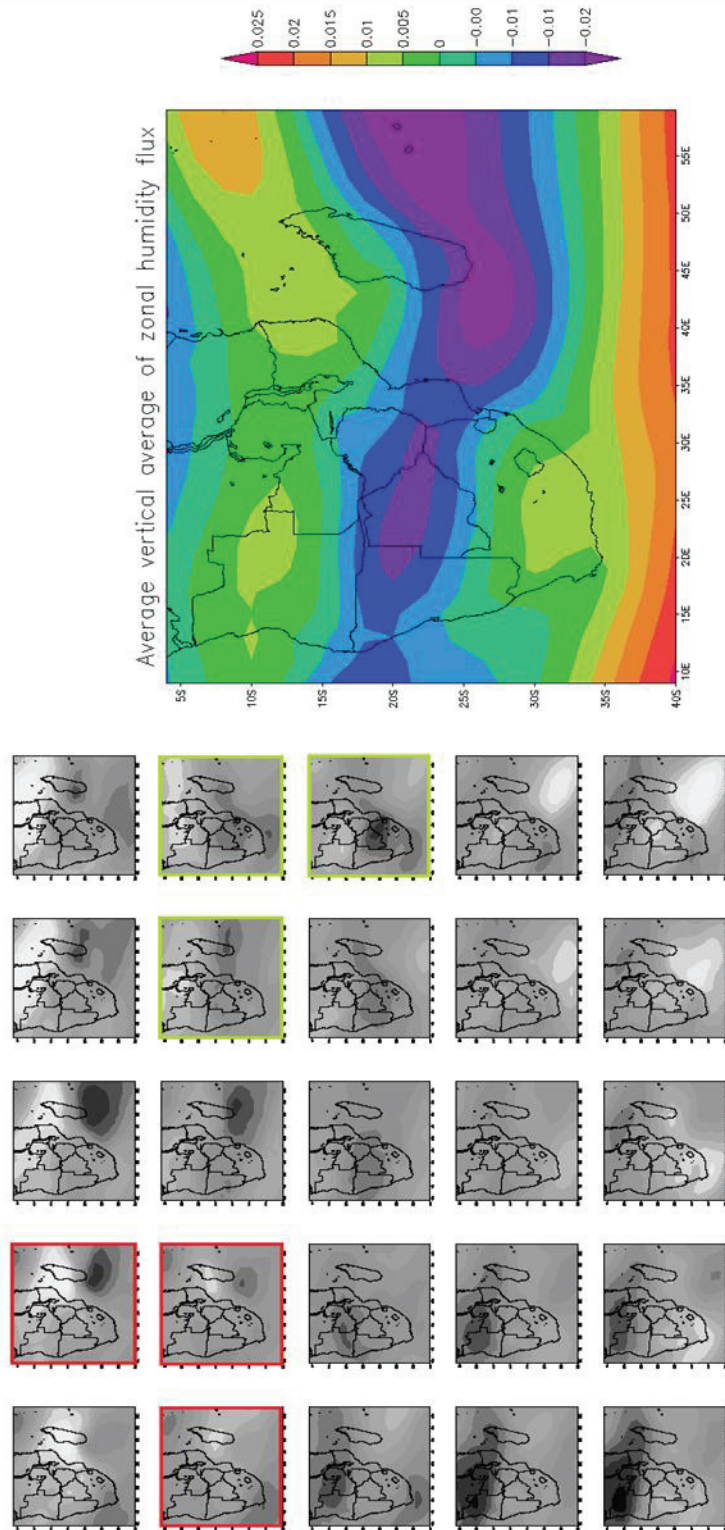


Fig. 8.6: The SOM for the vertical average of zonal humidity flux with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is also shown (right)

The situation, in terms of the deviations from the average, is reversed for the **non-landfalling** cases:

- allowing for the weakening of the mean easterly flux of humidity specifically to the southern parts of the Mozambique Channel and a westerly flux of humidity over most of the Mozambique Channel

- to the south of Madagascar weak to strong negative anomalies are found to occur, enhancing the mean easterly flux present over that region seen in the average values in Fig. 8.6 (right)

The vertical average of meridional humidity flux revealed similar results during the landfalling and non-landfalling events as the meridional winds and vertical average of meridional temperature flux. Fig. 8.7 shows the SOM for the vertical average of meridional humidity flux as well as the mean values. Together with the vertical average of zonal humidity flux, the existence of the anomalously anticyclonic versus cyclonic circulation during landfall versus non-landfall events through much of the depth of the troposphere is quite prominent.

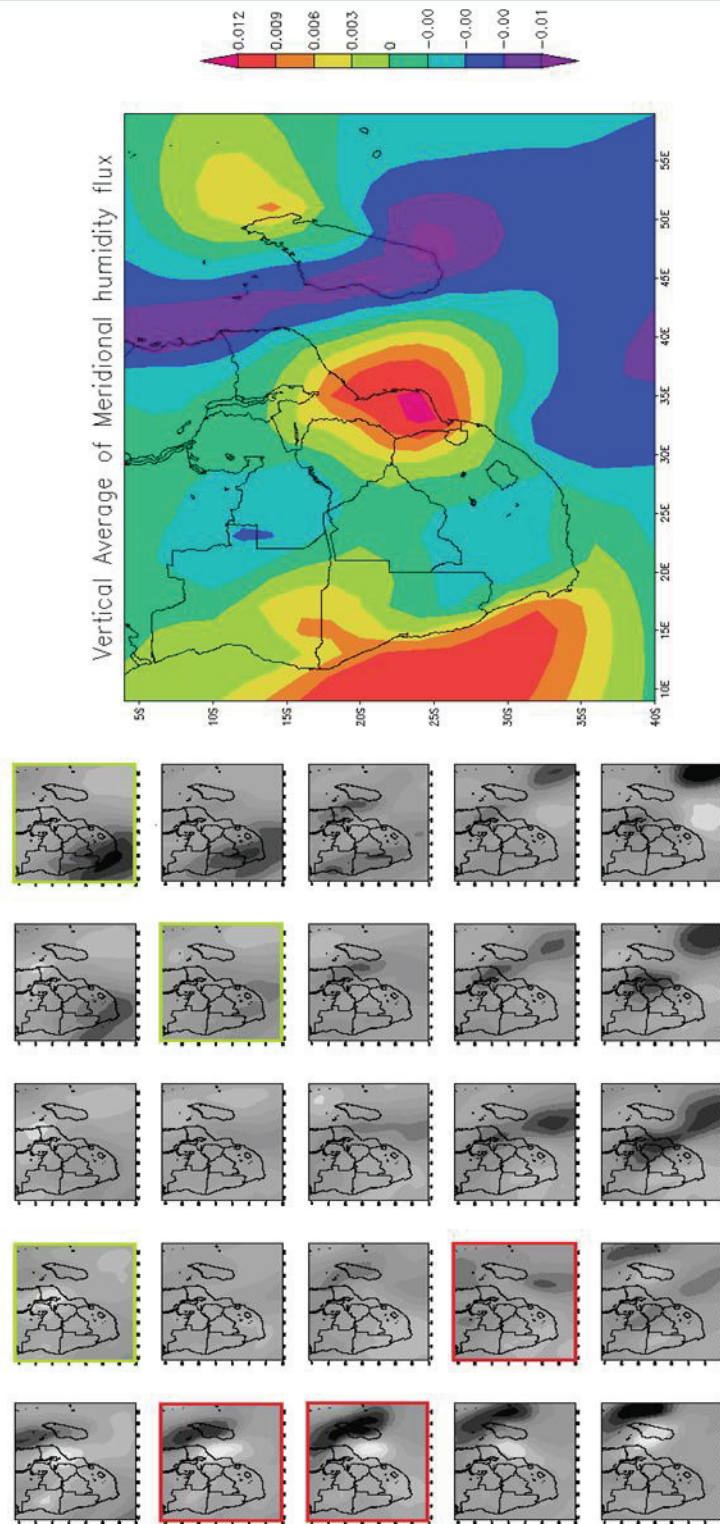


Fig. 8.7: Shows the SOM for the vertical average of meridional humidity flux with the green frame showing the top landfall classes and the red frame the top non-landfall classes. Anomalies shown are negative for dark grey to black and positive light grey to white. The average for the 26 tropical cyclone events is

8.3.7 Vertical structure of SOM-derived atmospheric signals related to landfall versus non-landfall events

The index enables the comparison between various levels of the atmosphere for the relative concentration of landfall/non-landfall event days in SOM nodes related to landfall/non-landfall events. Higher numbers of landfall/non-landfall event days classed into the landfall/non-landfall nodes give rise to a higher index value as opposed to a lower index value resulting from a more random distribution of landfall/non-landfall event days.

For geopotential height, there is not a large variance in the index value with respect to altitude at least for the levels considered from 1000 hPa to 100 hPa (Fig. 8.8). The mid-troposphere seems to reveal more importance than the lower and upper troposphere for geopotential height, while temperature becomes slightly more important towards the upper troposphere than lower troposphere.

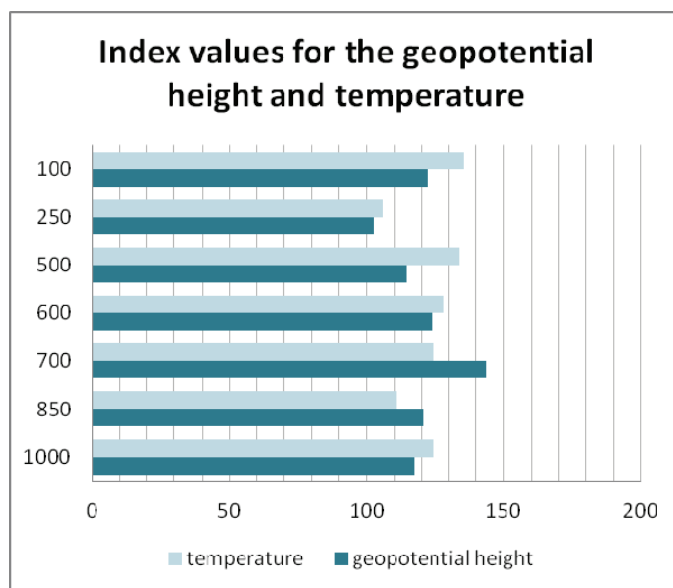


Fig. 8.8: The index values for geopotential height and temperature showing at which level the SOM has most significance in terms of tropical cyclone movement

The zonal and meridional winds' index values are shown in Fig. 8.9. The wind, according to the SOM analysis, seems to play a larger role between 600 hPa and 250 hPa than the regions. This is especially true for the zonal winds (dark blue bars). This is in accordance to findings regarding the importance of steering flow of the middle to upper troposphere for tropical cyclones in the SWIO region (Vitart et al., 2003; Reason and Keibel, 2004).

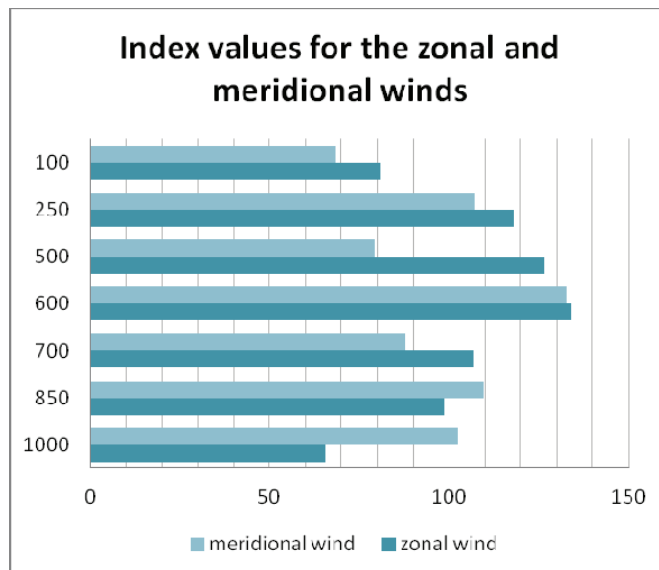


Fig. 8.9: The index values for the zonal and meridional winds showing at which level these variables have most significance in terms of tropical cyclone movement according to the SOM analysis

The index values for the relative humidity and mixing ratio can be seen in Fig. 8.10. The patterns associated with landfall/non-landfall events are more clearly organized from the 700 hPa to 250hPa levels, also including the steering levels. It seems that randomness or variation at smaller spatial and temporal scales of the values lower in the troposphere and generally very little moisture above 250 hPa results in weaker associations of anomalies at these altitudes to landfall/non-landfall over the AOI.

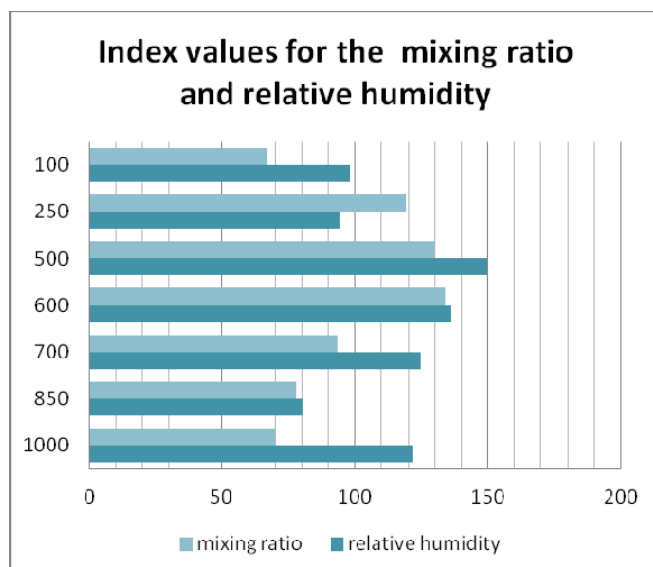


Fig. 8.10: The index values for the relative humidity and mixing ratios showing at which level the SOM has most significance in terms of tropical cyclone movement

8.4 Conclusion

Using SOMs, an objective analysis has been carried out to determine under what atmospheric conditions tropical cyclones were found to make landfall opposed to non-landfall. While it substantiates results in Ch. 7 based on average fields or case studies, it also highlights the importance of previous studies and provided some new findings. Previous studies showed that tropical cyclones were found to make landfall in the presence of a stronger easterly zonal steering flow opposed to a weak easterly flow where non-landfall was found to occur (Vitart et al., 2003). In addition it was also shown that during the landfall of tropical cyclone Eline in 2000, strong precursor conditions occurred, including a ridge to the south of the Mozambique Channel and an easterly steering current in the mid levels (Reason and Keibel, 2004). These two conditions shown to favour landfall of tropical cyclones are substantiated in this particular study. A ridge to the south of the Mozambique Channel is found to occur in the SOM analysis throughout the troposphere during the landfalling events. This ridge is found to be displaced to the north with height while from about 500 hPa upwards a trough occurs to the southwest of the subcontinent. The strengthening of the ridge allows for the strengthening of the easterly flow over the Mozambique Channel. In many of the non-landfalling cases, however, the occurrence of a trough is favoured to the south and over the Mozambique Channel, weakening the easterly flow.

The SOM analysis shows the importance of the enhanced easterly flow in the middle to upper troposphere during landfall of tropical cyclones, which was reported in the literature. This enhanced easterly flow allows for an enhanced vertical average of zonal easterly flux of humidity and temperature over the Mozambique Channel. During non-landfall a weakened easterly flow to more westerly flow is found to occur, weakening the vertical average of zonal temperature and humidity flux over the Mozambique Channel. Both the zonal and meridional winds and fluxes of temperature and humidity have stronger steering flow associated with anticyclonic anomalies during landfall and cyclonic during non-landfall to the east of the subcontinent.

Atmospheric moisture during the tropical cyclone events has a more zonal distribution during landfall with anomalously higher values over the eastern parts of the subcontinent. This pattern could be associated with the steering anomalies with enhanced easterly flow over the Mozambique Channel in the middle to upper troposphere. During the non-landfalling events an enhanced discontinuity in the moisture variables occurred over the Mozambique Channel with lower values over the eastern parts of the subcontinent and higher to the east. This could be associated with the weakening of the easterly flow over the Mozambique Channel.

The existence of the tropospheric characteristics of the zonal steering current, moisture and temperature distribution, specifically at tropospheric levels pointed out in the analysis, could signal strong precursor conditions for the landfall or non-landfall of tropical cyclones on multi-temporal scales during the summer months. The anomalies identified could also be used in the seasonal forecasting of landfall probabilities through the analysis of expected large-scale anomalies over the southern African region.

Chapter 9

Decadal-scale Dynamics as Related to Landfall of Tropical Systems and Subsequent Rainfall Contribution to the Limpopo River Basin

9.1 Introduction

Three multi-year periods have been identified where the contribution towards rainfall by tropical systems from the SWIO were larger than during the surrounding periods over the Limpopo River Basin (Malherbe et al., 2012; Fig. 9.1). These were during the late 1950s, the mid-1970s and the period from 1995 to 2003, coinciding with peaks in the Dyer-Tyson cycle, a quasi bi-decadal oscillation over the northeastern parts of South Africa and most of southern Africa between 20 and 30°S (Tyson et al., 2002).

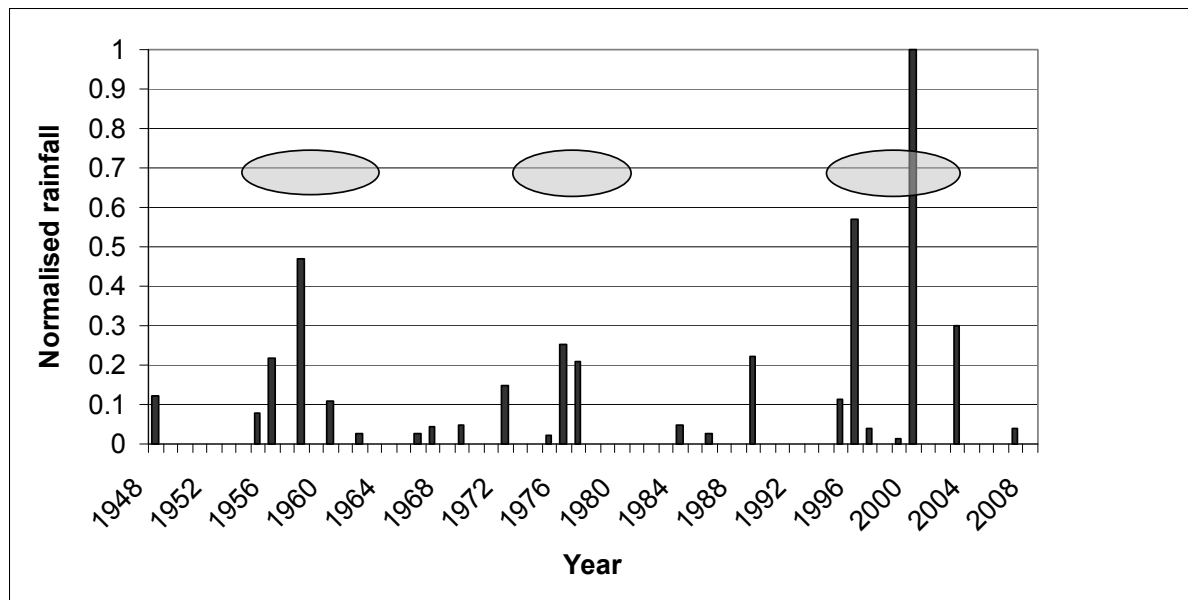


Fig. 9.1: Scaled annual rainfall totals contributed by tropical systems from the SWIO at four locations over the period 1948-2008 (similar results were obtained for Pafuri and Entabeni)

The mechanism responsible for the modulation of rainfall over the southern African region at the bi-decadal time scale has been related mostly to the influence from the west wind regime through the effect of standing waves on the frequency or intensity of synoptic systems over and surrounding southern Africa. A positive relation have been noted between above-normal rainfall and increased pressures in the region of the subtropical ridge associated with zonal standing wave 1 (about 10°W over the Atlantic Ocean) and lower pressures over the subcontinent (Tyson, 1981) when the summer half year is considered. Furthermore, the combined effect of waves 1 and 3 results in a ridge between 0° and 45°E at about 35°S (Trenberth, 1980, from Tyson, 1981), while the position of the first ridge of standing wave 3, located near 45°E, varies with a 10-11 year period throughout the troposphere. The ridge of wave 1 has also been shown to be influenced by ENSO (Trenberth, 1980, from Tyson, 1981). The influence of ENSO on the rainfall over the northeastern parts of South Africa is also modulated by the bi-decadal cyclicity evident in the rainfall time series with El-Niño being

wetter/drier and La Niña being drier/wetter during periods of above-normal/below-normal rainfall (Kruger, 1999; Reason and Rouault, 2002) also evident in the dry-spell frequency during El Niño and La Niña summers over the Limpopo River Basin specifically (Reason et al., 2005). Inter-decadal modulation of the ENSO signal is not only confined to southern Africa (e.g. Gershunov and Barnett, 1998).

More specifically, the circulation anomalies in the southern African region associated with periods of above-normal rainfall over the summer rainfall region include positive anomalies in the geopotential heights especially towards the southwest and to a lesser extent towards the southeast of the subcontinent throughout the lower to middle troposphere and negative pressure deviations over the southern African interior dominated by air flow from the north (Miron and Tyson, 1984). These anomalies have also been pointed out for wet conditions over the Limpopo River Basin specifically (Hofmeyr and Gouws, 1964). Circulation anomalies during wet conditions in January (late summer) over the interior further give rise to enhanced advection of moist air from the SWIO westward and southwestward into the southern African interior (D'Abreton and Lindesay, 1993) around an anomalously anticyclonic circulation to the southeast of southern Africa in the lower to upper troposphere (Jury, 1996; Matarira and Jury, 1992). This anticyclonic anomaly to the southeast is replaced by a cyclonic anomaly during drought conditions over southern Africa (Richard et al., 2001) and specifically also over parts of the Limpopo River Basin during non-ENSO-related droughts and to a lesser extent also ENSO-related droughts (Mulenga et al., 2003). Furthermore, there is a southward shift in the occurrence of westerly winds responsible for advection of dry air from the Atlantic Ocean towards the far southern parts of the subcontinent (D'Abreton and Lindesay, 1993) during wet conditions while increased easterly flow over the lower latitudes exhibits pulses of convection moving into the subcontinent (Jury and Nkosi, 2000). The easterly flow further increases atmospheric moisture from the SWIO over the subcontinent and, combined with increased barotropic instability, is responsible a higher frequency and higher growth rate of tropical disturbances (Mulenga et al., 2003). The upper air circulation anomalies during dry late summer seasons suggest a strengthening and northward displacement of the jet stream around 200 hPa leading to a displacement of cyclonic systems to the north while the enhanced westerly flow over the subcontinent causes a reduction in the influence of tropical systems over the interior of southern Africa (Richard et al., 2001).

The movement of tropical cyclones is dictated by the environmental steering flow defined to be the vertical weighted average or part thereof, mostly concentrating on the 700 hPa to 500 hPa level (Jury and Pathack, 1991; Franklin et al., 1996) located between 850 hPa and 200 hPa (Neumann, 1976), Coriolis force as well as interactions with vorticity in the surrounding environment influenced by coupling between latent heat release and vertical wind shear (Chan, 2005). Over the Indian Ocean, a southward shift and weakening of upper air westerlies, increased easterly trade winds and a strengthened Hadley circulation are characteristic of summers with more tropical cyclones over the SWIO (Jury, 1993) while an anomalously strong easterly zonal flow to the east of the subcontinent in the subtropics further increases the chances of tropical cyclone landfall (Vitart et al., 2003). It is expected that contrasting atmospheric circulation patterns associated with anomalously wet and dry conditions and a decadal variability in the prevalence of these atmospheric circulation regimes in the southern African region, manifesting in the position of the subtropical jet as well as the strength of easterlies towards the lower latitudes, magnitude of vertical wind shear and barotropic instability over the subcontinent, will alter the landfall tendency or the zonal nature of tracks after landfall of tropical cyclonic systems. These changes will then be responsible for modulating the influence of these types of systems in terms of the

contribution of rainfall over large parts of the subcontinent and more specifically also the direct contribution to rainfall over the Limpopo River Basin. The importance of the contribution to rainfall by tropical cyclonic systems from the SWIO over parts of the Limpopo Province and a positive correlation between the rainfall contributed by these systems and the quasi bi-decadal variation in rainfall over the region have been described recently (Malherbe et al., 2012). Severe flooding did occur as a direct result of tropical systems moving into the Limpopo River Basin during 1996 and 2000 during the previous expected maximum (Dyer and Tyson, 1977; Alexander, 1995) in the quasi bi-decadal rainfall cycle over the northeastern interior of South Africa. Except for the two major flood events, several tropical systems making landfall from the SWIO were responsible for significant rain over large parts of the Limpopo River Basin during the period 1995 to 2003. Since then, even though landfall did occur and there were three relatively strong La Niña events (2008, 2011, 2012), very little rain had been contributed by these systems over the Limpopo River Basin, showing that, at least for the direct contribution to rainfall by tropical cyclonic systems from the SWIO, the bidecadal rhythm is still present. This research describes the synoptic circulation anomalies associated with the surge in direct influence of tropical cyclonic systems from the SWIO over the Limpopo River Basin in the context of the latest maximum in the bi-decadal Dyer-Tyson cycle.

9.2 Data and Methodology

NCEP Reanalysis data (section 3.2) for the years 1979 to 2012 are used in the analysis of large-scale synoptic fields in the southern African region. The 2.5° gridded dataset is analyzed over AOI 1 (Fig. 3.2). The geopotential heights at all the standard pressure levels from 850 hPa to 30 hPa are considered.

Daily data are considered only for the first three months (January to March) of each year during this period. Historically, since 1948, landfalling systems causing significant rain over the eastern parts of southern Africa only occurred during these three months (Ch. 5). Rainfall data for the period, representing parts of the Limpopo River Basin, are derived from the six stations indicated in Fig. 3.1.

A Self-Organizing Map (SOM) analysis is performed to classify synoptic states at each of the levels mentioned earlier separately. By studying the synoptic patterns associated with the nodes of the SOM analysis and taking the error made when associating certain synoptic conditions with the SOM nodes into account, it was decided to use a 25-node – 5x5 – SOM. The periods surrounding landfall are identified in the time series of data and the synoptic classes resulting from the SOM analysis were evaluated in terms of the favourability for landfall and further inland movement. The favourability is expressed by counting the number of landfall event days associated with each node and expressing that as a percentage of the total number of days represented by that node. For each level, the resulting classes are ranked in terms of their favourability for landfall of tropical systems after which the time series are studied to identify the large-scale climate variability (since the mid-1970s) that gave rise to the oscillatory behaviour of rainfall contributed over the eastern parts of southern Africa by tropical systems from the SWIO.

The time series of the yearly occurrence of the complete set of SOM nodes, ranked in terms of favourability for landfall, are also constructed to study the variation in the occurrence of synoptic states, with a focus on the favourability for landfall, over the period. Attention is also given specifically to the occurrence and clustering of various types of synoptic conditions of relevance to landfall. The least favourable and most favourable patterns are

studied and four new classes are identified into which the existing 25 SOM nodes are grouped. Grouping is done by doing a new classification of the 25 nodes into the aforementioned four patterns using the SOM software. From the resulting groups representing the most favourable and least favourable conditions, an index is derived throughout the time series. This is done on a yearly basis for the period January to March. It is calculated by taking the difference between the number of occurrences of the group of favourable nodes and the number of occurrences within the group of unfavourable nodes. The time series of this index on a yearly basis is also calculated for each standard pressure level mentioned earlier. The index is used to identify whether an oscillatory pattern can be identified and whether this pattern coincides with the oscillatory pattern found in the rainfall contributed by tropical systems from the SWIO over the eastern parts of southern Africa. It is also done to identify on which levels of the troposphere the oscillatory pattern occurs.

Finally, several tropospheric elements are contrasted, for the period identified by rainfall data and synoptic patterns, to the rest of the period from 1979 to 2011.

9.3 Results and Discussion

9.3.1 Synoptic patterns and time series

The SOM analysis performed on the heights of the standard pressure levels from 1000 hPa to 100 hPa yielded nodes that made the association with landfall possible. Above the 100 hPa level, the nodes were dominated by trends over the time series (1979-2011). The results of the SOM analysis for the 850 hPa, 250 hPa and 100 hPa levels are shown in Figs. 9.2 to 9.4. The Sammon map, together with an indication of favourability for each of the nodes for landfall, is also indicated. The distances between the nodes in the Sammon map is an indication of how much the various synoptic states associated with each node differ from the synoptic states associated with surrounding nodes in terms of distribution of deviations and size of deviations.

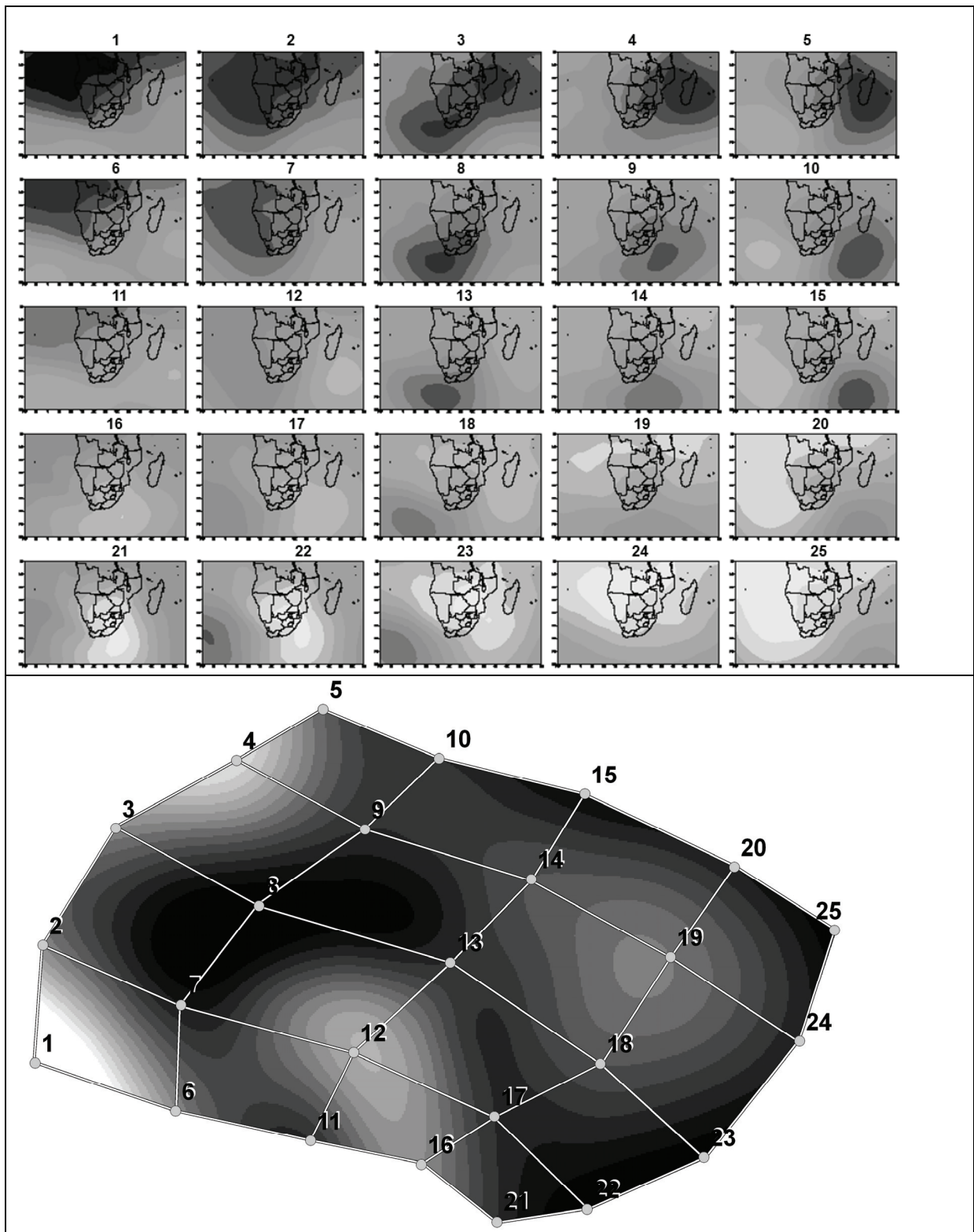


Fig. 9.2: SOM based on the standardized values with lighter (darker) shades showing positive (negative) deviations for the 850 hPa level (top) and Sammon map (bottom) showing the occurrence of days during landfall events within the identified nodes with lighter shades indicating a higher number of days with landfall events

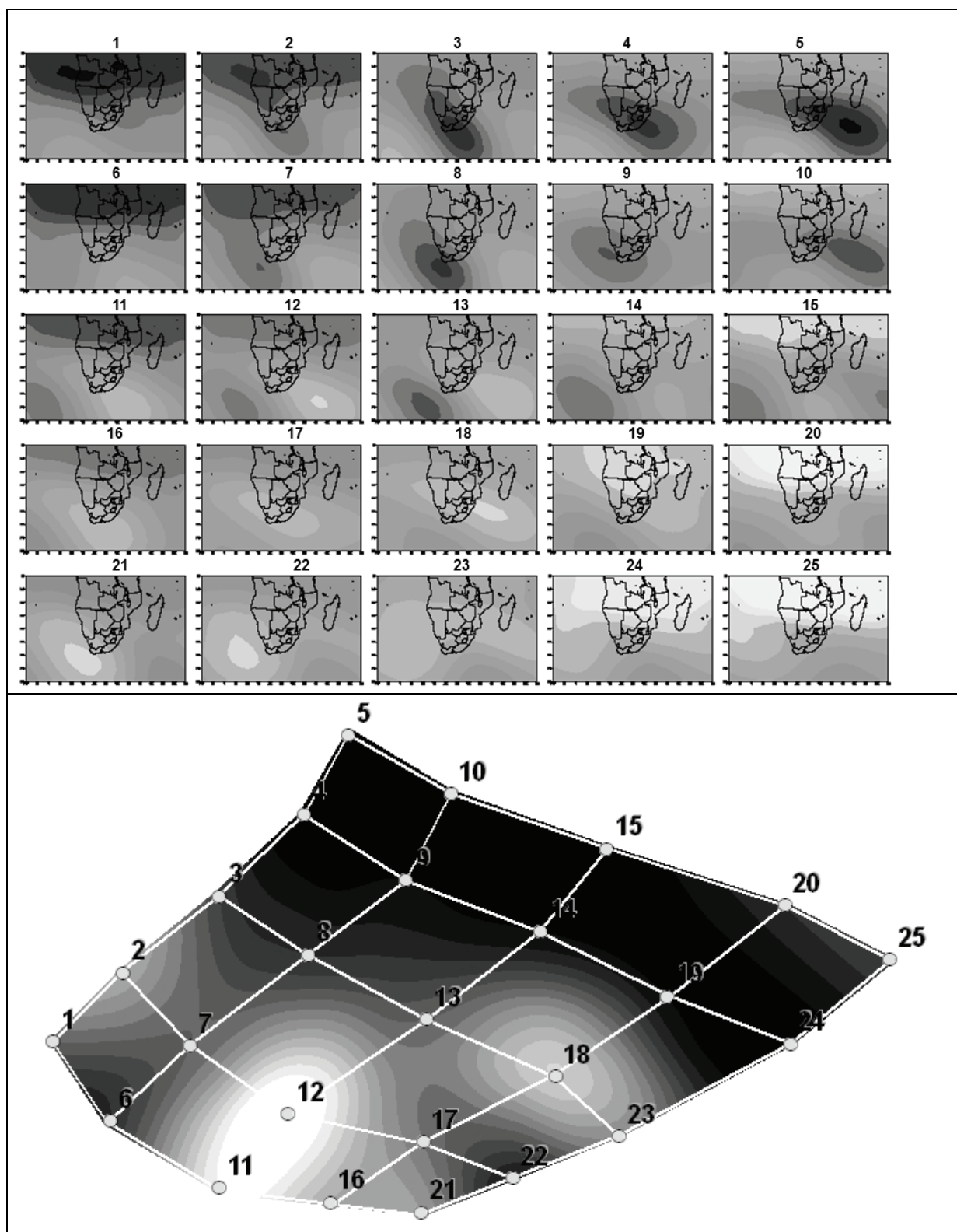


Fig. 9.3: SOM based on the standardized values with lighter (darker) shades showing positive (negative) deviations for the 250 hPa level (top) and Sammon map (bottom) showing the occurrence of days during landfall events within the identified nodes with lighter shades indicating a higher number of days with landfall events

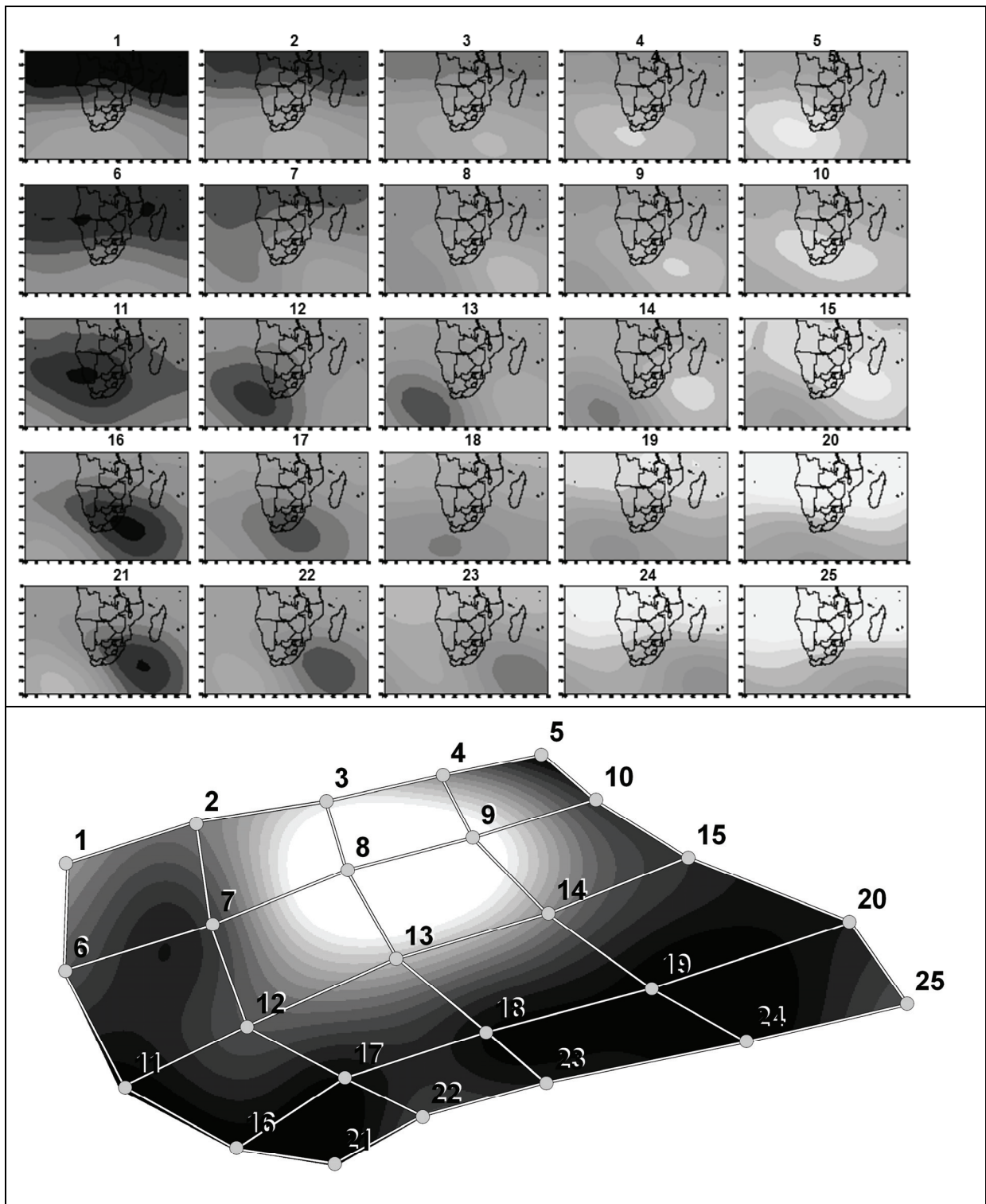


Fig. 9.4: SOM based on the standardized values with lighter (darker) shades showing positive (negative) deviations for the 100 hPa level (top) and Sammon map (bottom) showing the occurrence of days during landfall events within the identified nodes with lighter shades indicating a higher number of days with landfall events

The areas in each SOM that represent nodes that are relatively more favourable for landfall conditions are indicated in the Sammon maps in the lighter shades with the same grey scale used at all levels. With increasing height, there is an increase in the concentration of favourable with a more random distribution of favourable versus unfavourable nodes at the

lower levels. The following are the main categories identified from the SOMs with their relation to landfall indicated.

The patterns at the 850 hPa to 100 hPa levels that are associated with landfall events are the following:

- negative deviations in height to the north of southern Africa associated with weak or strong positive deviations to the south of the country
- positive deviations towards the southeast of the subcontinent
- positive deviations towards the southeast of the subcontinent and negative deviations towards the southwest of the subcontinent

The patterns at the 850 hPa to 100 hPa levels that are not associated with landfall events are the following:

- positive deviations in height to the north of southern Africa associated with weak or strong negative deviations to the south of the country
- strong negative deviations towards the southeast of southern Africa
- strong positive deviations over the subcontinent and towards the north

The patterns that do not have a strong association with landfall or no landfall are:

- strong negative deviations over the southern parts of the subcontinent and/or towards the southwest of that with weak deviations towards the north
- strong positive deviations towards the southwest of southern Africa with weaker deviations elsewhere

The time series for the nodes for the 1976-2011 period is shown for the 250 hPa level in Fig. 9.5. On the x-axis are the years from 1976 to 2011. On the y-axis are the nodes with the lighter shades of grey indicating higher numbers of occurrences. The nodes, in the vertical, have been arranged so that those that are associated with more landfall events are concentrated towards the bottom of the graph while nodes less associated are indicated towards the upper parts of the graph. On the right-hand side, the pattern associated with the nodes on the diagram is shown. El Niño and La Niña event years are also indicated in black and white lines respectively. The blue circles at the top represent significant rainfall events (responsible for more than 50 mm in one day at any of the six stations in Fig. 3.1) in dark blue and rainfall of a lighter nature in light blue. Numbers within the diagram indicate the number of the node as per Fig. 9.4.

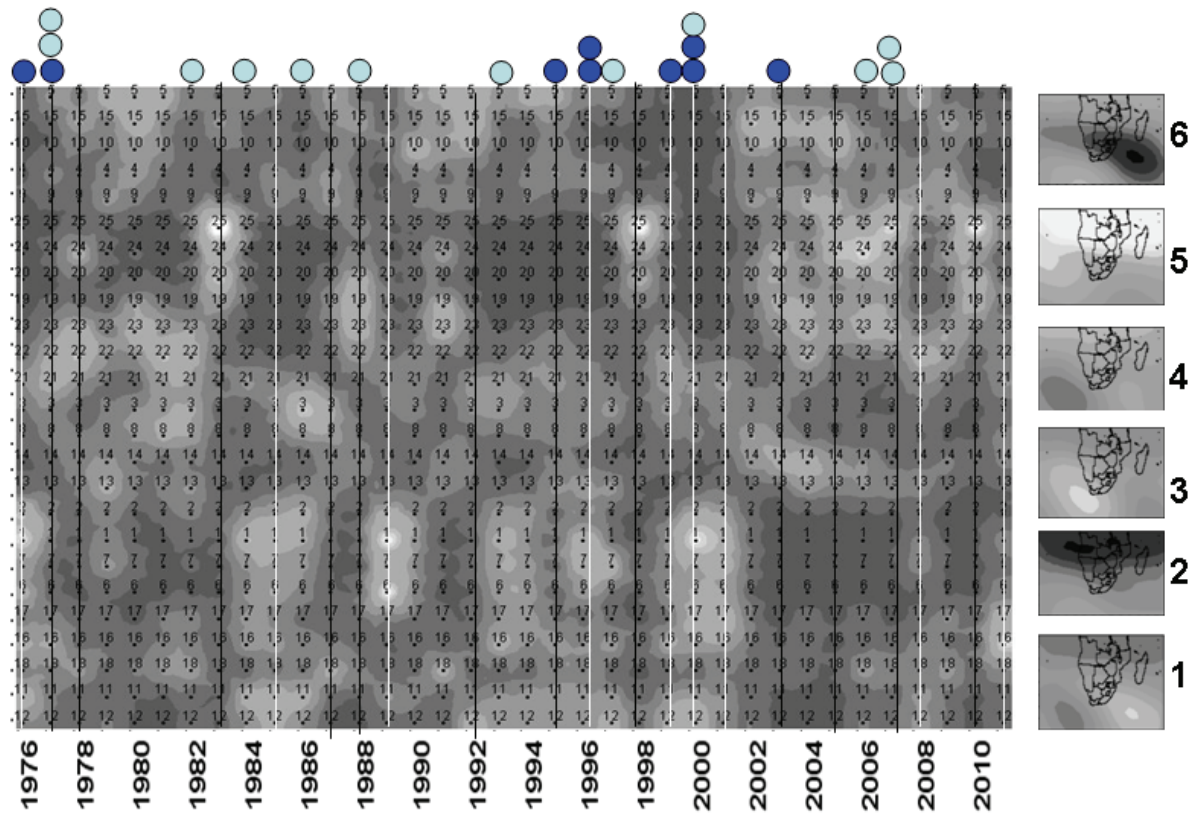


Fig. 9.5: Time series of the occurrence of days within the 25 nodes of the SOM for the 250 hPa level. All landfall events responsible for at least some rain over any of the 6 stations used for rainfall analysis have been indicated with circles and those responsible for more than 50 mm in one day at any of the stations with dark circles

From the time series of all the SOM nodes at 250 hPa, the patterns clearly associated with ENSO can be found in class 2 and class 5, with a focus on the north-south dipole of height anomalies in the southern Africa region. Classes 1 and 2 are more abundant during years with more landfalling systems contributing rainfall over the Limpopo River Basin being the north-south dipole and also the southwest-southeast dipole. Class 2 is also the class very closely related to ENSO cold events. Likewise, classes 5 and 6 are more abundant during years with no landfalling systems contributing rainfall over the Limpopo River Basin being the reverse of classes 2 and 1, with class 5 very closely related to ENSO warm events. Classes 1 and 6 are not as closely related to ENSO as classes 2 and 5, with clearly a higher (lower) concentration of class 1 (6) during years associated with landfall and greater contribution to rainfall over the Limpopo River Basin.

The dipole represented by classes 1 and 6 (north/south dipole sensitive to ENSO) will now be referred to as Pattern A while the East/West dipole to the south of southern Africa will be referred to as Pattern B.

The 25 nodes, divided into 4 groups, show variation in their association with landfall and rainfall contribution by tropical systems over the Limpopo River Basin, as shown in Fig. 9.6 for all the standard levels from 1000 hPa to 100 hPa.

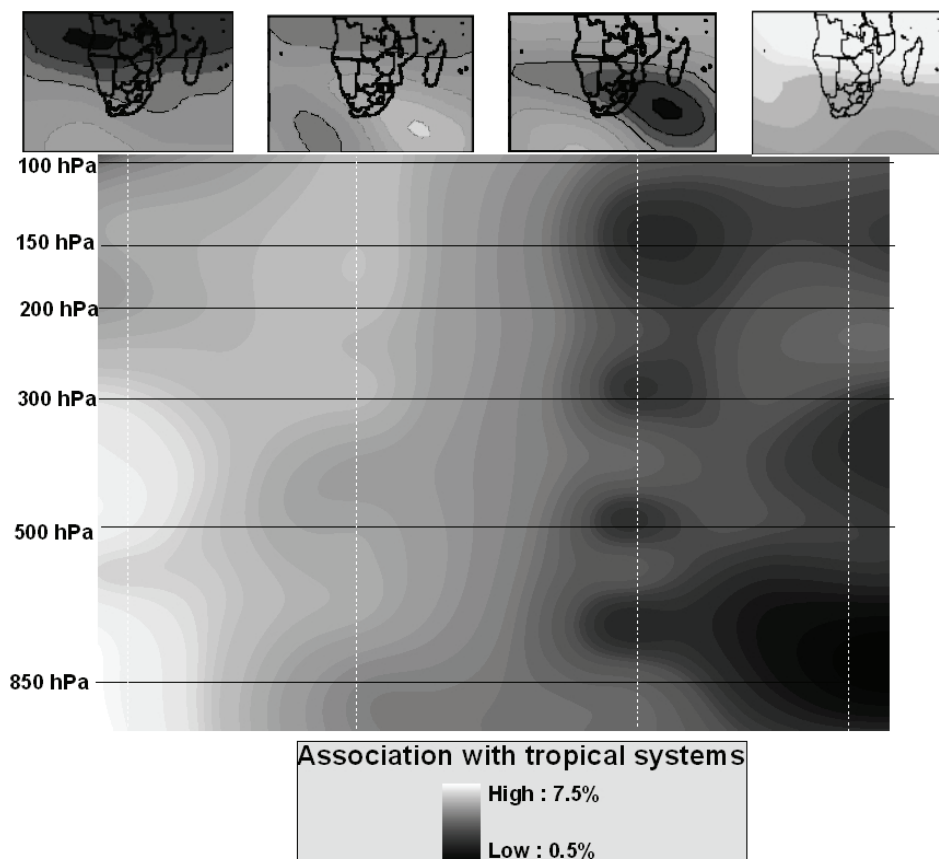


Fig. 9.6: Variation in concentration within each SOM of landfall event days per pressure level

From left to right at the top of Fig. 9.6, the groups associated more strongly with La Niña (Pattern A Positive), Landfall (Pattern B Positive), Non-landfall (Pattern B Negative) and El Niño (Pattern A Negative) are represented. The standard pressure levels are represented in the vertical from bottom (1000 hPa) to top (100 hPa). The concentration of landfall days is indicated in shades of grey with higher (lower) concentration shown in lighter (darker) shades. At the lower levels (from 1000 hPa to about 400 hPa), Pattern A is most strongly associated with favourability for landfall (largest contrast found between the far left and far right). From 300 hPa to 100hPa, the strongest signal comes from Pattern B (largest contrast between the middle left and middle right).

Break-point analysis based on the average rainfall contributed by tropical systems from the SWIO over the Limpopo River Basin indicates that the period 1993-2001 can be considered to be different from the rest of the time series spanning the period 1979-2011.

The variation in time on all the levels up to 50 hPa of the occurrence of these nodes can be used as an explanation for the much higher concentration of landfall and a larger contribution to rainfall over the eastern parts of the subcontinent during the 1993-2001 period as opposed to the periods surrounding that. Fig. 9.7 shows the average yearly total rainfall contributed by tropical systems from the SWIO over the six stations in the northeast of South Africa and southern Zimbabwe (Fig. 3.1), the time series of the balance of positive and negative occurrences of Pattern B as average over the entire 250-100 hPa depth and a decomposition of the components of that difference index (showing the time series of the positive and negative occurrences of Pattern B). This pattern reaches a maximum during the 1993-2001

period during which an enhanced contribution to rainfall by tropical systems from the SWIO occurred together with the previous maximum in the theoretical Dyer-Tyson cycle. This variation in time is not displayed by Pattern A (not shown). Fig. 9.8 shows the positive and negative contributions of Pattern B and how it is reflected through the troposphere at least up to 100 hPa.

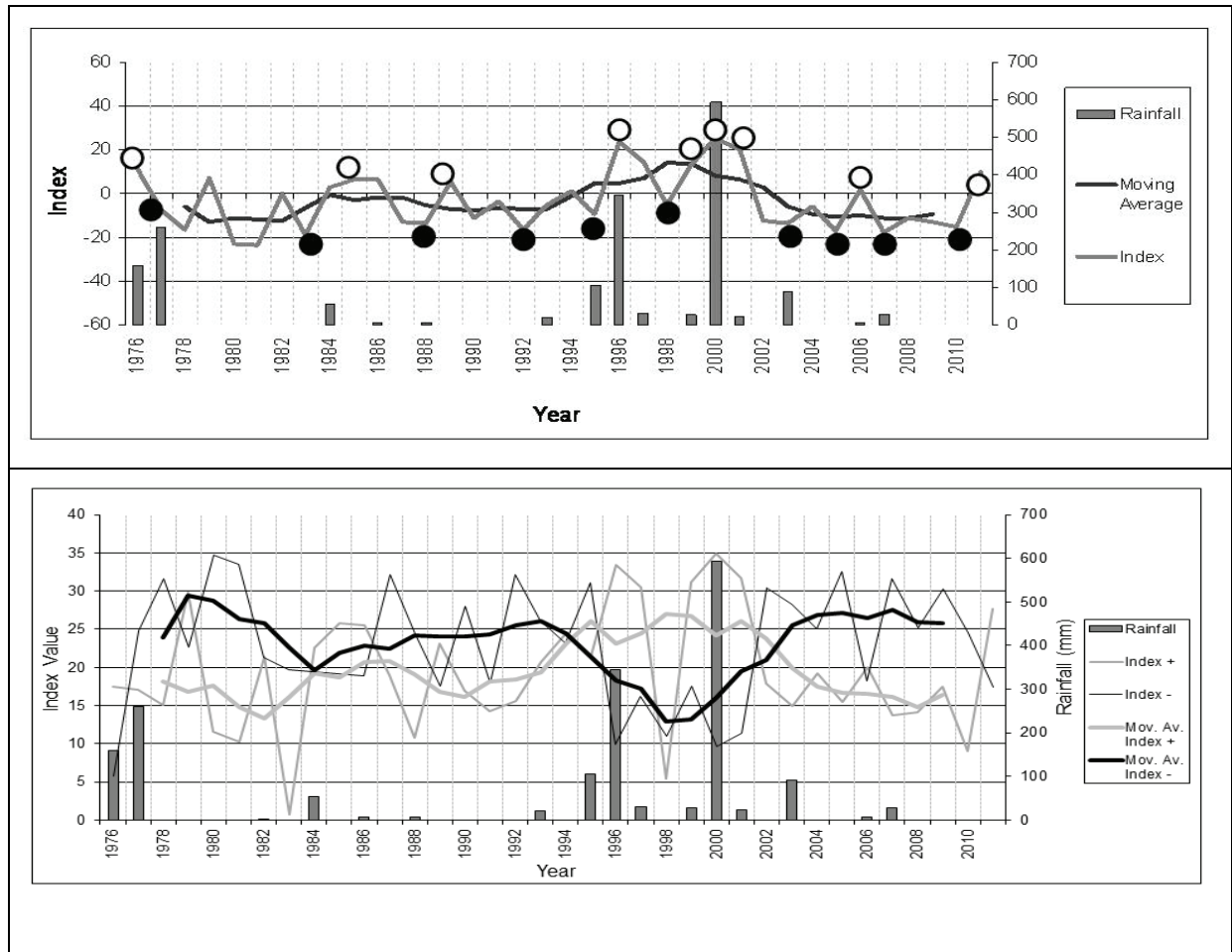


Fig. 9.7: Time series of the difference between Pattern B Positive and Pattern B Negative (line) and annual rainfall (bars), Pattern B Positive and Pattern B Negative (bottom)

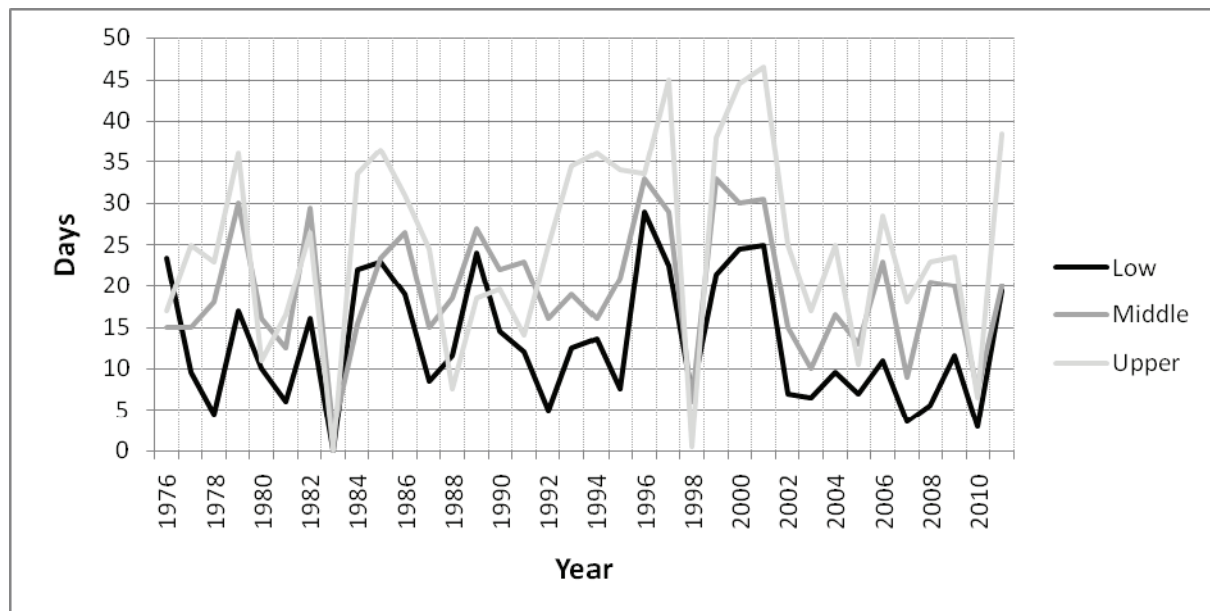
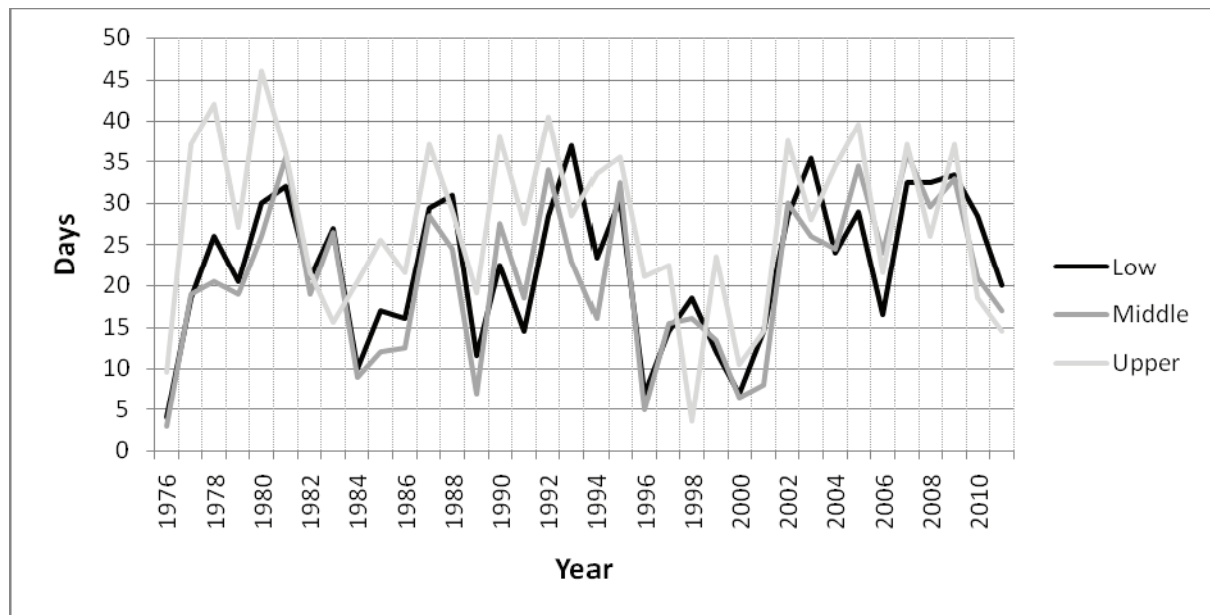


Fig. 9.8: Time series of the yearly number of days with Pattern B Negative (top) and Pattern B Positive (bottom) of the subcontinent for the lower (850 and 700 hPa), middle (600 and 500 hPa) and upper (250 and 100 hPa) levels in the atmosphere

The variation in pressure pattern is visible throughout the troposphere up to 100 hPa. Apart from the larger maximum in the 1993-2001 period, there is also a secondary maximum around 1984-1986. Interestingly, there was also an increase in the frequency of landfall events contributing some rain over the Limpopo River Basin during that period (e.g. Domoina), but the interior tracks of these systems limited their contribution to rainfall. This secondary peak in Pattern B was, however, much weaker than the 1993-2001 peak and other factors contributing to the more zonal tracks during the 1993-2001 period were not present during this smaller peak around 1985.

9.3.2 Anomalous patterns during the previous maximum

A change-point analysis performed on the rainfall contributed by tropical systems from the SWIO (as recorded by the six stations in Fig. 3.1) suggests that the period 1993-2001 was

anomalous to the rest of the time series in terms of the influence of tropical systems over the Limpopo River Basin.

Fig. 9.9 shows the difference calculated between 1993 and 2001 and the other years in the 1979-2011 period for the zonal wind to the east of southern Africa.

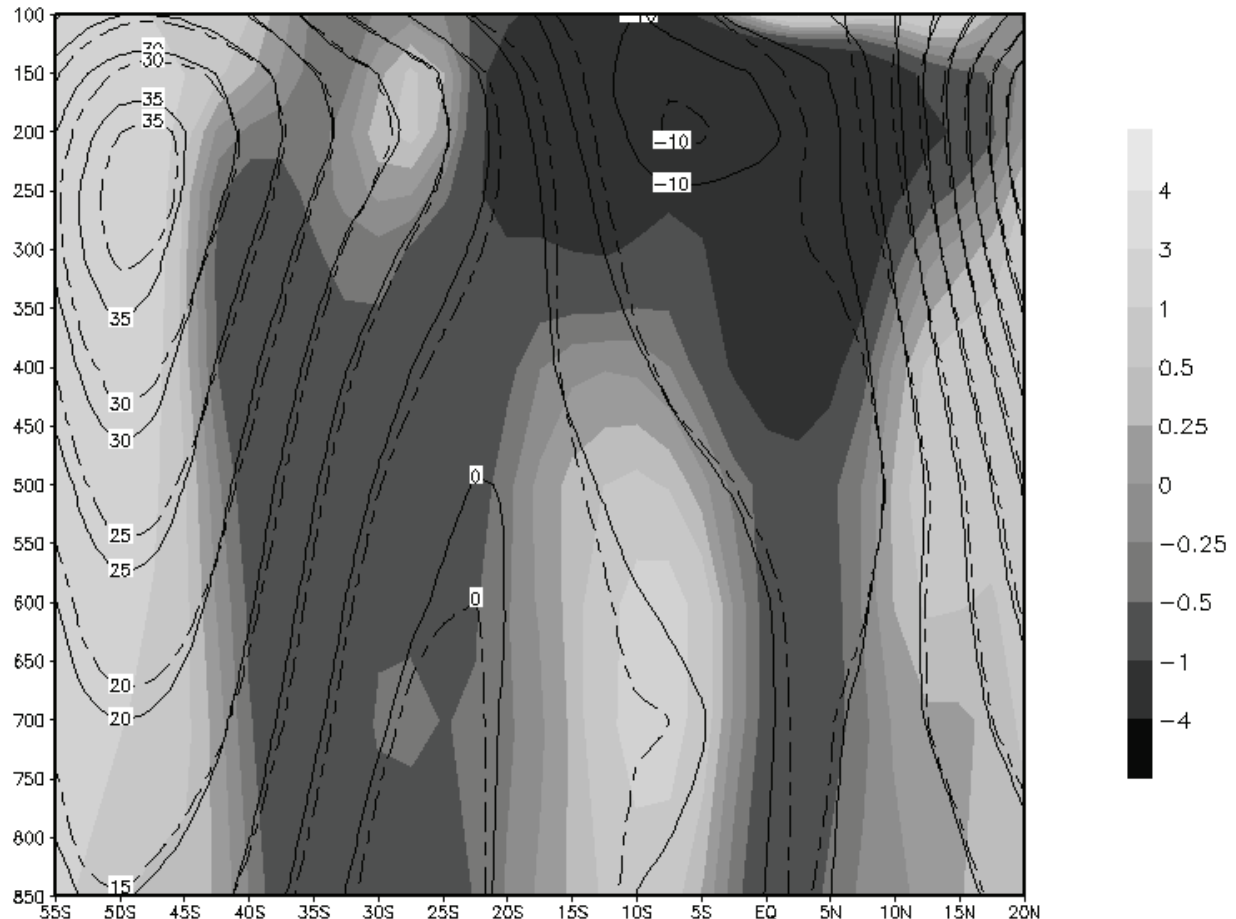


Fig. 9.9: Average zonal wind along 40°E during the 1993-2001 period (solid contours), 1979-1992 and 2002-2011 periods (dashed contours) and the difference (shaded)

The zonal flow, as altered during the 1993-2001 period, displays the following characteristics:

- stronger jet stream towards 50°S and also 35°S at 250-200 hPa
- southward contraction of the jet at 30°S
- stronger equatorial upper jet
- intensification of subtropical easterlies throughout the column
- stronger equatorial westerlies from the surface to 400 hPa

This shows a broader Hadley cell during this period. The strengthened easterlies over the subtropics in the lower to middle troposphere indicates a strengthened westward steering flow for tropical systems from 700-500 hPa (Fig. 9.10).

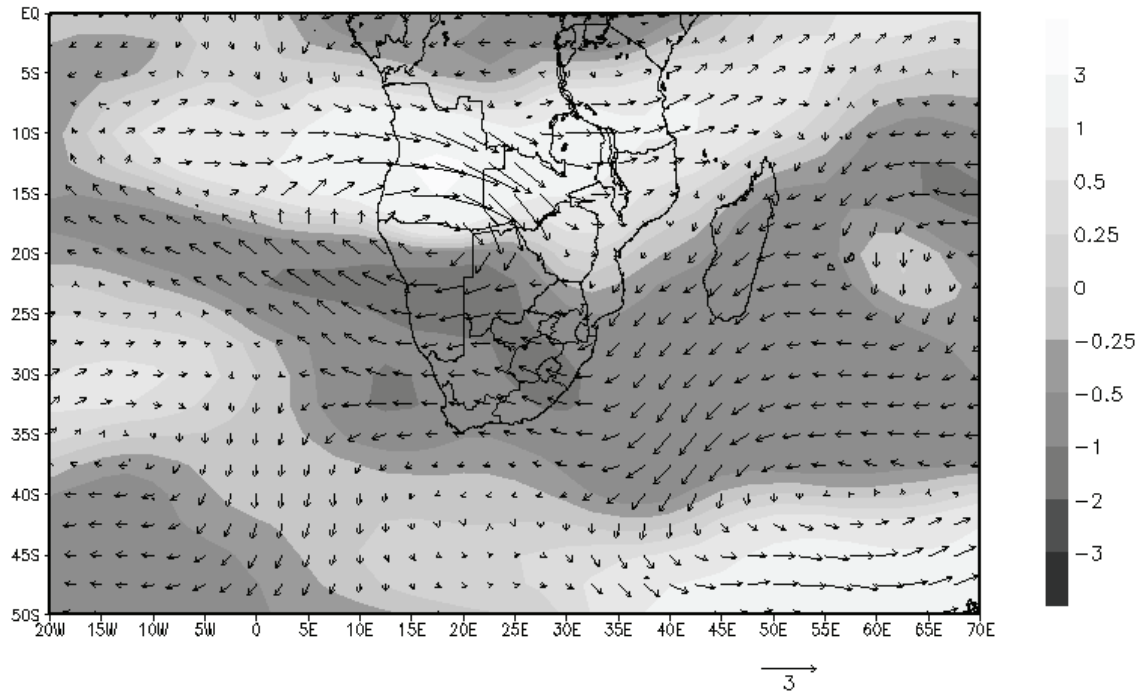


Fig. 9.10: Difference wind vectors of the 700-500 hPa flow for the period 1993-2001 with the 1979-1992 and 2002-2011 periods with the difference in zonal component shaded

The mid-level cyclonic anomaly over northern Namibia/southern Angola represents the deepening of the Angola low, located to the north of the strengthened subtropical easterlies associated with the broadening of the Hadley cell. In the upper air, the strengthened westerly subtropical jet at 50°S occurs to the north of a cyclonic anomaly at 15°W and to the south of an anticyclonic anomaly to the southeast of southern Africa (Fig. 9.11).

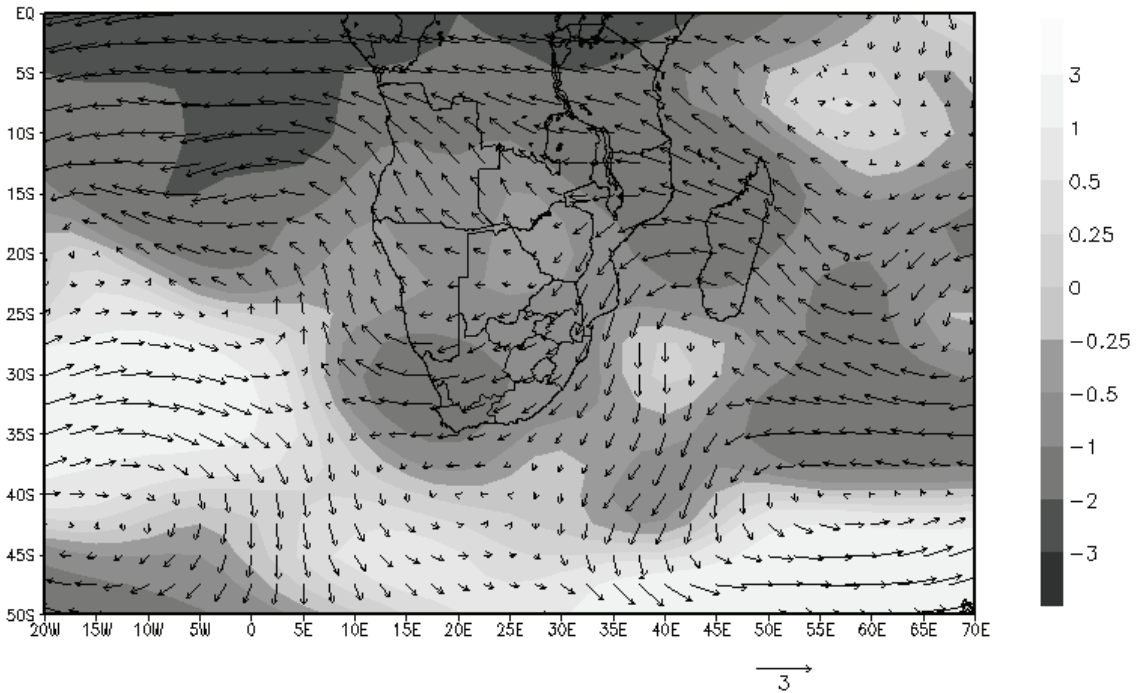


Fig. 9.11: Difference wind vectors of the 250 hPa flow for the period 1993-2001 with the 1979-1992 and 2002-2011 periods with the difference in zonal component shaded

Lower level convergence over Africa is displaced further south (Fig. 9.12), showing possibly a southward displacement of the ITCZ, coupled with increased moisture content over the middle layers of the atmosphere (not shown).

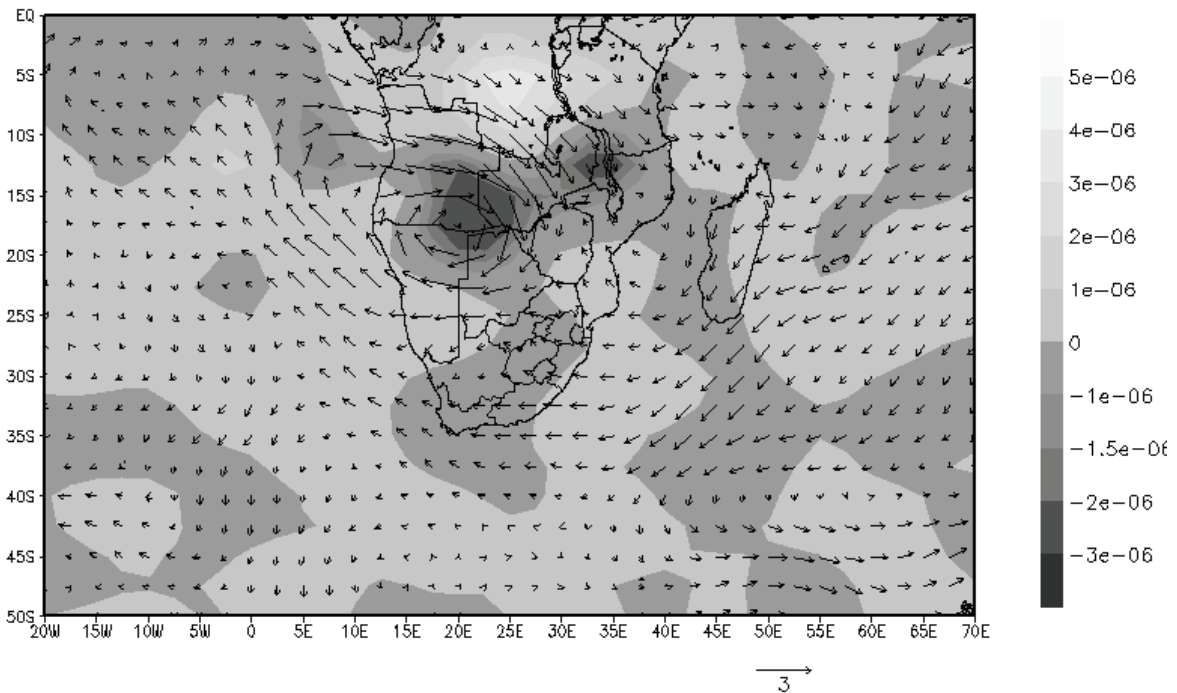


Fig. 9.12: Difference on 850 hPa convergence and difference in wind vectors for the period 1993-2001 with the 1979-1992 and 2002-2011 periods with the difference in zonal component shaded

Relative vorticity decreases in the boundary between intensified easterlies to the south and intensified westerlies to the north, roughly between 10 and 20°S (Fig. 9.13).

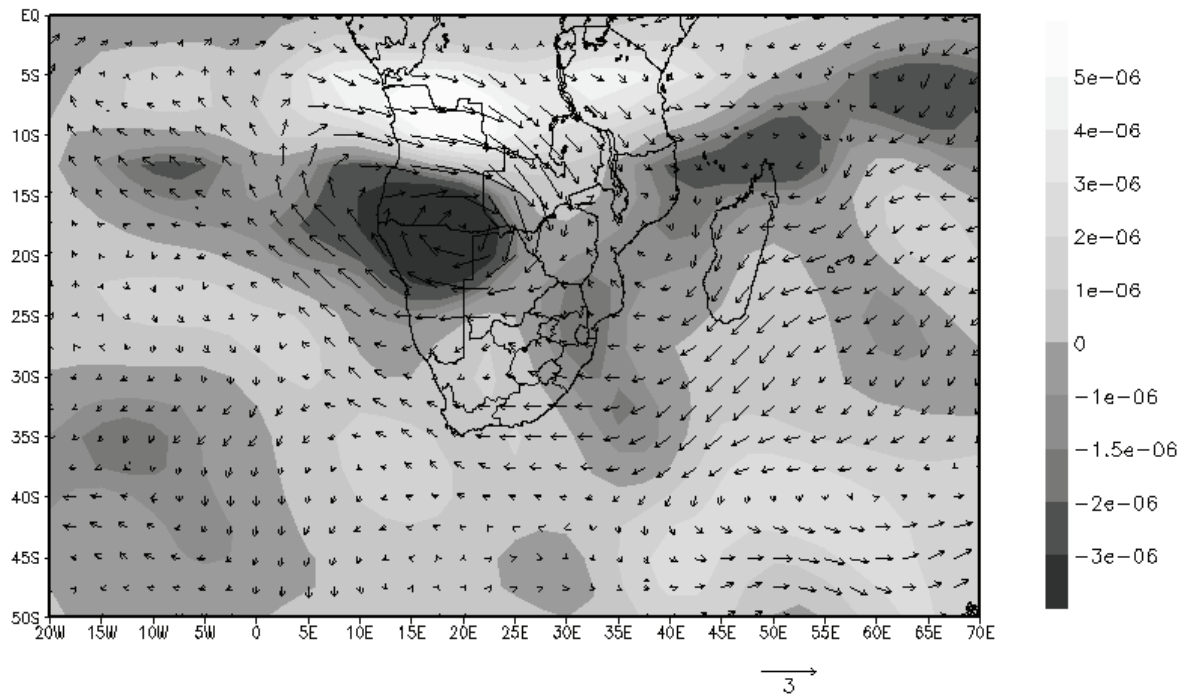


Fig. 9.13: Difference in 850 hPa relative vorticity (shaded) and difference in wind vectors for the period 1993-2001 with the 1979-1992 and 2002-2011 periods

The strengthened negative vorticity over the northern half of the subtropics is an indication of increased dynamical potential for the development of tropical systems over these latitudes over the southern African subcontinent and over the SWIO to the east. The anticyclonic anomaly to the southeast of the subcontinent has been linked to steering of tropical systems towards the west over the subcontinent (Ch. 8 & 9).

On a hemispheric scale, the anomalies during the 1993-2001 period display a shift towards a positive Southern Annular Mode (SAM) compared to the rest of the 1979-2012 period (shown for the 500 hPa level – Fig. 9.14).

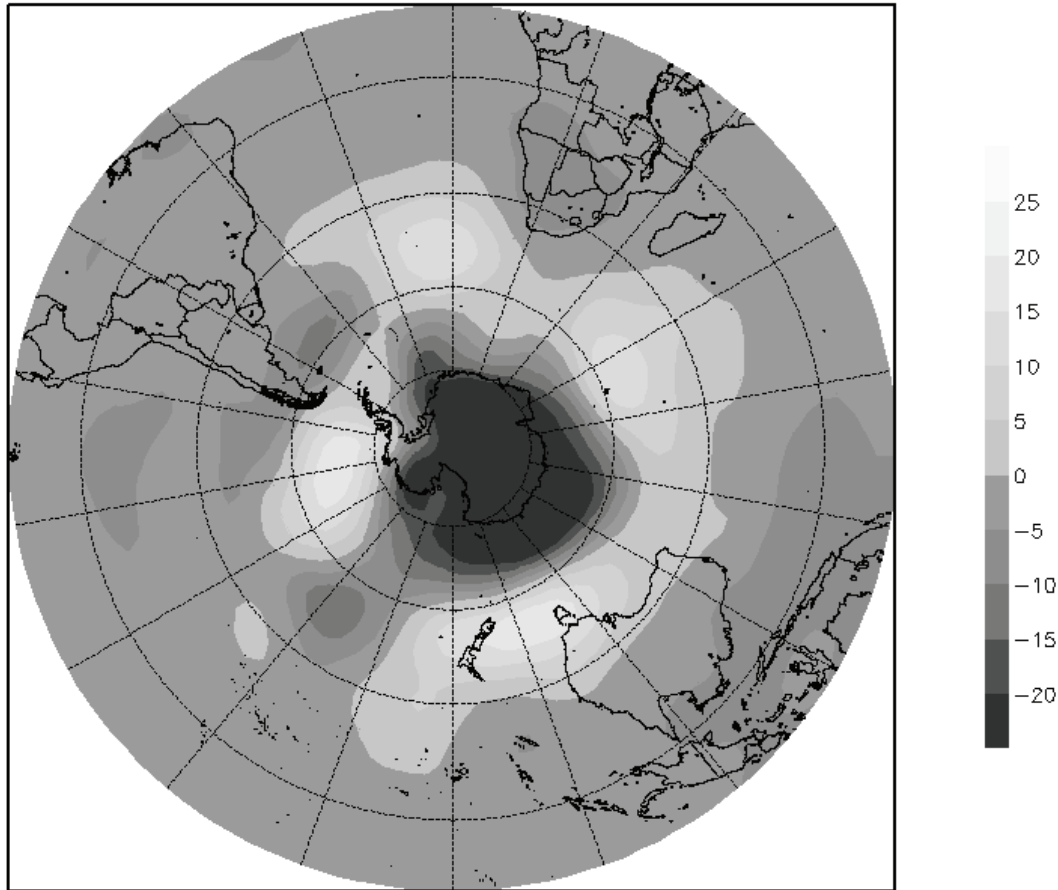


Fig. 9.14: 500 hPa height difference (m) for the period 1993-2001 with the 1979-1992 and 2002-2012 periods

In terms of hemispheric circulation anomalies, the anticyclonic circulation towards the south and southeast of southern Africa is part of a general tendency towards positive anomalies over the mid-latitudes while negative anomalies are indicated over the high latitudes. This is a positive deviation in the SAM. To conclude, we show the variation in contribution of tropical systems and also total January-March rainfall together with the SAM is parallel, indicating a positive correlation. We approximate this relation between mid and high latitudes by taking the difference in the weighted average occurrence of positive versus negative anomalies between 50° to 30°S and 90° to 65°S , detrended. Fig. 9.15 shows the variation in the moving 9-year moving average SAM calculated as mentioned above. Also indicated is the rainfall for the South African stations in Fig. 3.1 (January-March rainfall totals per year), normalized and categorized (into 8 categories), together with the 9-year moving average of this total. Additionally, solid circles indicate years with the largest seven events of rainfall contributed by tropical systems from the SWIO over this area (2 of these happened in 2000).

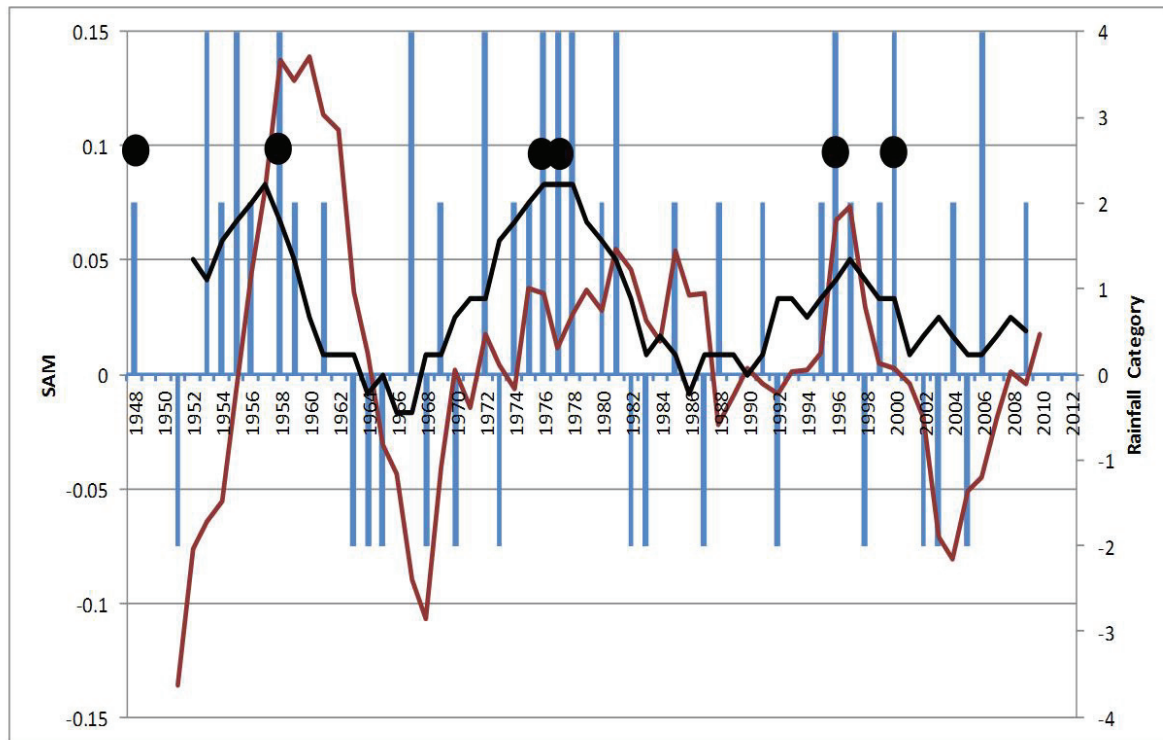


Fig. 9.15: Categorized January-March rainfall (bars), 9-year moving average of SAM index (brown line), 9-year moving average of rainfall category (black line) and years with largest rainfall events associated with tropical systems from the SWIO (circles)

The positive relation of the decadal scale variability over the Limpopo River Basin (in terms of January-March rainfall and the influence of tropical systems from the SWIO) to the latitudinal distribution of height anomalies in the mid and high latitudes (shown here by a rather broad approximation of the SAM) during the full study period (1948-2012) is indicated by the time series.

9.4 Conclusion

The period from 1993 to 2001 displays several tropospheric anomalies that increased the rainfall contribution of tropical systems from the SWIO over the Limpopo River Basin and that were also conducive to increased rainfall over southern Africa. The main large-scale feature is a positive deviation in the SAM (associated with positive anomalies in geopotential heights towards the south (mid-latitudes) and surrounding South Africa) as well as a broadening of the Hadley cell over the subtropics related earlier to summers with higher tropical cyclone activity in the SWIO region (Jury, 1993) and having the following consequences related to tropical systems from the SWIO:

- Enhanced dynamic potential for the development of tropical systems over the SWIO as well as over the southern African subcontinent.
- Enhanced easterly steering flow directing tropical systems towards southern Africa and also responsible for further inland penetration – also related to higher landfall risk (Vitart et al., 2003).

- Increased moisture content of the lower to middle troposphere over southern parts of southern Africa due to low-level convergence over the region and an increase in subtropical easterlies – increasing the favourability of the troposphere for sustaining tropical systems moving into the subcontinent.

This period also coincided with the expected rainfall maximum in the 18-year Dyer-Tyson cycle (Ch. 5). The regular return of higher rainfall since 1948 over the Limpopo River Basin, coupled with an increase in the influence of tropical systems from the SWIO (Malherbe et al., 2012), linked to variation on hemispheric scale over the mid to high latitudes, can contribute towards identifying the drivers of decadal-scale variability in the region as the SAM is sensitive to external forcing. It can further also explain the variation in rainfall over northeastern South Africa as related to phenomena on a similar time scale over other areas in the southern hemisphere, such as glacial advances over southern New Zealand (Tyson et al., 1997).

Chapter 10

Projections of Change in the Climatology of Tropical Systems over the SWIO and into Southern Africa under Enhanced Anthropogenic Forcing

10.1 Introduction

The rainfall contributed by landfall of tropical cyclones over southern Africa from the SWIO has been responsible for serious flooding events over the eastern parts of the southern African interior, including the Limpopo River Basin (Reason and Keibel, 2004; Crimp and Mason, 1999), a semi-arid region including parts of Mozambique, Zimbabwe, South Africa and Botswana. While on average about three tropical cyclones make landfall each year over Mozambique and/or Madagascar (Mavume et al., 2009), a total of only 45 tropical systems (including tropical cyclones) caused rainfall over the Limpopo River Basin after making landfall over the southern African subcontinent since 1948 (Malherbe et al., 2012).

The analysis of the future attributes of tropical systems, as projected by atmospheric global circulation models (AGCMs) or by coupled global circulation models (CGCMs), has been performed either by the tracking of simulated TCLVs (a vortex revealing tropical-cyclone-like characteristics, not exactly a tropical cyclone due to model-resolution constraints) or by inferring the general favourability of simulated large-scale conditions from indices based on atmospheric and surface variables relevant to tropical cyclone formation and intensity in indices considering.

CGCMs (e.g. Bengtsson et al., 1995; Tsutsui, 2002; Bengtsson et al., 2007) even when applied at spatial resolutions too coarse to represent realistically the intensity of tropical cyclone vortices, nested or nudged regional climate models (RCMs) (Walsh and Katzfey, 2000; Nguyen and Walsh, 2001; Walsh et al., 2004) and AGCMs forced by the SST fields of CGCMs (Sugi et al., 2009) have also been used to study the effect of enhanced anthropogenic forcing on tropical cyclone occurrence. The skill demonstrated by AGCMs of various resolutions in simulating the present day tropical cyclone climatology has, however, been reported to be low over the southern Indian Ocean, with an equator-ward displacement of tropical cyclone tracks over the region (Camargo et al., 2007; Sugi et al., 2009). The topography of the island of Madagascar furthermore influences the low level flow and hence the tropical vortices' evolution over the SWIO (Landman and Goddard, 2005). This necessitates, in the case of limited-area RCMs, the inclusion of the island in the domain of the regional model. Recent years have seen the advent of variable resolution AGCMs as an alternative to limited-area models, for the purpose of downscaling the output of CGCMs (McGregor, 2005). The present study is based on the output of the variable resolution AGCM CCAM with forcing from six AR4 GCM SST and Sea Ice fields, as described in Ch. 3. Assessment Report Four (AR4) of the IPCC points out that only a small number of studies have to date explored the potential impact of enhanced anthropogenic forcing on the attributes of tropical cyclones over the SWIO (Christensen et al., 2007).

Apart from the explicit tracking of TCLVs in climate models, the influence of large-scale environmental conditions such as SSTs, vertical wind shear, steering flow, vorticity as well as tropospheric temperature and humidity profiles on the occurrence of tropical cyclones (e.g. Gray, 1988; Holland, 1997) has facilitated the development of several tropical cyclone indices from seasonal averages of atmospheric and oceanic variables (e.g. Gray, 1998; Emanuel and Nolan, 2004; Zeng et al., 2007) employed on the output of simulated climate model datasets on a seasonal average basis to ascertain varying favourability for tropical

cyclone development over space and time. The Tropical Cyclone Seasonal Genesis Parameter (TCSGP; Gray, 1998) is one such index. This index focuses on the dynamical and thermodynamical properties of the environment. Dynamical variables include the Coriolis parameter, low-level absolute vorticity and tropospheric vertical wind shear, while the thermal variables include two atmospheric components, namely relative humidity in the mid-troposphere and difference in equivalent potential temperature between the surface and 500 hPa levels as well as an ocean component where thermal energy (temperature exceeding 26° to a depth of 60 m) is considered. A variation of this index is the Convective Genesis Parameter, where the thermal component of this index is replaced by the convective potential, which is a scaled representation of the mean convective rainfall (mm/day) for the relevant period (Royer et al., 1998; Caron and Jones, 2008). The alteration to the TCSGP index is based on the assumption that the thermal potential within the TCSGP is a representation of the potential for cumulonimbus convection, which is related to the supply of latent heat in the atmosphere (Royer et al., 1998) and which is the main energy source of tropical cyclones.

Several of the input atmospheric variables have also been used individually in studies into the temporal variability in the occurrence of tropical cyclones (e.g. Bell and Chelliah, 2006; Bengtsson et al., 2007; Chan and Liu, 2004; Martin and Chan, 2004; Sall et al., 2006; Singh et al., 2001; Vitart et al., 1999). Dynamic mechanisms over the North Pacific have been found to be key in driving changes in that region's tropical cyclone climatology (Chan and Liu, 2004), while an enhancement specifically of vertical wind shear between the lower and upper levels of the troposphere has been related to a projected decrease in Atlantic Ocean hurricane activity (Garner et al., 2009). A thermal property of the lower to middle troposphere, namely an increase in static stability, has been linked to a lower projected frequency of tropical cyclones over the northern hemisphere (e.g. Bengtsson et al., 2007).

Large-scale circulation patterns over the SWIO that have been linked to deviations in the occurrence of tropical cyclones include upper and lower tropospheric wind anomalies over the subtropics and tropics, and in particular the quasi-biennial oscillation (Jury, 1993). The Madden Julian Oscillation and ENSO have also been linked to the track and frequency characteristics of tropical cyclones over the region (Liebmann et al., 1994) while the landfall of tropical cyclones has been related to ENSO through an associated increase in the tropical easterlies over the SWIO during La Niña events (Vitart et al., 2003). An increase in low-level vorticity over parts of the SWIO has been linked to reduced tropical cyclone genesis over the SWIO during El Niño years (Vitart et al., 1999).

Anthropogenically enhanced warming could lead to a worldwide decrease in the frequency of tropical cyclones towards the end of the 21st century (e.g. Bengtsson et al., 2007; Yoshimura and Sugi, 2005). However, an increase is expected in the occurrence of the most intense systems (Bengtsson et al., 2007; Walsh and Ryan, 2000). A decrease in expected frequency of tropical cyclones under enhanced anthropogenic forcing has also been reported for the southern Indian Ocean specifically (Caron and Jones, 2008) as well as a slight increase in the maximum wind speed obtainable (Oouchi et al., 2006; Yu et al., 2010). Uncertainty in the projections is reflected to some extent in the finding of Tsutsui (2002) reporting an expected increase in the number of days with tropical cyclones over the region.

Findings based on the analyses of historical datasets over the SWIO include a lack of any trend in the observed number of tropical cyclones between 1960 and the late 1990s (Henderson-Sellers et al., 1998), a downward trend for the period 1952-2007 but with an upward trend for the latter part (1980-2007) of this period (Mavume et al., 2009), and an upward trend in the intensity of observed tropical cyclones over the region since the late 1970s into the 21st century (Webster et al., 2005; Elsner et al., 2008). Concerning landfall in

the region of the SWIO, a decrease has been observed in the landfall of tropical cyclones (Muvame et al., 2009) but no trend has been found either in landfall occurrences of all tropical warm-cored closed lows over southern Africa from the SWIO, or in the rainfall contributed by these systems (Malherbe et al., 2012).

This study is an evaluation of the simulated current and projected future climatologies of tropical systems over the SWIO under enhanced anthropogenic forcing by a variable resolution atmospheric model with a focus also on the inland tracks as these systems have been shown to play an important role in significant rainfall over the eastern parts of the southern African subcontinent, specifically the Limpopo River Basin.

10.2 Methodology

The tracking of warm-cored closed lows and TCLVs (a subset of the wider warm-cored closed lows based on minimum wind speed criteria) is described in Ch. 4. The area of interest, to allow for sufficient tracking over a large part of the SWIO and including the areas to the east of Madagascar, is 20°E to 80°E and 5°S to 40°S (Fig. 3.3). Results are presented in terms of the number of times that the simulated tracks of warm-cored low pressure systems/TCLVs migrate through each grid point per relevant 30-year period (1961-1990 or 2071-2100) as per CCAM ensemble member of ensemble mean. Comparison with the observed is done through the application of the tracking algorithm to the NCEP Reanalysis dataset and calculating the pattern correlation between these simulated and observed values, as well as the root mean square error and standard deviation (Taylor, 2001), interpreted as a measure of the CCAM ensemble's ability to represent the current climatology of tropical systems over the SWIO (including those moving into the subcontinent).

The Convective Tropical Cyclone Genesis Parameter (CTCGP; Royer et al., 1998) is calculated for the January to March period for an understanding of projected changes as inferred by tracking of warm-cored closed lows and TCLVs. The CTCGP is calculated from the averages of the relevant simulated fields. The multiplication factor used with the simulated convective daily rainfall is 0.1379, which is a value derived from the ERA40 dataset for the approximation of the average observed number of tropical cyclones globally (Caron and Jones, 2008).

10.3 Results and Discussion

10.3.1 Simulation of warm-cored closed low pressure systems and TCLVs compared to observed data

Fig. 10.1 shows the average number of all landfalling warm-cored closed lows over the region of interest for the average of the six CCAM simulations for the period 1961-1990 (top) and for NCEP over the same period (bottom). The landfall region along the eastern parts of southern Africa as well as the westward penetration into the subcontinent is captured well by the CCAM ensemble while other spatial characteristics are also captured. Over the interior of southern Africa, however, it can also be seen that there has been a larger tendency for closed warm-cored low pressure systems from the SWIO to move a bit further south (as observed from the NCEP Reanalysis dataset) than found in the CCAM simulations.

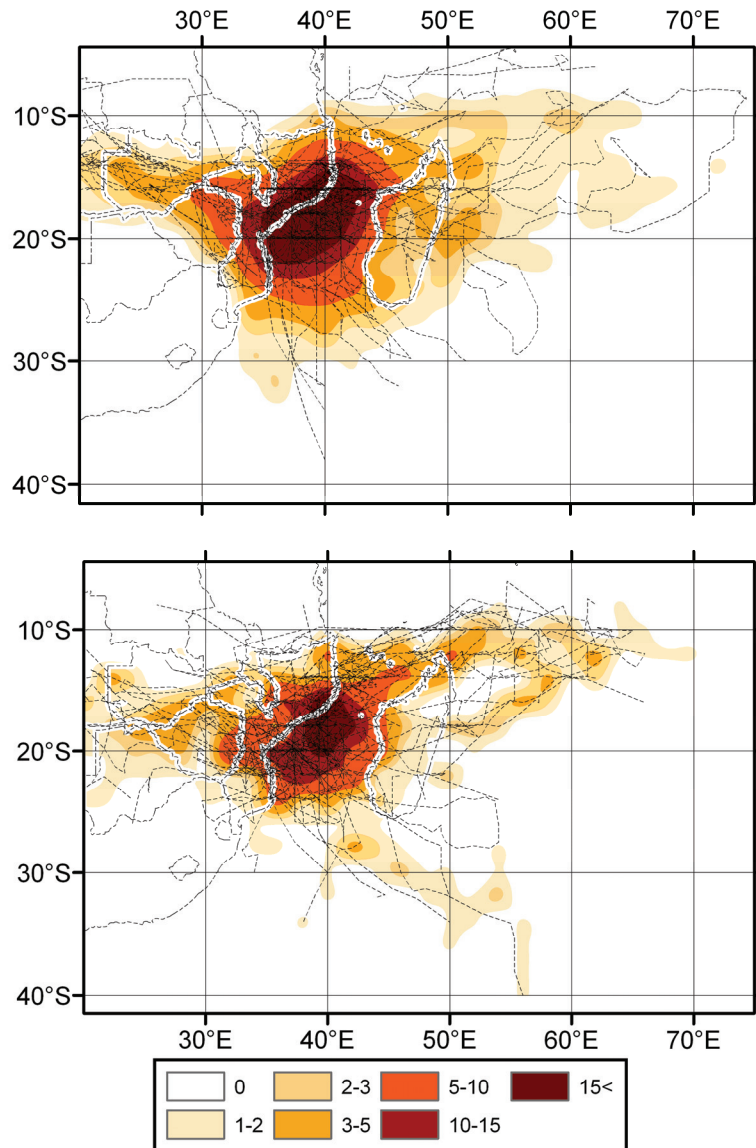


Fig. 10.1: Tracks per grid point (shaded) of warm-cored closed lows originating over the SWIO and making landfall into southern Africa as simulated by the CCAM ensemble for 1961-1990 (top) and observed tracks per grid point according to NCEP Reanalysis data (bottom). Actual tracks are also indicated (for the CCAM-CSIROMk3.5 on the CCAM-Ensemble map)

Fig. 10.2 focuses on how the passage of warm-cored closed low pressure systems from the SWIO during the period 1961-1990 as observed in NCEP reanalysis data and as simulated by the CCAM ensemble was captured along 28°E (longitude crossing the eastern parts of Botswana).

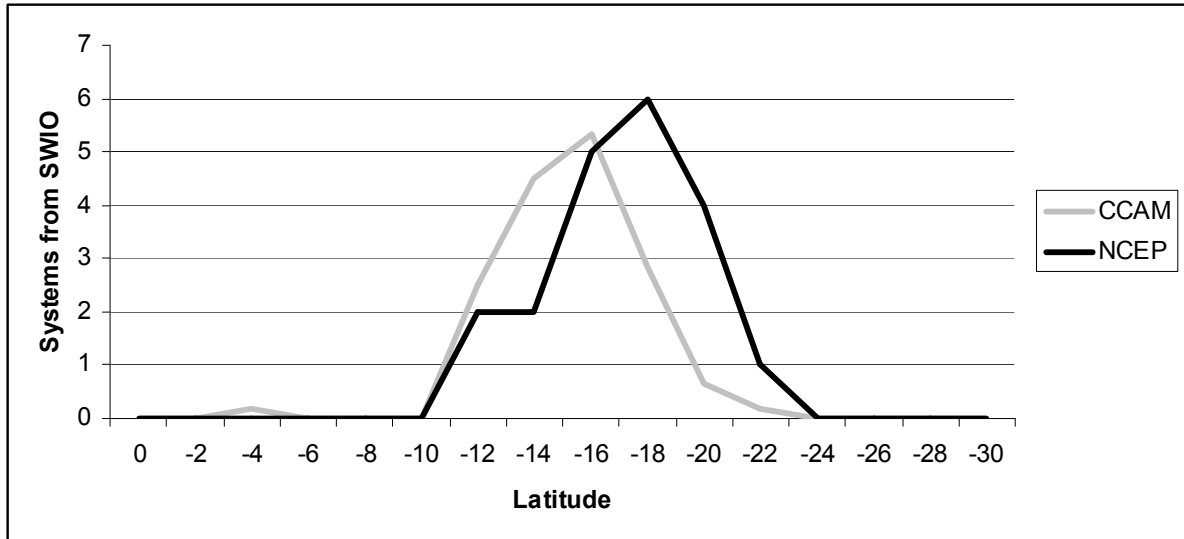


Fig. 10.2: Latitudinal occurrence of warm-cored closed low pressure systems from the SWIO moving across 28°E as simulated by CCAM ensemble (grey) and observed in NCEP (black) for the period 1961-1990

The number of systems moving into the subcontinent simulated by the CCAM ensemble compares well with the observed. As mentioned earlier, the tracks simulated by the CCAM ensemble are positioned somewhat north (about 2°) to the observed with westward movement into the subcontinent.

Fig. 10.3 shows for the period 1961-1990 the simulated CCAM ensemble average of the frequency of occurrence of simulated TCLVs per grid point (top, shaded) and as observed from the NCEP Reanalysis data (bottom). The individual TCLV tracks, from which the frequency of occurrence per grid point is calculated, are also shown for the observed tracks (bottom) and for one of the CCAM ensemble members (top).

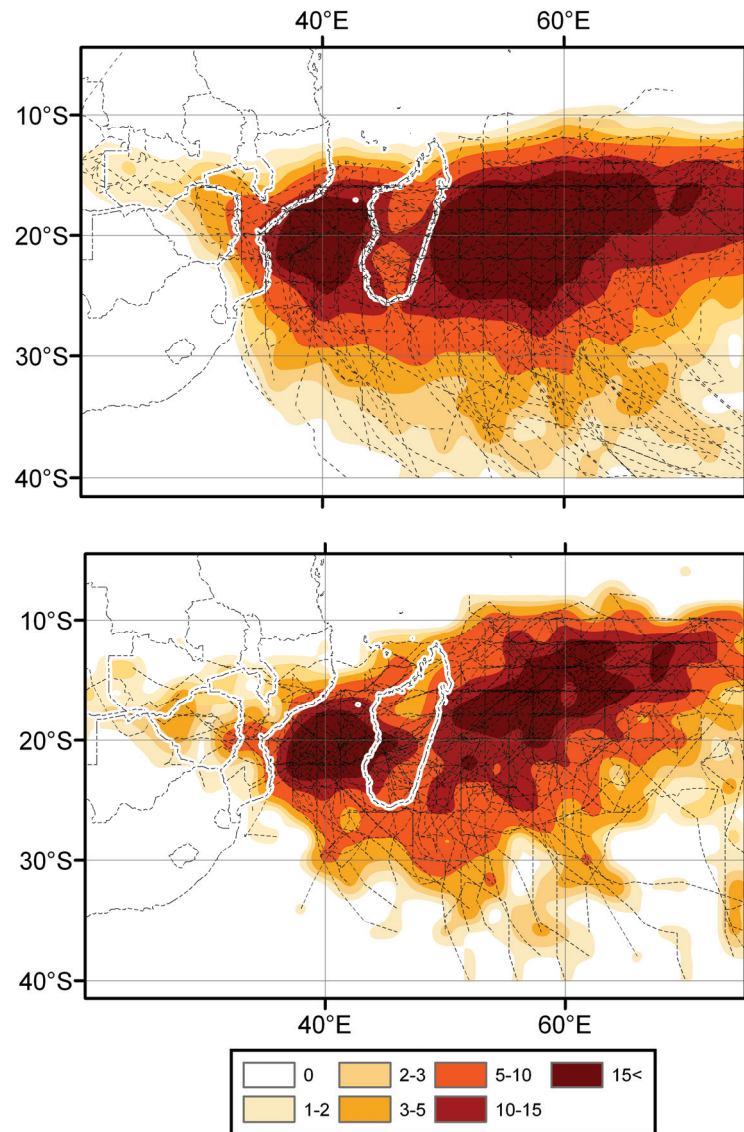


Fig. 10.3: Tracks per grid point (shaded) of TCLVs originating or moving across the SWIO as simulated by the CCAM ensemble for 1961-1990 (top) and observed tracks per grid point according to NCEP Reanalysis data (bottom). Actual tracks are also indicated (for the CCAM-CSIROmk3.5 on the CCAM-Ensemble map)

Fig. 10.4 shows the overlap over the SWIO of observed versus simulated TCLVs for the period 1961-1990. The slight northward displacement over the southern African interior can also be noticed from this map.

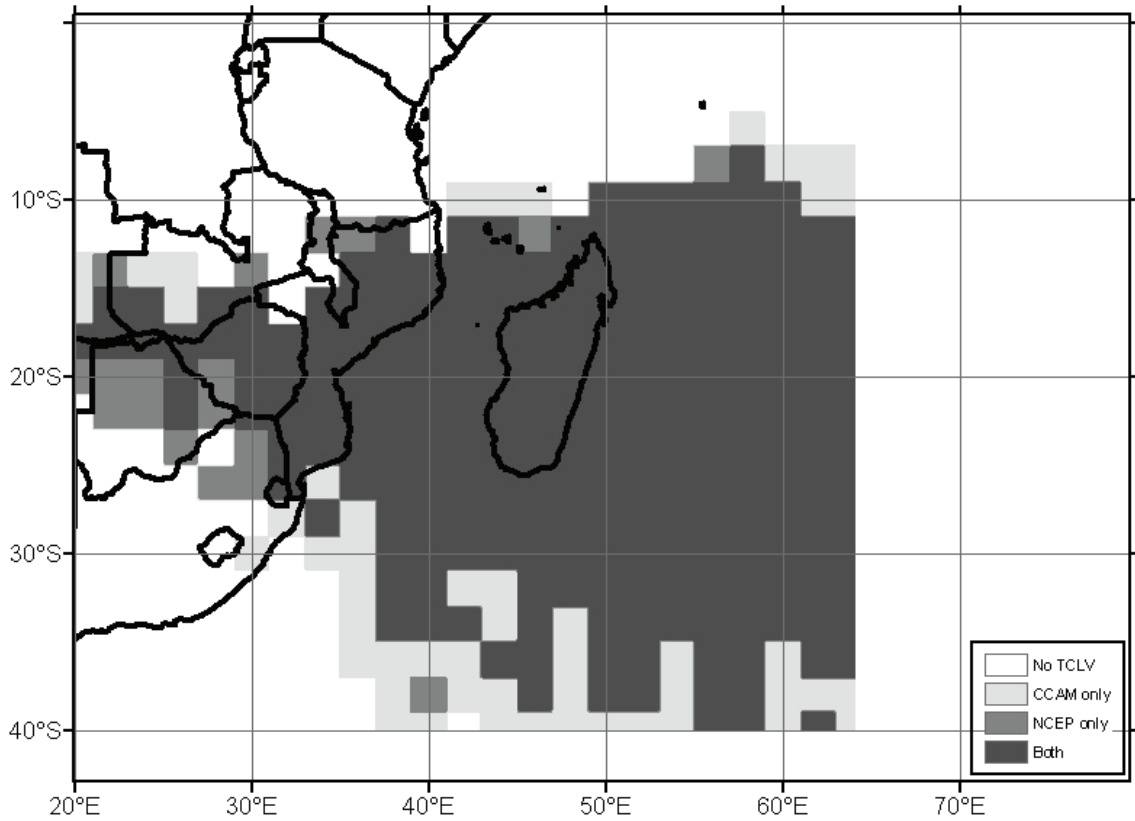


Fig. 10.4: Distribution of areas where TCLVs were only simulated by CCAM (light grey), observed in NCEP Reanalysis data and not simulated in the CCAM ensemble (grey) and where TCLVs were both observed in NCEP Reanalysis data and simulated by the CCAM ensemble

The overall agreement between the two areas (simulated versus observed) is high according to Fig. 10.4. For each $2 \times 2^\circ$ grid point in the region of interest, Fig. 10.5 shows the correlation between the fractions of the total number of TCLVs per dataset that moved across that point as observed in the NCEP Reanalysis data and the CCAM ensemble.

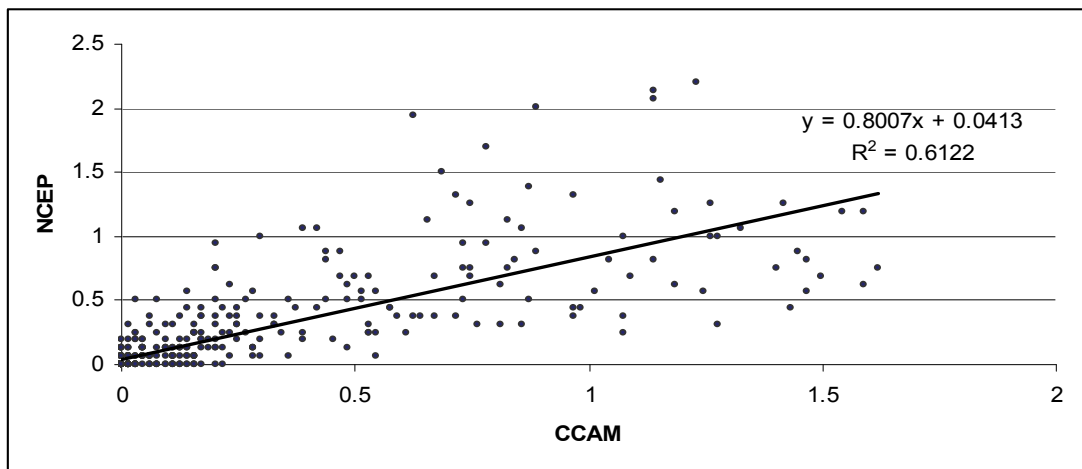


Fig. 10.5: Correlation between the CCAM ensemble simulation (x-axis) and observed NCEP Reanalysis data (y-axis) of TCLVs in the region of interest

The positive correlation indicated when the fractions of the total number of tracks per point of the observed are plotted against the simulated is also seen in the correlation matrix between the CCAM ensemble and NCEP datasets with a Pearson correlation coefficient of 0.78 between the two datasets.

Table 10.1: Correlation matrix of CCAM ensemble and NCEP Reanalysis occurrence of tracks per grid point

	<i>CCAM</i>	<i>NCEP</i>
<i>CCAM</i>	1	
<i>NCEP</i>	0.78242	1

The correlation matrix demonstrates a high agreement of the ensemble mean with the observed tracks of TCLVs. The pattern correlation of the individual members and mean ranges between 0.76 and 0.82 (Taylor diagram – Fig. 10.6) – the highest being with the ensemble mean, having also the smallest root mean square error, suggesting that the ensemble mean is an improvement to the simulations of the individual members.

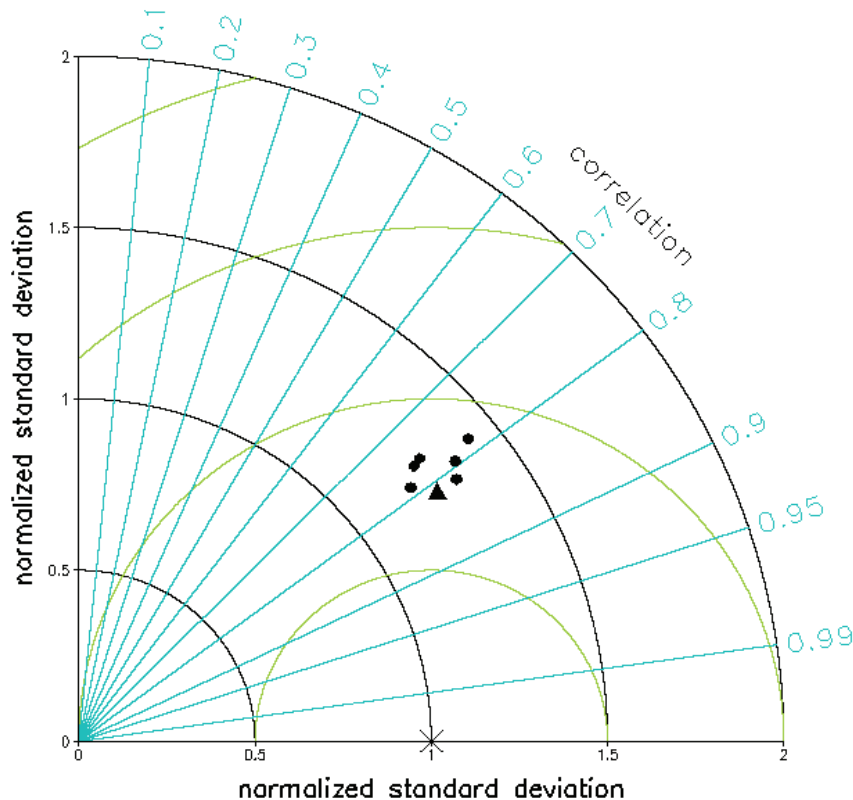


Fig. 10.6: Taylor diagram showing the statistical comparison of the simulation of TCLVs frequency of occurrence per grid point of the individual CCAM ensemble members (circles) and the CCAM ensemble mean (triangle) over the SWIO with the reanalysis distribution (cross)

10.3.2 Projected changes as simulated by the CCAM ensemble

The simulated change in warm-cored closed lows as well as TCLVs is shown in Fig. 10.7 as an ensemble average as well as per difference number of ensemble members projecting a positive and negative change.

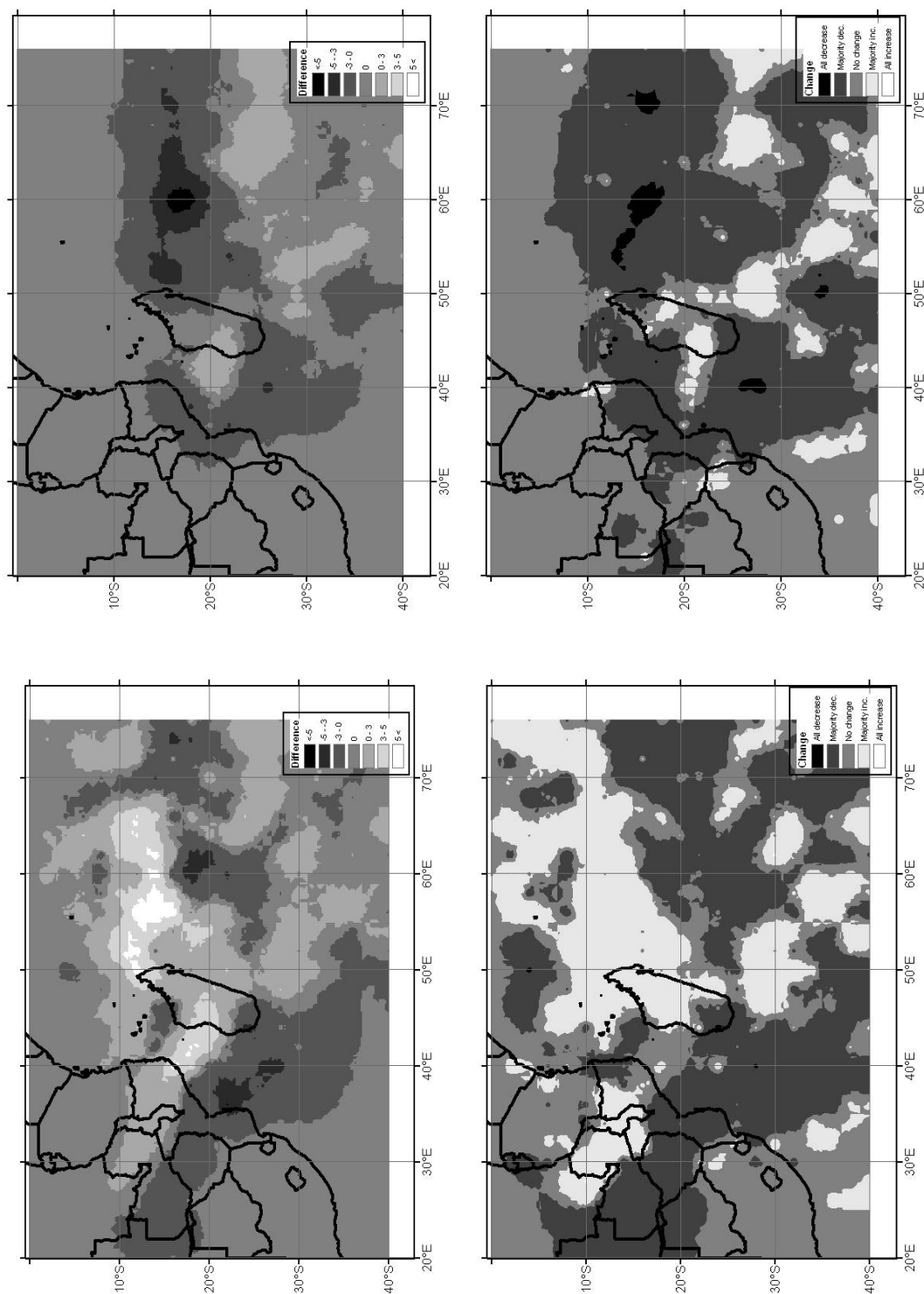


Fig. 10.7: Ensemble mean difference (top) and number of ensemble members indicating an increase/decrease (bottom) in number of tracks of warm-cored closed lows (left) and TCLVs (right) originating or moving across the SWIO between 1961-1990 and 2071-2100

Fig. 10.7 shows that the CCAM ensemble simulates a decrease in the number of TCLVs over much of the region, and a decrease of all warm-cored closed low pressure systems over the southern parts of the Mozambique Channel into the Limpopo River Basin and the southern parts of the Indian Ocean in general while an increase in warm-cored closed lows, but not of the intensity of TCLVs, is simulated over the areas surrounding northern Madagascar into the northern parts of Mozambique. The change in average rainfall and 700 hPa heights for the January-March period simulated as calculated from the 30-year averages for the 1961-1990 and 2070-2100 periods, however, provides insight into this northward displacement (Fig. 10.8).

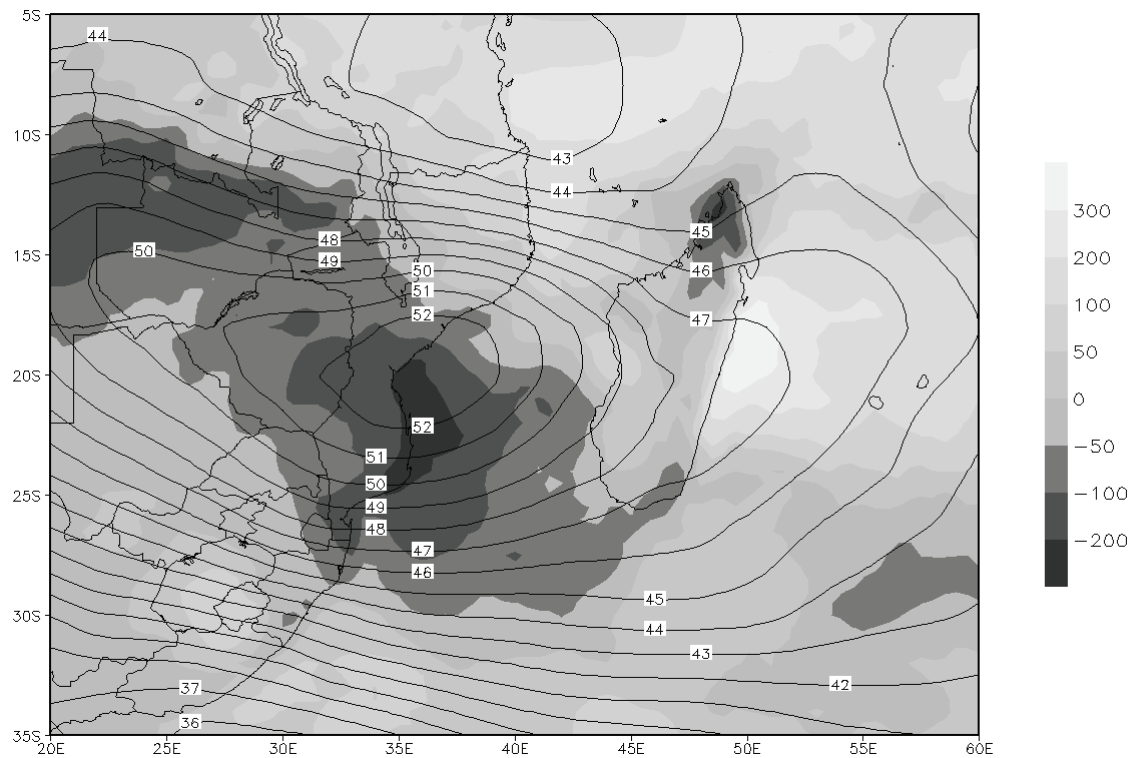


Fig. 10.8: CCAM ensemble average projected change in January-March rainfall (shaded) and 700 hPa heights (contours) between the 1961-1990 and the 2071-2100 periods

Relatively larger increases of the 700 hPa heights are simulated over the eastern parts of the subcontinent stretching northwestward and also into the southern parts of the Mozambique Channel, indicating a relatively larger strengthening of the subtropical high pressure over that area, related to lower rainfall over the eastern parts of the subcontinent. The strengthening of the high can also be related to a northward displacement of warm-cored closed lows moving westward into the subcontinent around the northern edges of the strengthened mid-level high pressure area. Furthermore, the decrease in TCLVs specifically over the southern parts of the Mozambique Channel and to the east of Madagascar can be explained by projected changes in large-scale thermal and dynamic features of the atmosphere. The change in the CTCGP gives an indication that the dynamic and/or thermal properties are projected to become more favourable for TC genesis over the central to northern parts of the SWIO, but less favourable towards the southern parts, specifically towards the southern half of the Mozambique Channel and also east of 60°E, south of 15°S (Fig. 10.9).

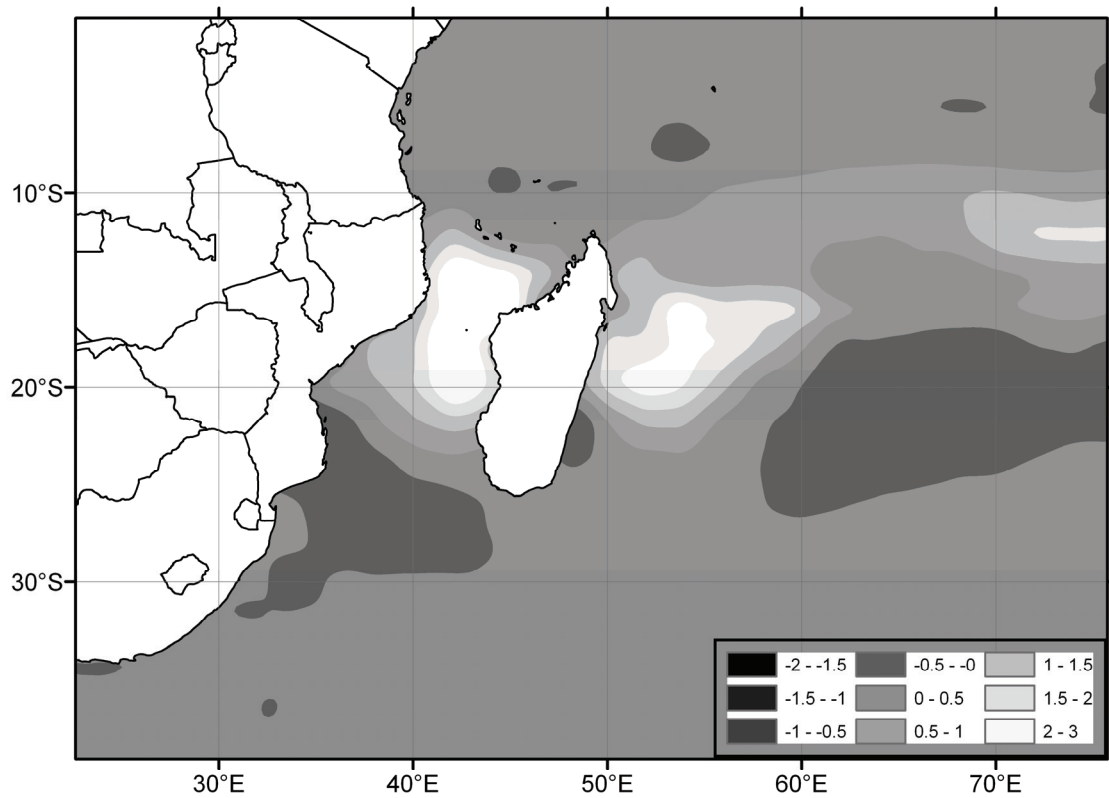


Fig. 10.9: Median change in the Convective Tropical Cyclone Genesis Parameter (unit: approximate average 10-yearly total number of tropical cyclones per 2° grid point) as calculated from the ensemble of CCAM members for the period 2071-2100 relative to 1961-1990

The change in the components of the CTCGP suggests that dynamic and thermal properties of the troposphere are both responsible for changes projected over several areas (Fig. 10.10).

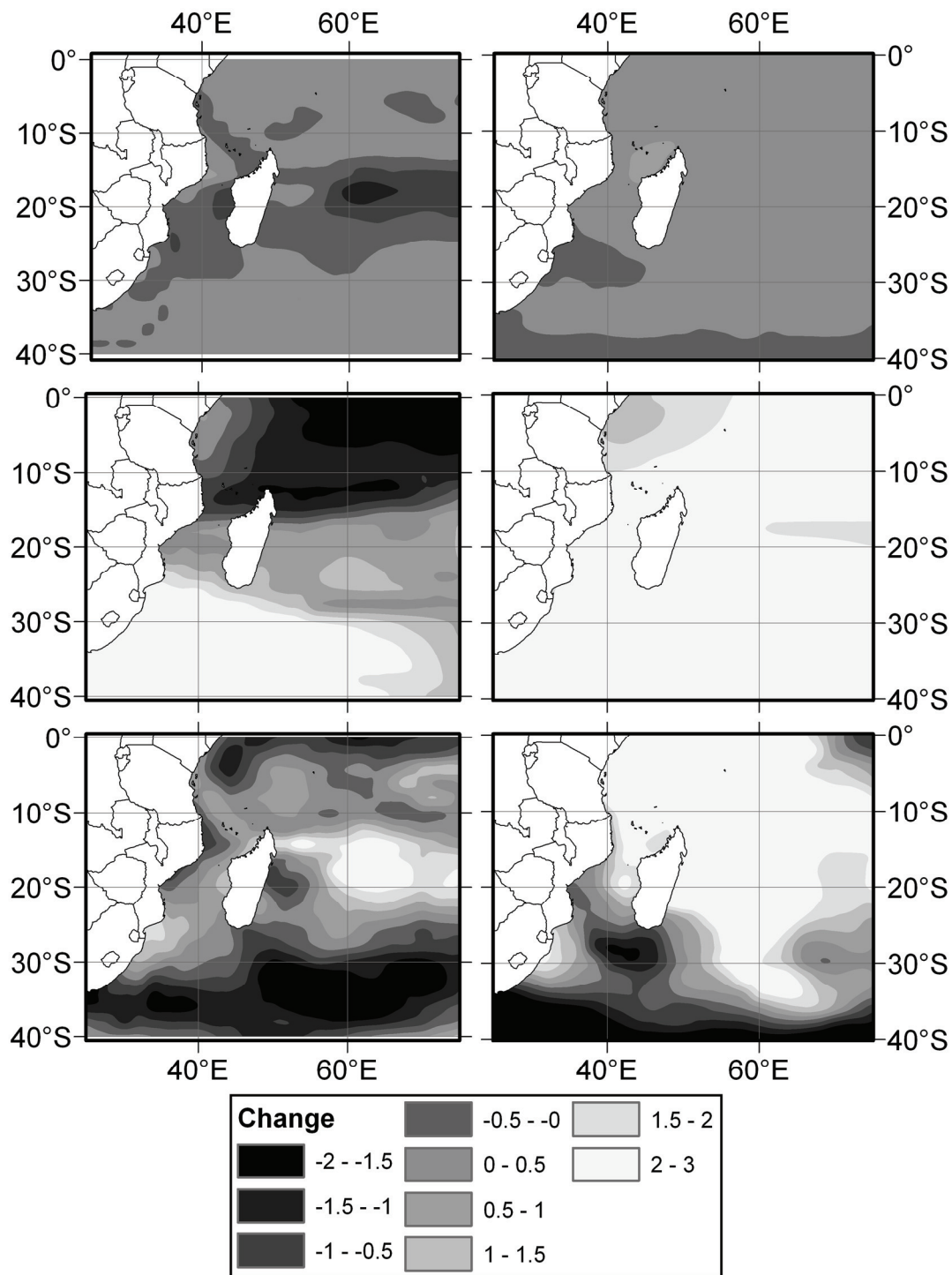


Fig. 10.10: The change in the dynamic (top left) and convective (top right) components of the Convective Tropical Cyclone Genesis Parameter, change in the average wind shear ($\text{ms}^{-1} / 600 \text{ hPa}$) between 800 and 200 hPa (middle left) and the change in vorticity (10^{-6} s^{-1} – bottom left) dynamical components and change in the gradient of potential temperature ($10^{-1} \text{ }^{\circ}\text{C}$) between the surface and 500 hPa (middle right) and the change in relative humidity (%) at 600 hPa (bottom right) as calculated from the ensemble of CCAM members between the 1961-1990 and 2071-2100 periods

From Fig. 10.10 it is suggested that over the areas adjacent to southern Africa, the projected decrease in TCLVs simulated is related to dynamical components (increasing wind shear and a decrease in large-scale vorticity) as well as thermal properties (increase in potential temperature gradient and decrease in relative humidity at 600 hPa) while the decrease to the east of 60°E is related more to the dynamical component of the atmosphere. Over the northern parts of the Mozambique Channel and around northern Madagascar, conditions become more favourable for the development of tropical cyclones according to the projected changes in the dynamical and thermal properties of the atmosphere and this is at least reflected in the simulated increase in warm-cored closed lows over that region, as well as (over a more limited area) the increase in simulated TCLVs.

10.4 Conclusions

The six-member CCAM ensemble has captured the general spatial characteristics of the occurrence of warm-cored closed lows and TCLVs over the SWIO such as preferred tracks and landfall into southern Africa realistically for the 1961-1990 period.

According to the set of projections, based on the business as usual SRES scenario, a general decrease in the frequency of tropical cyclones may occur towards the end of the 21st century over most of the SWIO region, even though the total number of warm-cored closed lows over the region is projected to increase. While the increase in the total number of warm-cored closed lows in a zonal band over and to the north of northern Madagascar is associated with increasing favourability in thermal and dynamic features of the atmosphere, the projected decrease in TCLVs over the southern areas is related to a projected change mainly in the dynamic properties of the atmosphere and in both dynamic and thermal properties close to southern Africa. While mainly thermal properties are associated with a decrease in the expected occurrence of TCs over the SWIO to the east of Madagascar, the decrease over the areas adjacent to southern Africa is related to:

- an increase in static stability in the lower to middle troposphere
- a decrease in relative humidity and an enhancement of vertical wind shear between the lower and upper levels of the troposphere
- an increase in vorticity (decrease in cyclonic vorticity) in the lower levels over especially the regions of the SWIO close to the southeastern parts of southern Africa where the strengthening of the subtropical high pressure is expected to be larger than over the surrounding areas

The projected northward shift in the tracks of tropical systems from the SWIO over southern Africa and decrease in average January-March rainfall is also the result of a relatively larger strengthening in the subtropical high pressure system over the northern to eastern parts of southern Africa relative to the areas surrounding that. The decrease in the simulated future occurrence and intensity of tropical systems from the SWIO moving into the Limpopo River Basin under enhanced anthropogenic forcing doesn't bode well for the Limpopo River Basin where the widespread rainfall associated with these types of systems is essential in this semi-arid area with a large rural population dependent on agriculture.

Chapter 11

Projected Impact of Climate Change on Extreme Hydrological Events in Two Primary Drainage Basins in Limpopo

**TG Lumsden, RE Schulze, DM Knoesen
and RP Kunz**

*School of Bioresources Engineering and Environmental Hydrology
University of KwaZulu-Natal, Pietermaritzburg, South Africa*

PLEASE CITE THIS CHAPTER, IN CONTEXT OF THE FULL REPORT, AS:

Lumsden TG, Schulze RE, Knoesen DM and Kunz RP. 2011. Projected impact of climate change on extreme hydrological events in two primary drainage basins in Limpopo Province. School of Bioresources Engineering and Environmental Hydrology, University of KwaZulu-Natal, Pietermaritzburg, RSA, *ACRUcons Report* 64. 15 pp.

DISCLAIMER

While every reasonable effort has been made by the authors to obtain objective and realistic results in this study, neither the authors, the School of Bioresources Engineering and Environmental Hydrology nor the University of KwaZulu-Natal, nor any of their employees, make any warranty, express or implied, or assume any legal liability or responsibility for the accuracy, completeness or usefulness of any information, product or process disclosed by this report.

11.1 Introduction

Climate change is projected to induce significant changes to streamflow in South Africa (Schulze et al., 2005; Schulze, 2011). The potential impact on extreme hydrological events is of concern, particularly in regions which are subject to potentially destructive tropical weather systems, such as the Limpopo Province. The aim of this component of the project is therefore to investigate the potential impact of climate change on extreme hydrological events in a Limpopo catchment affected by tropical weather systems.

The methodology employed involves the application of a hydrological model, together with the regional climate projections developed in the project. The impact on extreme hydrological events is assessed by considering the impact on design floods, as estimated from the output of the hydrological model. The estimation of design floods does not, however, consider the nature of the rainfall causing the flood events. These events may therefore be caused by both tropical and non-tropical weather systems. Elsewhere in the project, an attempt is made to isolate and quantify the influence of changes in tropical weather systems on extreme hydrological events.

This chapter details the methodology and results of the study to assess climate change impacts on design floods, which was focussed on two primary drainage basins in the Limpopo Province.

11.2 Methodology

11.2.1 Study area

The study area in this component of the project is defined to be the Department of Water Affairs (DWA) Primary Drainage Basins A and B. These basins cover almost the entire Limpopo Province, as well as parts of adjacent provinces. The primary river draining Basin A is the Limpopo River while the primary river draining Basin B is the Olifants River. The basins correspond to the South African portion of the Limpopo River Basin, which also spans parts of Botswana, Zimbabwe and Mozambique. Basins A and B drain into Mozambique at separate locations and later converge some 200 km from the Mozambique coastline. The portions of the Limpopo River Basin in the above-mentioned neighbouring countries are excluded from the study area owing to a lack of readily available data. A map of the study area is given in Fig. 11.1.

11.2.2 Hydrological modelling and input data

The *ACRU* hydrological model (Fig. 11.2) was selected for application in hydrological modelling. Warburton et al. (2010) have shown that the *ACRU* model adequately simulates historical observed streamflow in three catchments in different climatic regions of South Africa, one of which was the Luvuvhu Catchment located in Primary Basin A within the study area. This implies that the model can be applied with confidence to simulate streamflow in the study area. In terms of process representation and structure, *ACRU* satisfies many of the requirements of a model for application in assessing the hydrological impacts of climate change, including, for example (Schulze, 2010):

- capturing the sensitivity of hydrological responses (e.g. runoff, evaporation, transpiration) to climate;

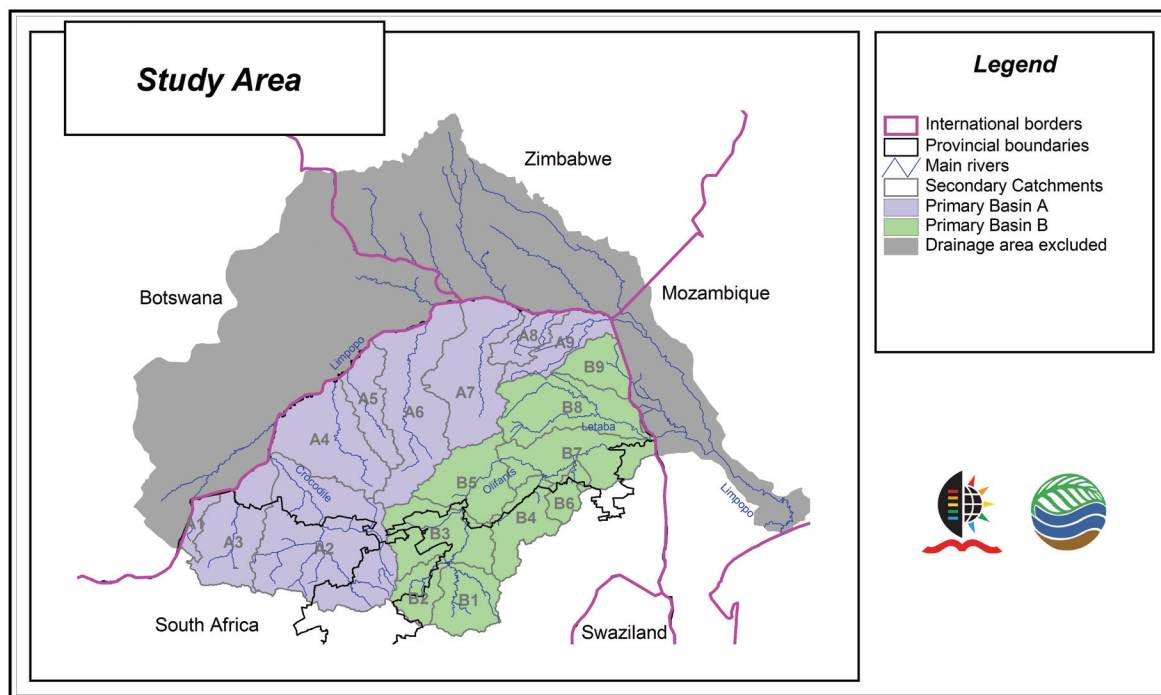


Fig. 11.1: Map of the study area for assessing projected climate change impacts on extreme hydrological events

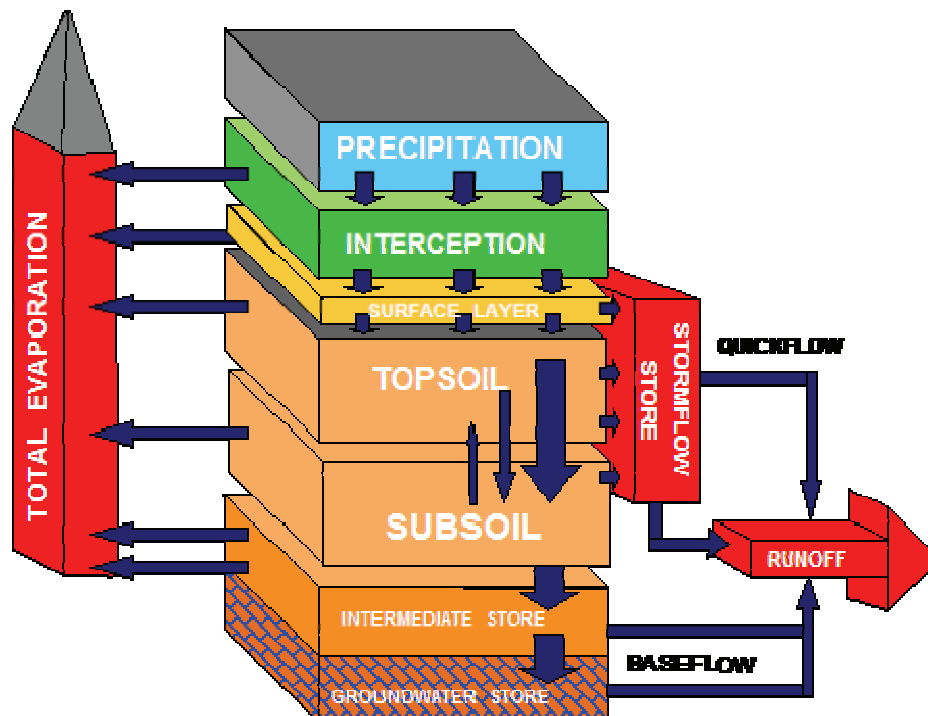


Fig. 11 2: The ACRU agrohdrological modelling system: general structure (after Schulze, 1995)

- operating at a time step appropriate (daily) for relevant processes likely to be affected by climate change (e.g. evaporation, transpiration);
- being conceptual-physical in nature, where processes are adequately represented and physical boundary conditions exist, thus enabling the model to be extrapolated to new contexts (e.g. climate change).

For operational decision making, South Africa has been delineated by DWA into 1946 hydrologically interlinked and cascading Quaternary Catchments. This represents a fourth level of subdivision of the landscape, with the previously mentioned Primary Catchments being the first level subdivision. It has been shown, however, that the Quaternary Catchments still exhibit considerable physiographic heterogeneity (Schulze, 2004) and that a further catchment subdivision was required to obtain more homogenous hydrological response units. Hence the Quaternary Catchments were further subdivided into Quinary Catchments using an altitude based methodology (Schulze and Horan, 2010). Each Quaternary Catchment was divided into three Quinary Catchments (upper, middle and lower) where the boundaries of these Quinaries were defined according to “natural breaks” in altitude. Altitude was used as the driver in the subdivision process as changes in altitude are in turn often associated with changes in hydrological drivers such as climate, soils and vegetation. The outflow of the Lower Quinary of a Quaternary Catchment was configured to enter the Lower Quinary of the next downstream Quaternary Catchment (rather than the Upper Quinary), since the latter may be at a higher altitude (Schulze and Horan, 2010).

In this project the *ACRU* model was applied in the study area at the scale of Quinary Catchments (QCs). In Primary Basin A there are 417 QCs while in Primary Basin B there are 435 QCs.

An extensive database of model inputs has been developed for the QCs and is known as the Quinary Catchments Database (Schulze et al., 2010). These inputs relate to, *inter alia*, daily historical observed climate (1950-1999), soils and land cover. The development of these inputs for application in the *ACRU* model is described in Schulze et al. (2010). These inputs were applied in *ACRU* for this study. Hydrological simulations were performed using both historical observed climate data (described in Schulze et al., 2010) and the regional climate projections developed in this project. Hydrological simulations were performed using historical observed climate data to provide a context within which the changes in hydrological extremes (derived from simulations using regional climate projections) could be viewed.

The regional climate projections developed in the project consisted of dynamically downscaled (via CCAM) global climate projections generated by six IPCC AR4 coupled climate models (for the A2 SRES emissions scenario), namely:

- CCAM-CSIROmk3.5
- CCAM-GFDLcm2.0
- CCAM-GFDLcm2.1
- CCAM-MIROC
- CCAM-ECHAM5
- CCAM-UKHADcm3

The regional projections were made available for the continuous period from 1961 to 2100 for daily rainfall and maximum and minimum temperatures.

The manner in which the regional climate projections were used to develop climate inputs to *ACRU* at QC scale is described below. Consideration was first given to the relative scales of the regional climate projections (generated on a 0.5 degree resolution grid) and the QCs, as

shown in Fig. 11.3. It is apparent from Fig. 11.3 that while there are a number of QCs within each grid cell, there are also numerous cases where a QC is overlaid by more than one grid cell. It was decided that the grid cell that had the largest degree of overlap with a QC by area would be the cell that represented climate in that QC. Thus each QC was associated with a unique representative grid cell. However, since the grid is relatively coarse in scale there will, in reality, be a fairly large spatial variation of rainfall within each cell.

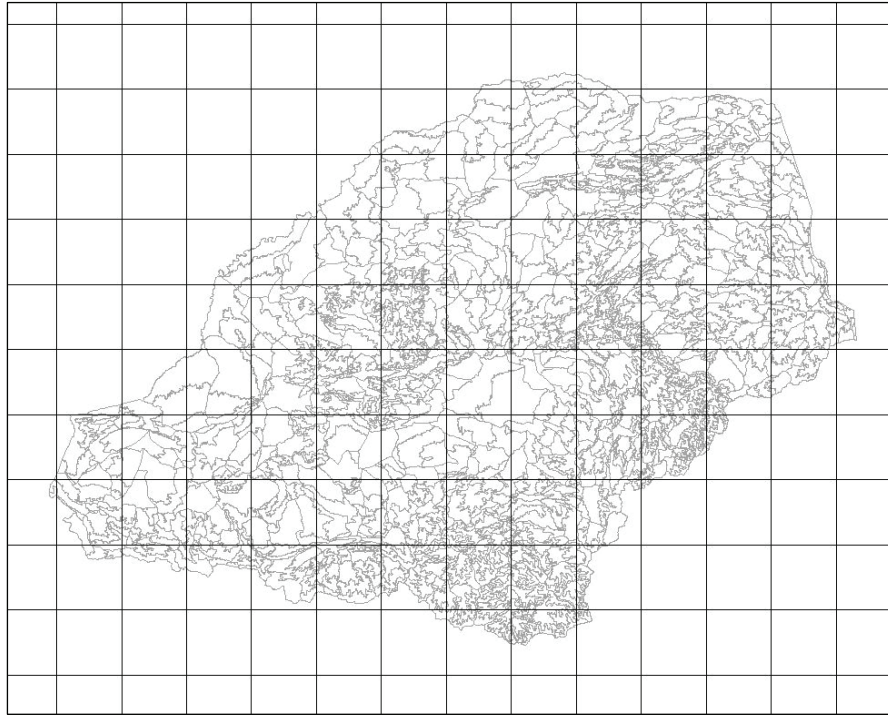


Fig. 11.3: Quinary Catchments in the study area overlaid by a 0.5° resolution grid representing the scale of the regional climate projections

Thus for each QC an adjustment factor was derived and applied to the rainfall dataset of the representative grid cell to develop a unique rainfall record for that QC. In effect this introduced a more realistic spatial variability in rainfall over the QCs represented by a particular grid cell. The adjustment factor for a QC, y , was calculated for each month ($j=1$ to 12) of the year as follows:

$$AF_j = QC_{rain} / QC_{rain_{ave}}$$

where:

- AF_j = adjustment factor for month j
- QC_{rain} = median monthly rainfall for month j of QC y (mm)
- $QC_{rain_{ave}}$ = area weighted average for month j of median monthly rainfalls of all QCs overlapping the representative grid cell for QC y (mm)

The median monthly rainfalls applied in the calculation of the adjustment factor were derived from the observed daily climate record available for each QC in the Quinary Catchments Database. $QC_{rain_{ave}}$ was calculated over the domain of the representative grid cell for QC y , i.e. the area weighting excluded portions of the component catchments that lay outside of the representative grid cell. The adjustment factors were restricted to lie within the range $0.5 < AF_j < 2.0$ to prevent extreme adjustments being made to the climate projection time series.

Since temperature is less spatially variable than rainfall, no adjustment was applied to the grid cell temperature data to represent temperatures in the individual QCs.

The land cover in the study area was assumed to be a baseline land cover, as represented by the Veld Types of Acocks (1988). Furthermore, no abstractions from the water resources of the area were considered, nor was any water resources infrastructure (e.g. dams, transfer pipelines, etc.) represented in the simulations. Thus, the hydrological simulations in this study corresponded to “natural” catchment conditions. This implies that the influence of climate on the streamflow simulations is not masked by the influence of land use.

11.2.3 Estimation of design floods

The estimation of design floods involves performing an extreme value analysis on the daily time series of simulated streamflow for each subcatchment. In this case the streamflow considered was the accumulated streamflow volume including contributions from all upstream subcatchments. This involves first extracting the Annual Maximum Series (AMS) from the daily streamflow time series. In the case of, for example, 2-day duration events, this series comprises of the highest two consecutive day duration flow values from each year of simulation. Hydrological years (running from 1 October to 30 September) were considered in this process. Thereafter an extreme value distribution is fitted to the AMS to allow for the estimation of flow values corresponding to different return periods. The Log-Pearson Type III (Kite, 1988) distribution was selected for this purpose as it is widely accepted and applied internationally. Design floods were determined for the 1, 2, 3, 4, 5, 6 and 7 day duration events. In each case the floods were estimated for the 2, 5, 10 and 20 year return periods.

11.2.4 Assessing climate change impacts on design floods

Design floods were first estimated based on the observed climate record for the period 1950-1999. These estimates were generated to provide a context within which the changes in hydrological extremes (derived from simulations using regional climate projections) could be viewed. The changes in hydrological extremes were assessed by considering a present and future period within the 140-year time series of climate produced by the climate models. These present and future periods were 1961-1990 and 2071-2100, respectively, as defined elsewhere in the project. Changes in floods were expressed in terms of a ratio where the future flood value was divided by the present value, such that a ratio greater than 1 indicated an increasing trend (blue shades) while a value less than 1 indicated a decreasing trend (red shades).

11.3 Results

11.3.1 Design floods based on observed climate data

A sample of the design flood estimates (based on observed climate) produced for the various event durations and return periods is presented as follows:

- First, the design floods for various return periods for the 7 day duration event are given in Fig. 11.4;
- Secondly, the design floods for the 20 year return period are given for various event durations in Fig. 11.5.

In Figs. 11.4 and 11.5 accumulated catchment streamflows are mapped (i.e. flows in a particular catchment include contributions from upstream catchments). The boundaries of secondary catchments within the primary catchments are indicated on the maps. It should be

noted that the flows into the Limpopo River from the various secondary catchments in Basin A are not accumulated since they are considered to be exit flows leaving South Africa. In addition the flow contributions from Botswana and Zimbabwe were not available for inclusion in the accumulation of flows.

In Fig. 11.4 the expected trend is evident, this being that streamflows increase as return period increases. The catchments containing major river channels become clearly visible in the dark blue shade assigned to the highest category of flow ($> 500 \times 10^6 \text{ m}^3$). The catchments containing the Limpopo River do not follow this trend; however, since flows are not accumulated in these catchments (see above paragraph for explanation).

In Fig. 11.5 the expected trend is also evident, this being that streamflows increase with increasing event duration. It is, however, apparent that there are minimal differences in the maps for the 5, 6 and 7 day events, suggesting that flood events seldom extend beyond 5 days.

11.3.2 Climate change impacts on design floods for different return periods

The projected impact of climate change on the 1-day duration design flood is shown in Fig. 11.6 for the six regional climate projections at the 2, 10 and 20 year return periods. Examination of the ratios of change reveals the following:

- In the southwest of the study area (i.e. the upper part of Basin A) there is a pattern of increasing flood magnitudes shown for 4 out of 6 of the climate projections (i.e. excluding CCAM-GFDLcm2.0 and CCAM-GFDLcm2.1). For 3 out of 4 of these projections (i.e. excluding CCAM-CSIROmk3.5), the intensity of the increase is greater and/or covers a larger area with increasing return period. In some cases the area of increasing flood magnitude extends into the upper section of Basin B.

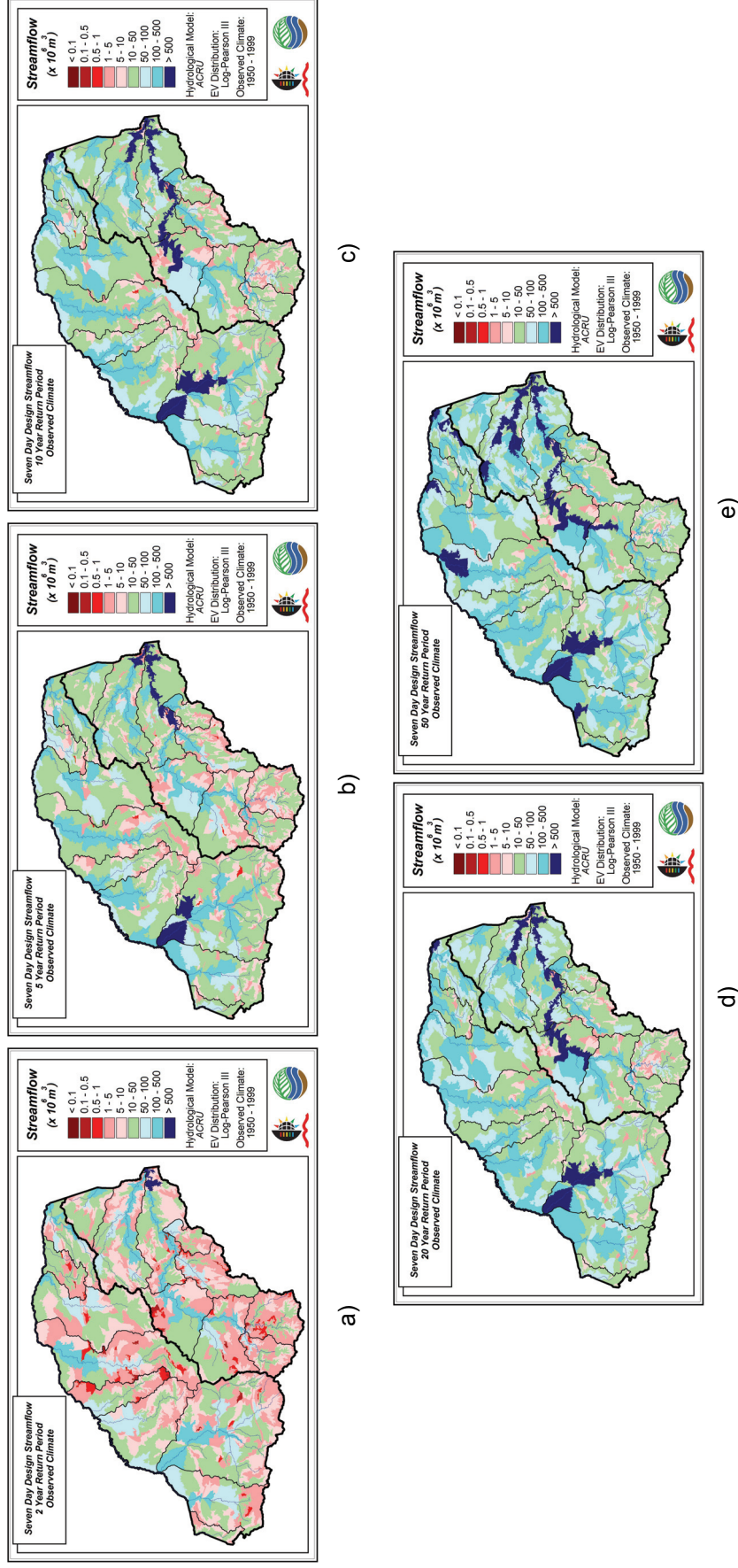


Fig. 11.4: Design flood estimates based on observed climate (1950-1999) for 7-day duration events at a) 2 year, b) 5 year, c) 10 year, d) 20 year and e) 50 year return periods

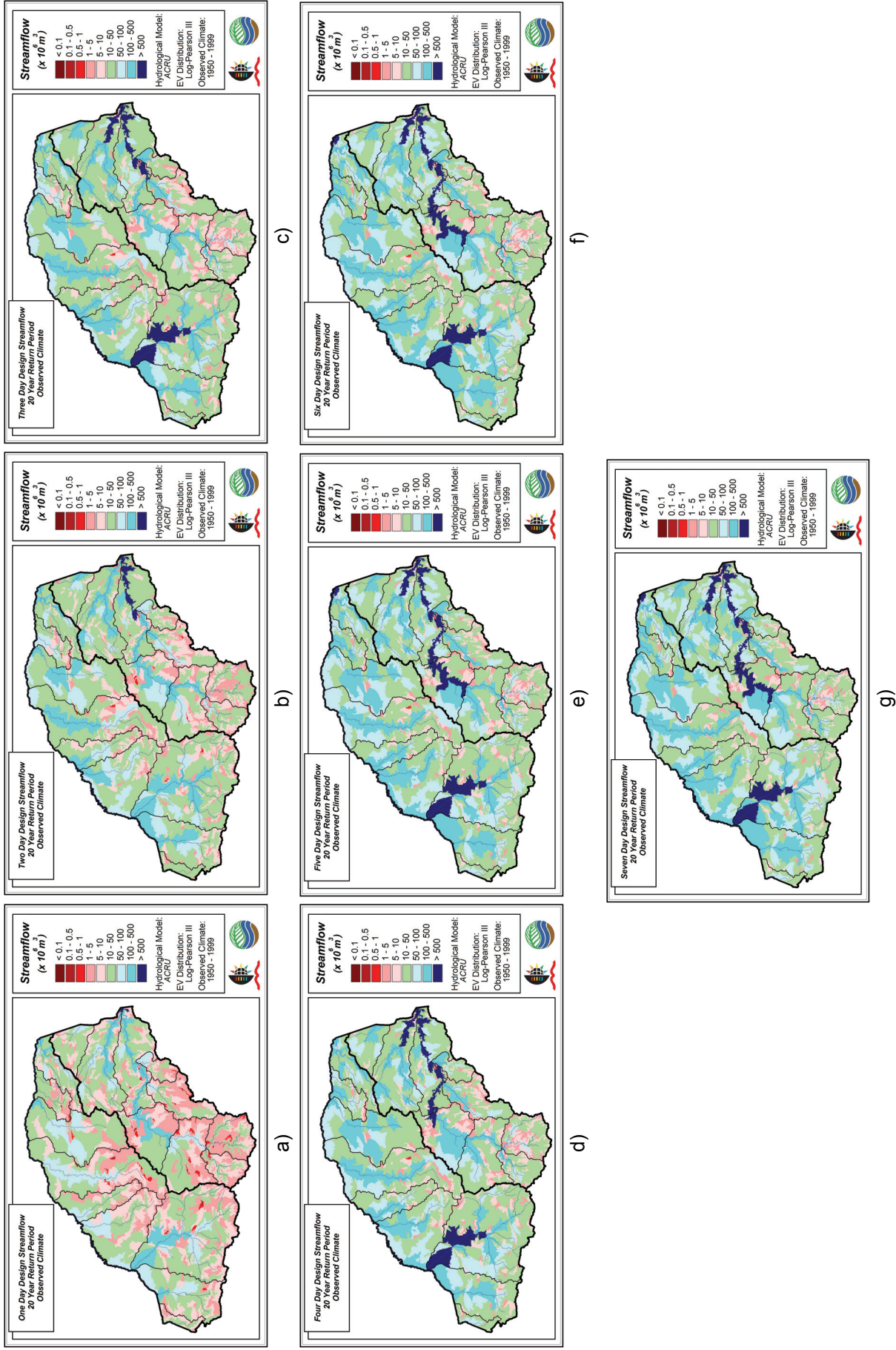


Fig. 11.5: Design flood estimates based on observed climate (1950-1999) for a 20-year return period for a) 1 day, b) 2 day, c) 3 day, d) 4 day, e) 5 day, f) 6 day and g) 7 day duration events

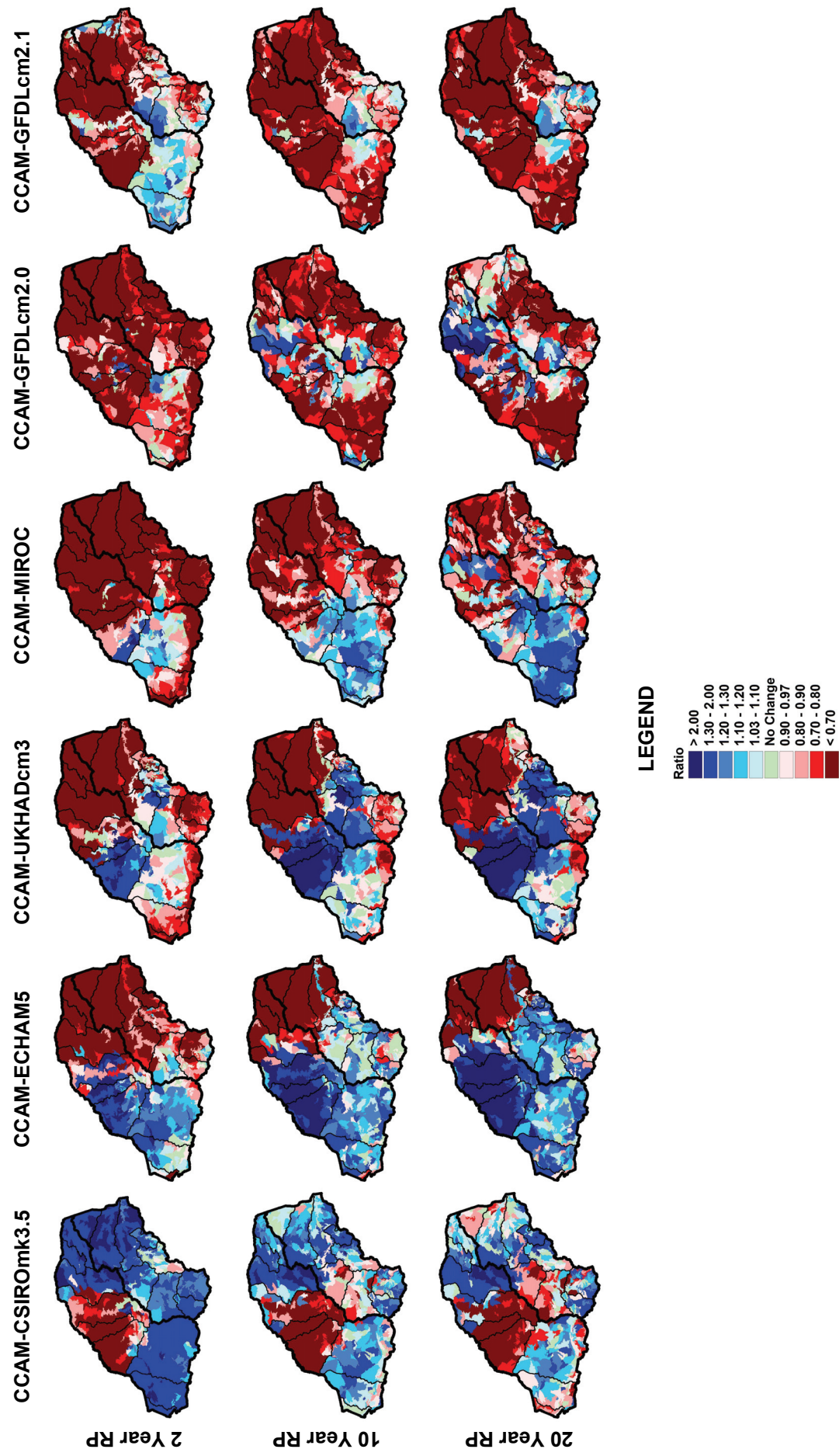


Fig. 11.6: Ratios of future to present 1-day duration design flood estimates derived from different regional climate projections for the 2, 10 and 20 year return periods

- In the north-east of the study area (i.e. lower sections of Basins A and B), there is a pattern of decreasing flood magnitudes shown for 5 out of 6 of the climate projections (i.e. excluding CCAM-CSIROmk3.5). For 4 out of 5 of these projections (i.e. excluding CCAM-GFDLcm2.1), the intensity and/or area of the decrease reduces with increasing return period.

The degree of concurrence amongst the climate projections with respect to the patterns identified was assessed further by considering concurrence at three different possible magnitudes of change (10% or more, 20% or more and 30% or more) for positive (increasing) and negative (decreasing) directions of change. The degree of concurrence was assumed to be a measure of confidence in the patterns identified, where confidence was defined in a manner similar to that in the IPCC AR4 (IPCC, 2007), as shown in Table 11.1.

Table 11.1: Definition of degrees of confidence relating to patterns of change in design flood estimates

Degree of Confidence	Number of Projections Depicting a Particular Pattern
No Confidence	0 out of 6
Very Low	1 out of 6
Low	2 out of 6
Medium	3 out of 6
High	4 out of 6
Very High	5 out of 6
Extremely High	6 out of 6

Maps of the degree of confidence associated with increases or decreases in design flood estimates are shown in Figs. 11.7, 11.8 and 11.9 for possible changes of 10% or more, 20% or more and 30% or more, respectively. For each possible of magnitude of change, confidence maps are presented for the 2, 10 and 20 year return periods. Figs. 11.7 to 11.9 reveal that:

- The pattern of increasing flood magnitudes in the southwest of the study area is only evident at High or Very High confidence at the 10% or more level of change. Areas of confidence increase with increasing return period, in line with trends identified previously. For the higher magnitudes of change (20% or more, 30% or more), a pattern of increasing flood estimates exists (mostly) at only Low or Very Low confidence.
- The pattern of decreasing flood magnitudes in the northeast of the study area is clearly evident at High and greater levels of confidence for all three magnitudes of change, with significant areas of confidence still evident at the 30% or more magnitude of change (suggesting a sharp decrease is likely).

11.3.2 Climate change impacts on design floods for different event durations

To assess whether the patterns in Fig. 11.6 would be consistent for longer duration events, the changes in the 2, 10 and 20 year return period design floods were mapped for the 1, 2, 3, 4, 5, 6 and 7 day event durations for each climate projection. The analysis showed that for each climate projection there was almost no discernible difference in the maps for various event durations for a particular return period. The maps for the CCAM-CSIROmk3.5 climate projection are shown in Fig. 11.10 as an example. The map for the 7-day event duration is omitted owing to space constraints in the figure. The above analysis implies that the patterns identified in Fig. 11.6 are consistent for longer duration events.

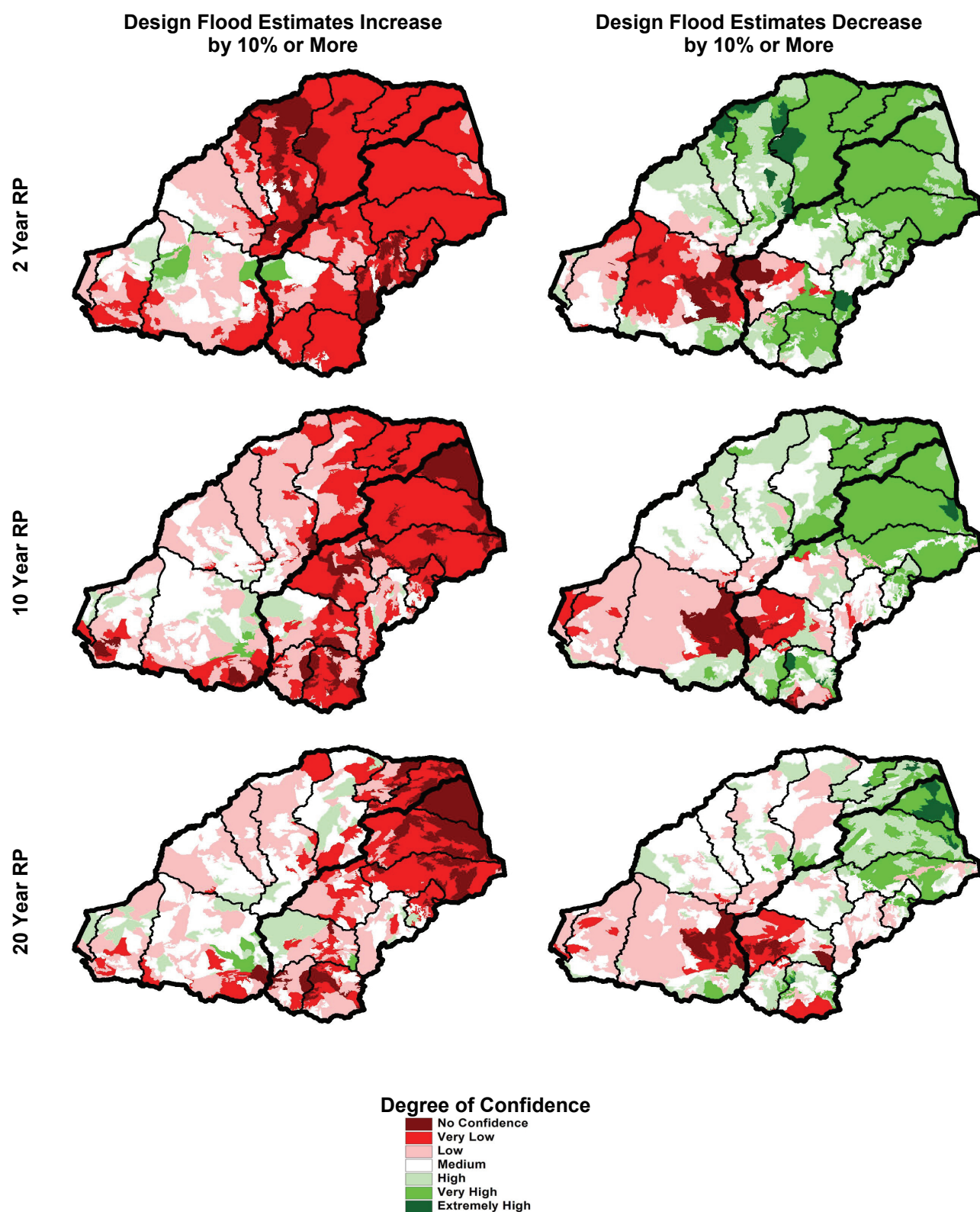


Fig. 11.7: Degree of confidence that design flood estimates will increase or decrease by 10% or more for the 2, 10 and 20 year return periods

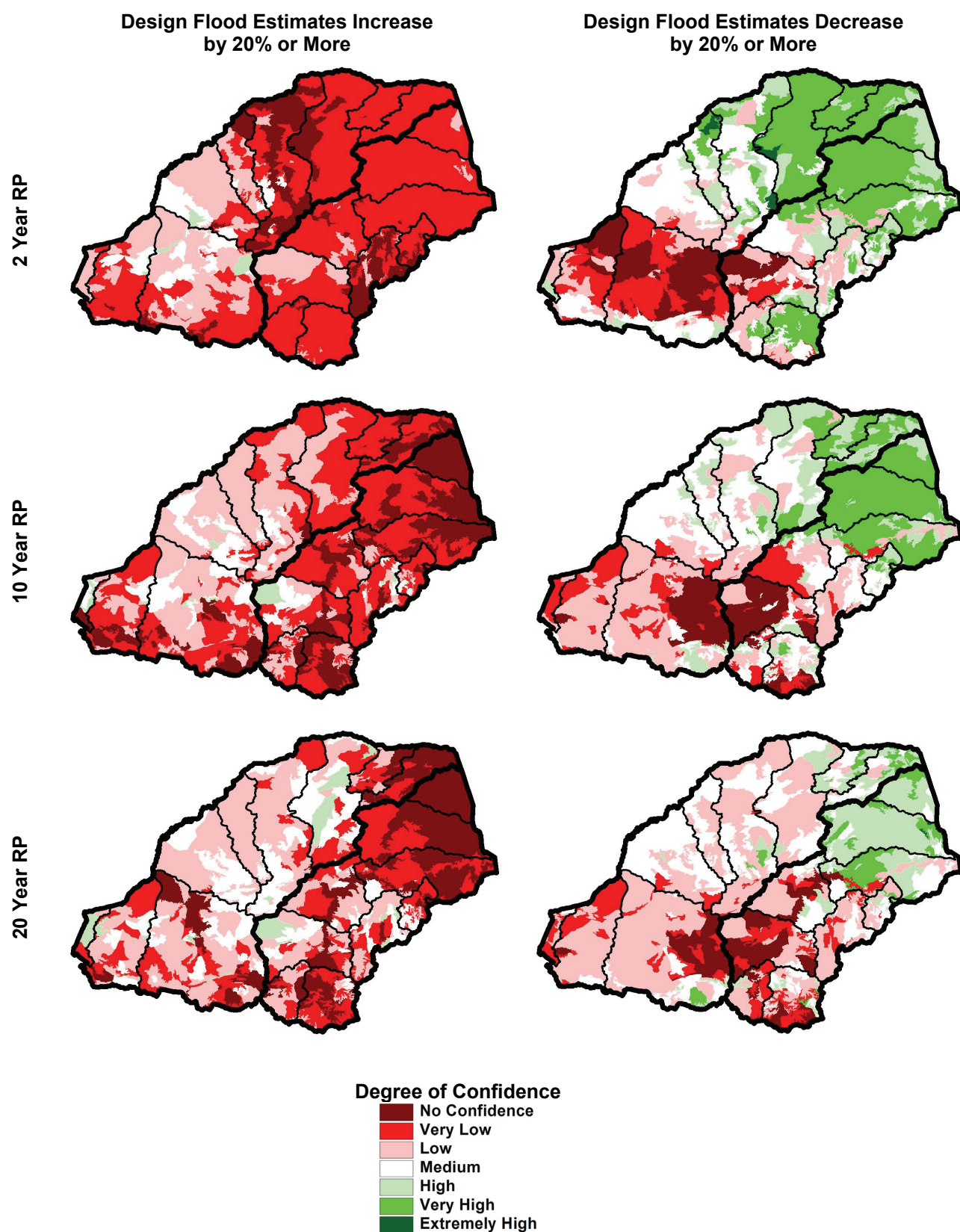


Fig. 11.8: Degree of confidence that design flood estimates will increase or decrease by 20% or more for the 2, 10 and 20 year return periods

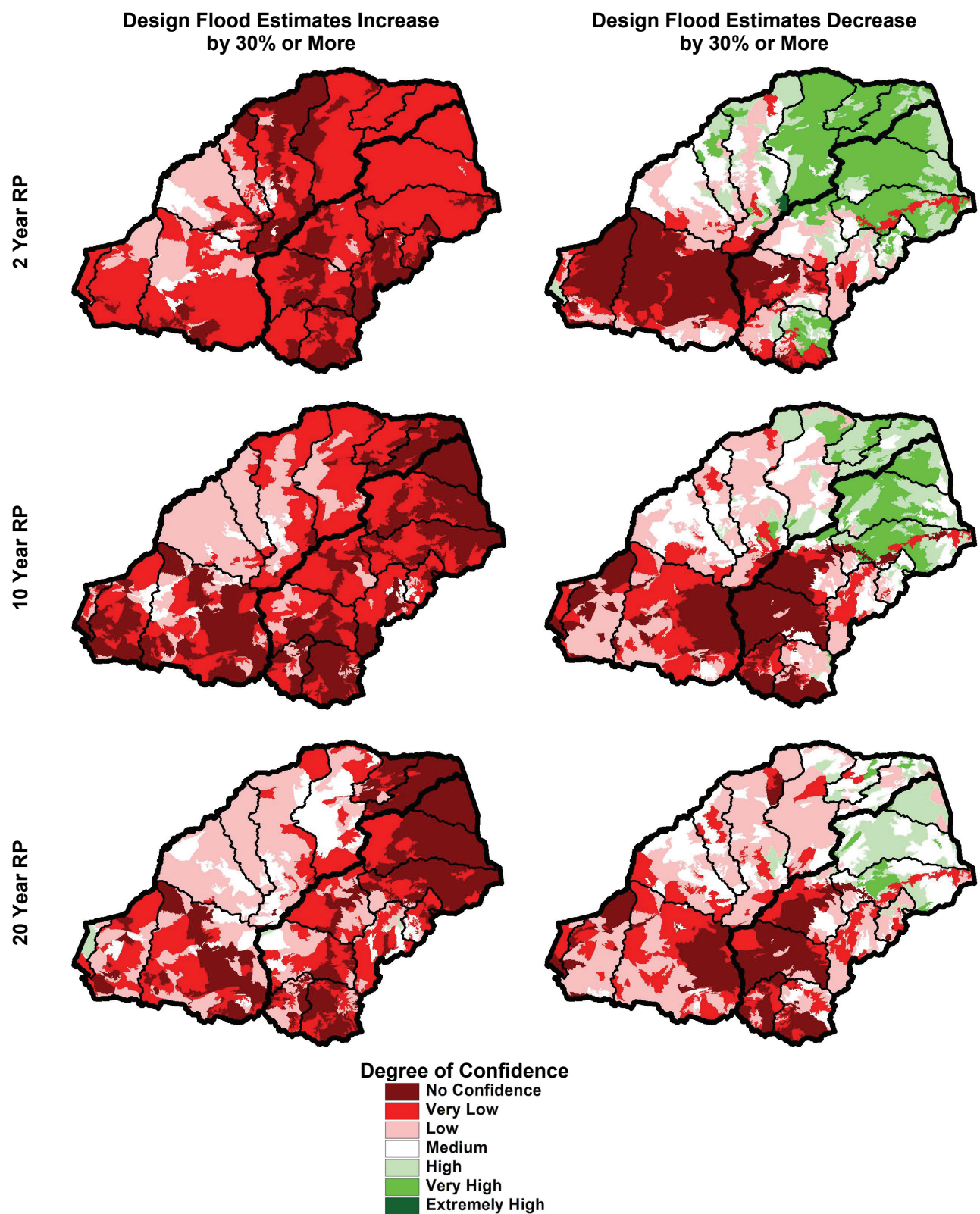


Fig. 11.9: Degree of confidence that design flood estimates will increase or decrease by 30% or more for the 2, 10 and 20 year return periods

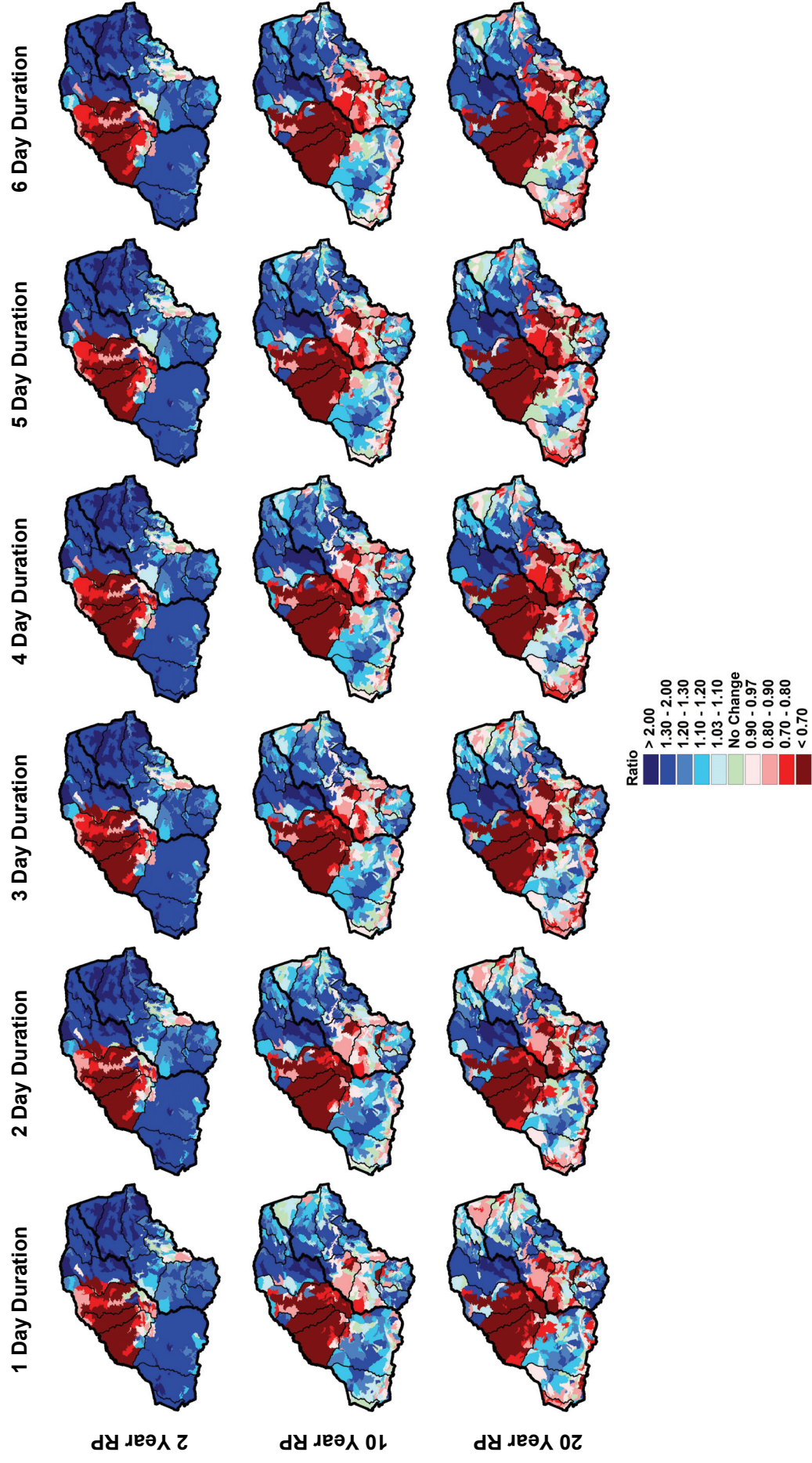


Fig. 11.10: Ratios of future to present design flood estimates derived from the CCAM-CSIROmk3.5 climate projection for the 1 to 6 day event durations for the 2, 10 and 20 year return periods

11.4 Discussion and Conclusions

The pattern of decreasing flood magnitudes in the northeast of the study area (lower sections of Primary Basins A and B) reflects the finding of the climate analyses (reported elsewhere) that the preferred continental tracks of the tropical low pressure systems and tropical cyclones shift northward with a concurrent decrease in the January to March rainfall over affected areas. This part of the study area (northeast) is currently most affected by tropical weather systems. The finding that the decreasing flood magnitudes are least evident at higher return periods, possibly indicates that the more severe (and less frequent) tropical weather systems still penetrate the study area in future.

The pattern of increasing flood magnitudes in the southwest of the study area (upper sections of Primary Basins A and B) is likely to be the result of other changes in the climate, for example, an increase in rainfall convectivity. These projected increases are severe in some cases (e.g. more than double the current flood magnitude). If these projected increases were to materialize (the associated confidence is relatively low), this would place a strain on current water related infrastructure (e.g. dams, bridges, culverts, etc.) and require a change in design procedures for future infrastructure.

The finding that projected impacts of climate change on design flood estimates vary with return period, rather than with event duration, concurs with a study by Schulze and Knoesen (2011) which had a national focus and also used multiple climate projections (based on a different downscaling methodology and mostly different global climate models). Schulze and Knoesen (2011) did not, however, show the marked reductions in design flood estimates in the lower sections of Primary Basins A and B that are a key finding of this report.

Chapter 12

Conclusions and Recommendations for Further Research

The project has given an overview of the influence of tropical systems from the Southwest Indian Ocean over the Limpopo River Basin and through this created an awareness of the collective impacts and characteristics of a synoptic weather system that has to date only been considered on an individual basis in scientific papers. It has further revealed the importance of considering not only precipitation data for determining climate variability but also synoptic systems responsible for variability. By focussing on tropical systems from the SWIO, an important feature of decadal-scale variability has been revealed, namely the specific positioning of large-scale circulation anomalies surrounding southern Africa and the influence thereof on the steering of tropical systems from the SWIO – having an impact on the rainfall in the region. The link between the regional and hemispherical modes of variation will further assist in identifying and explaining the drivers of the variability towards meaningful seasonal and decadal-scale predictability.

On climate change time scales, the findings of the project suggest under anthropogenic forcing a northward displacement of tropical systems from the SWIO with regional circulation changes such as the strengthening of the subtropical ridge over parts of southern Africa, increasing vertical wind shear and increasing static stability. This leads to a simulated decrease in precipitation over the Limpopo River Basin.

Recommendations based on the experiences during the project are listed according to two of the main themes of the study, namely decadal-scale variability and climate change research.

12.1 Decadal-scale variability

- The study demonstrated that taking a synoptic system approach instead of only considering total rainfall can bring new insights to the reasons behind variability. Utilization of a similar approach to investigate variation of climate on decadal scale in other regions is recommended.
- While we now have a clearer understanding of synoptic-scale variability responsible for the bi-decadal oscillation noted previously over the region, the drivers of that variability still need to be identified and explained.
- Further research should be conducted to explain the variation (through inclusion of external forcings also such as solar/aerosol/ozone variation and feedback effects within and from the stratosphere to subsurface ocean to understand the variation found in the troposphere affecting frequencies, intensities and tracks of synoptic weather systems within the larger scale variation of planetary waves, jet streams and circulation cells. It is proposed that a coupled model be used to understand the decadal-scale variability taking the aforementioned into account.

12.2 Climate change research

- As new climate simulations become available through the COordinated Regional climate Downscaling Experiment (CORDEX), the study of the effect of

anthropogenically enhanced climate forcing on Tropical Cyclone Climatology in the region can be supplemented/repeated with output from a series of AR5 simulations.

- The gridded time series of rainfall produced by the CCAM downscalings of global model output should be assessed against station data in terms of the daily distribution of rainfall. The gridded nature of the projections implies that rainfall is calculated as an areal average (over a 0.5 degree resolution grid cell). Thus, there is a potential that the gridded projections will over-simulate frequency and under-simulate magnitudes of daily rainfalls, relative to point-scale observed station data. This would impact *ACRU* streamflow simulations, since the model's development was based on the use of point-scale station data. An over-simulation of rainfall frequency and under-simulation of rainfall magnitude would lead to an under-simulation of streamflow since the threshold of rainfall for runoff generation would not be exceeded as often. In terms of assessing changes in streamflow, this effect may be offset by the ratio approach to assessing changes, where future flows are divided by present flows, thus potentially cancelling out errors in streamflow simulation.
- There is a need for a more in-depth understanding of the changes in climate that produce the changes in design floods. For example, how does rainfall change at annual, seasonal and daily time scales? How frequently is the rainfall threshold for runoff generation exceeded? How does rainfall intensity change? How is antecedent soil moisture affected?
- The future period investigated in this study was 2071-2100. Earlier periods within the 1961-2100 time series should be investigated to assess the timing of changes and when action may be required.
- Longer windows of climate (i.e. greater than 30 years) should be considered to allow for the assessment of longer return periods, e.g. 50 and 100 years. Longer return periods become critical for the design of larger dams.
- The climate projections that were developed and applied in this study were based on the A2 (business-as-usual) emissions scenario. Projections should be developed and applied for other emissions scenarios to allow planners to consider alternate development pathways.
- Actual land use, dams and abstractions of water should be considered in hydrological modelling for a more comprehensive assessment of flood risks to water related infrastructure.
- Depending on the outcome of the above investigations, possible changes in design criteria for future water related infrastructure should be investigated.

Chapter 13

Knowledge Dissemination and Capacity Development

13.1 Presentations at conferences

- Malherbe J. 2008. Westward moving tropical systems over northeastern South Africa: Trends and Climatology. 24th Annual Conference of the South African Society for Atmospheric Sciences, September 2008, Pretoria.
- Malherbe J. 2009. Cyclicity and trends of landfalling tropical systems over the eastern interior of southern Africa. 25th Annual Conference of the South African Society for Atmospheric Sciences, September 2009, Tulbach.
- Malherbe J. 2010. Multi-model analysis of expected future trends in the landfall of tropical systems from the southwest Indian Ocean over the eastern parts of southern Africa. 26th Annual Conference of the South African Society for Atmospheric Sciences, September 2010, Gariep.
- Malherbe J. 2011. High resolution projections of tropical cyclone landfall over southern Africa under enhanced anthropogenic forcing. 27th Annual Conference of the South African Society for Atmospheric Sciences, September 2011, Pretoria.
- Theale C. 2011. Determining the synoptic conditions that favour landfall of tropical cyclones over southern Africa with the aid of Self-Organizing-Maps. 27th Annual Conference of the South African Society for Atmospheric Sciences, September 2011, Pretoria.

13.2 Scientific papers

- Malherbe J, Engelbrecht FA, Landman WA and Engelbrecht CJ. 2012. Tropical systems from the southwest Indian Ocean making landfall over the Limpopo River Basin, southern Africa: A historical perspective. *International Journal of Climatology* **32**: 1018-1032. DOI: 10.1002/joc.2320.
- Engelbrecht FA. 2010. *The scientific basis for climate change over southern Africa*. In: Climate Change and Trade: The challenges for Southern Africa. South African Institute for International Affairs.
- Malherbe J, Engelbrecht FA and Landman WA. Projected changes in tropical cyclone climatology and landfall in the southwest Indian Ocean region under enhanced anthropogenic forcing. *Climate Dynamics*. Accepted.
- Tshiala MF, Olwoch JM and Engelbrecht FA. Trends in the occurrence of heat waves over the Limpopo Province in South Africa. *Climate and Development Journal*. Submitted.

Chapter 14

References

- Arblaster JM, Meehl GA. 2006. Contributions of external forcings to southern annular mode trends. *Journal of Climate* **19**: 2896-2905.
- Acocks JPH. 1988. Veld Types of southern Africa. Botanical Research Institute, Pretoria, RSA. *Botanical Survey of South Africa Memoirs* **57**: 146 pp.
- Alexander WJR. 1995. Floods, droughts and climate change. *South African Journal of Science* **91**: 403-407.
- Becker S, B ker ML, Matyas CJ and Rohli RV. 2010. Assessing links between atmospheric vorticity patterns and directional changes in hurricane tracks. *Theoretical and Applied Climatology* **102**: 397-392.
- Bell GD and Chelliah M. 2006. Leading tropical modes associated with interannual and multidecadal fluctuations in North Atlantic hurricane activity. *Journal of Climate* **19**: 590-612.
- Bengtsson L, Botzet M and Esch M. 1995. Hurricane-type vortices in a general circulation model. *Tellus* **47A**: 175-196.
- Bengtsson L, Hodges KI, Esch M, Keenlyside N, Kornbleuh L, Luo J-J and Yamagata T. 2007. How may tropical cyclones change in a warmer climate? *Tellus* **59A**: 539-561. DOI: 10.1111/j.1600-0870.2007.00251.x.
- Blender R and Schubert M. 2000. Cyclone tracking in different spatial and temporal resolutions. *Monthly Weather Review* **128**: 377-384.
- Camargo SJ, Emanuel KA and Sobel AH. 2007. Use of a genesis potential index to diagnose ENSO effects on tropical cyclone genesis. *Journal of Climate* **20**: 4819-4834. DOI: 10.1175/JCLI4282.1.
- Caron LP and Jones GJ. 2008. Analysing present, past and future tropical cyclone activity as inferred from an ensemble of coupled global climate models. *Tellus* **60A**: 80-96.
- Cavazos T. 2000. Using Self-Organizing Maps to investigate extreme climate events: An application to wintertime precipitation in the Balkans. *Journal of Climate* **13**: 1718-1732.
- Chan JCL. 2005. The physics of tropical cyclone motion. *Annual Reviews in Fluid Mechanics* **37**: 99-128.
- Chan JCL, Ko FMF and Lei YM. 2002. Relationship between potential vorticity tendency and tropical cyclone motion. *Journal of Atmospheric Sciences* **59**: 1317-1336.
- Chan JCL and Liu KS. 2004. Global warming and western North Pacific typhoon activity from an observational perspective. *Journal of Climate* **70**: 4590-4602.
- Chan JCL and Xu M. 2009. Inter-annual and inter-decadal variations of landfalling tropical cyclones in East Asia. Part I: time series analysis. *International Journal of Climatology* **29**: 1285-1293. DOI: 10.1002/joc.1782.

- Christensen JH, Hewitson B, Busuioc A, Chen A, Gao X, Held I, Jones R, Kolli RK, Kwon W-T, Laprise R, Magaña Rueda V, Mearns L, Menéndez CG, Räisänen J, Rinke A, Sarr A and Whetton P. 2007. Regional Climate Projections. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
- Crimp SJ and Mason SJ. 1999. The extreme precipitation event of 11 to 16 February 1996 over South Africa. *Meteorology and Atmospheric Physics* **70**: 29-42. DOI: 10.1007/s007030050023.
- D'Abreton, PC and Lindesay JA. 1993. Water vapour transport over Southern Africa during wet and dry early and late summer months. *International Journal of Climatology* **13**: 151-170. DOI: 10.1002/joc.3370130203.
- Dong K and Neumann CJ. 1986. The relationship between tropical cyclone motion and environmental geostrophic flows. *Monthly Weather Review* **114**: 115-122.
- Dunn P. 1985. An investigation into tropical cyclones in the SW Indian Ocean, Flood studies: *Technical Note 1*. Department of Water Affairs, Pretoria, RSA.
- Dyer TGJ and Tyson PD. 1977. Estimating above and below normal rainfall periods over South Africa, 1972-2000. *Journal of Applied Meteorology* **16**: 145-147. DOI: 10.1175/1520-0450(1977)016<0145:EAABNR>2.0.CO.
- Dyson LL. 2009. Heavy daily-rainfall characteristics over the Gauteng Province. *Water SA* **35**: 627-638.
- Dyson LL and van Heerden J. 2002. A model for the identification of tropical weather systems over South Africa. *Water SA* **28**: 249-258. ISSN 0378-4738.
- Elsner BE, Kossin JP and Jagger TH. 2008. The increasing intensity of the strongest tropical cyclones. *Nature* **455**: 92-95. DOI: 10.1038/nature07234.
- Emanuel KA and Nolan DS. 2004. Tropical cyclone activity and global climate. In: *Proc. of 26th Conference on Hurricanes and Tropical Meteorology*. pp. 240-241. American Meteorological Society, Miami, FL, USA.
- Engelbrecht CJ, Engelbrecht FA and Dyson LL. 2012. High-resolution model-projected changes in mid-tropospheric closed-lows and extreme rainfall events over southern Africa. *International Journal of Climatology*. DOI: 10.1002/joc.3420. 15 pp.
- Engelbrecht FA, McGregor JL and Engelbrecht CJ. 2009. Dynamics of the Conformal-Cubic Atmospheric Model projected climate-change signal over southern Africa. *International Journal of Climatology* **29**: 1013-1033. DOI: 10.1002/joc.1742.
- Franklin JL, Feuer SE, Kaplan J and Aberson SD. 1996. Tropical cyclone motion and surrounding flow relationships: Searching for beta gyres in omega dropsonde datasets. *Monthly Weather Review* **124**: 64-84.
- Garner ST, Held IM, Knutson T and Sirutis J. 2009. The roles of wind shear and thermal stratification in past and projected changes of Atlantic tropical cyclone activity. *Journal of Climate* **22**: 4723-4734. DOI: 10.1175/2009JCLI2930.1.
- Gershunov A and Barnett TP. 1998. ENSO influence on intraseasonal extreme rainfall and temperature frequencies in the contiguous United States: Observations and model results. *Journal of Climate* **11**: 1575-1586.

- Gray WM. 1988. Environmental influences on tropical cyclones. *Australian Meteorological Magazine* **36**: 127-139.
- Gray WM. 1998. The formation of tropical cyclones. *Meteorology and Atmospheric Physics* **76**: 37-69.
- Harr PA and Russel LE. 1991. Tropical cyclone track characteristics as a function of large-scale circulation anomalies. *Monthly Weather Review* **119**: 1448-1467.
- Henderson-Sellers A, Zhang H, Emanuel K, Gray W, Lnadsea C, Holland G, Lighthill J, Shieh S-L, Webster P and McGuffie K. 1998. Tropical cyclones and global climate change: A post IPCC assessment. *Bulletin of the American Meteorological Society* **79**(1): 19-38.
- Hewitson BC and Crane RG. 2002. Self-organizing maps: applications to synoptic climatology. *Climate Research* **22**: 13-26.
- Hines KM, Bromwich DH and Marshall GJ. 2000. Artificial surface pressure trends in the NCEP-NCAR Reanalysis over the Southern Ocean and Antarctica. *Journal of Climate* **13**: 3940-3952.
- Ho CH, Kim JH, Jeong JH, Kim HS and Chen D. 2006. Variation of tropical cyclone activity in the South Indian Ocean: El Niño-Southern Oscillation and Madden-Julian Oscillation effects. *Journal of Geophysical Research* **111**: D22101, DOI:10.1029/2006JD007289.
- Hofmeyr WL and Gouws V. 1964. A statistical and synoptic analysis of wet and dry conditions in the northwestern Transvaal. *Notos* **13**: 37-48.
- Holland GJ. 1982. Tropical cyclone motion: environmental interaction plus a beta effect. *Journal of the Atmospheric Sciences* **40**: 328-342.
- Holland GJ. 1984. Tropical cyclone motion: A comparison of theory and observation. *Journal of the Atmospheric Sciences* **41**: 68-75.
- Holland GJ. 1997. The maximum potential intensity of tropical cyclones. *Journal of the Atmospheric Sciences* **54**: 2519-2541.
- IPCC. 2007. Summary for Policymakers. In: Parry ML, Canziani OF, Palutikof JP, van der Linden PJ and Hanson CE (eds). *Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. pp. 7-22. Cambridge University Press, Cambridge, UK.
- Jury M. 1993. A preliminary study of climatological associations and characteristics of tropical cyclones in the SW Indian Ocean. *Meteorology and Atmospheric Physics* **51**: 101-115. DOI: 10.1007/BF01080882.
- Jury MR. 1996. Regional teleconnection patterns associated with summer rainfall over South Africa, Namibia and Zimbabwe. *International Journal of Climatology* **16**: 135-153. DOI: 10.1002/(SICI)1097-0088(199602)16:2<135:AID-JOC4>3.0.CO;2-7.
- Jury MR and Nkosi SE. 2000. Easterly flow in the tropical Indian Ocean and climate variability over south-east Africa. *Water SA* **26**(2): 147-152.
- Jury MR and Pathack B. 1991. A study of climate and weather variability over the tropical southwest Indian Ocean. *Meteorology and Atmospheric Physics* **47**: 37-48. DOI: 10.1007/BF01025825.

- Kalnay E, Kanamitsu M, Kistler, R, Collins W, Deaven D, Gandin L, Iredell M, Saha S, White G, Woollen J, Zhu Y, Leetmaa A, Reynolds R, Chelliah M, Ebisuzaki W, Higgins W, Janowiak J, Mo KC, Ropelewski C, Wang J, Jenne R, and Joseph D. 1996. The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society* **77**: 437-457. DOI: 10.1175/1520-0477(1996)077<0437: TNYRP>2.0.CO.
- Kanamitsu M, Kistler RE and Reynolds RW. 1997. NCEP/NCAR Reanalysis and the use of satellite. *Advances in Space Research* **19(3)**: 481-489.
- Karoly DJ and Vincent DG (eds). 1998. Meteorology of the Southern Hemisphere. *Meteorological Monographs* vol. 27, no. 49. American Meteorological Society, Boston, USA. 410 pp.
- Katzfey JJ, McGregor JL, Nguyen KC and Thatcher M. 2009. Dynamical downscaling techniques: impacts on regional climate change signals. In: Anderssen RS, Braddock RD and Newham LTH (eds). *18th World IMACS Congress and MODSIM09 International Congress on Modelling and Simulation. Modelling and Simulation Society of Australia and New Zealand and International Association for Mathematics and Computers in Simulation*, July 2009, pp. 2377-2383. ISBN: 978-0-9758400-7-8.
- Kite GW. 1988. Frequency and Risk Analysis in Hydrology. Water Resources Publications, Fort Collins, Colorado, USA.
- Klinman MG and Reason CJC. 2008. On the peculiar storm track of TC Favio during the 2006-2007 Southwest Indian Ocean tropical cyclone season and relationships to ENSO. *Meteorology and Atmospheric Physics* **100**: 233-242. DOI: 10.1007/s00703-008-0306-7.
- Kohonen T. 2001. Self-Organising Maps, 3rd edition. Springer-Verlag, Berlin Heidelberg New York.
- Kohonen T, Hynninen J, Kangas J and Laaksonen J. 1996. SOM_PAK: The Self-Organizing Map Program Package. Technical Report A31. Laboratory of Computer and Information Science, Helsinki University of Technology, Finland. 24 pp.
- Kruger AC. 1999. The influence of the decadal-scale variability of summer rainfall on the impact of El Niño and La Niña events in South Africa. *International Journal of Climatology* **19**: 59-68.
- Lambert SJ. 1988. A cyclone climatology of the Canadian Climate Centre general circulation model. *Journal of Climate* **1**: 109-115.
- Landman WA and Goddard L. 2005. Predicting southern African summer rainfall using a combination of MOS and perfect prognosis, *Geophysical Research Letters* **32**: L15809. DOI: 10.1029/2005GL022910.
- Liebmann B, Hendon HH and Glick JD. 1994. The relationship between tropical cyclones of the Western Pacific and Indian Oceans and the Madden-Julian Oscillation. *J Met Soc Jap* **72(3)**: 401-412
- Livezey RE and Chen WY. 1983. Statistical field significance and its determination by Monte Carlo techniques. *Monthly Weather Review* **111**: 46-59.
- Luengo F, Cofino AS and Gutierrez JM. 2004. GRID Oriented implementation of Self-organizing Maps for data mining in meteorology. *Lecture Notes in Artificial Intelligence* **2970**: 163-171.

- Malherbe J, Engelbrecht FA, Landman WA and Engelbrecht CJ. 2012. Tropical systems from the southwest Indian Ocean making landfall over the Limpopo River Basin, southern Africa: a historical perspective. *International Journal of Climatology* **32**: 1018-1032. DOI: 10.1002/joc.2320.
- Mandal M, Mohanty UC, Sinha P and Ali MM. 2007. Impact of sea surface temperature in modulating movement and intensity of tropical cyclones. *Natural Hazards* **41**: 413-427.
- Mann H. 1945. Nonparametric tests against trend. *Econometrica* **13**: 245-259.
- Martin LM and Chan CL. 2004. Tropical cyclone intensity and vertical wind shear. *Bulletin of the American Meteorological Society* **61**: 1859-1876.
- Matarira CH and Jury MR. 1992. Contrasting meteorological structure of intra-seasonal wet and dry spells in Zimbabwe. *International Journal of Climatology* **12**: 165-176. DOI: 10.1002/joc.3370120205.
- Mavume AF, Rydberg L, Rouault M and Lutjeharms JRE. 2009. Climatology and landfall of tropical cyclones in the southwest Indian Ocean. *WIO Journal of Maritime Science* **8(1)**: 15-36.
- McGregor JL. 2005. C-CAM: Geometric aspects and dynamical formulation [electr. publication]. *CSIRO Atmospheric Research Tech. Paper* No. 70, 43 pp.
- Miron O and Tyson PD. 1984. Wet and dry conditions and pressure anomaly fields over South Africa and the adjacent oceans, 1963-1979. *Monthly Weather Review* **112**: 2127-2132.
- Mulenga HM, Rouault M and Reason CJC. 2003. Dry summers over northeastern South Africa and associated circulation anomalies. *Climate Research* **25**: 29-41.
- Neumann CJ. 1976. Statistical prediction of tropical cyclone motion over the southwest Indian Ocean. *Monthly Weather Review* **104**: 76-85.
- Nguyen KC and Walsh KJE. 2001. Interannual, decadal, and transient greenhouse simulation of tropical cyclone-like vortices in a regional climate model of the South Pacific. *Journal of Climate* **14**: 3043-3054.
- Oouchi K, Yoshimura J, Yoshimura H, Mizuta R, Kusunoki S and Noda A. 2006. Tropical cyclone climatology in a global-warming climate as simulated in a 20 km-mesh global atmospheric model: frequency and wind intensity analysis. *Journal of Meteorological Society Japan* **84(2)**: 259-276.
- Ose T and Arakawa O. 2009. Characteristics of the CMI3 models simulating realistic response of Tropical Western Pacific precipitation to Nino3 SST variability. *Journal of the Meteorological Society of Japan* **87(4)**: 807-819. DOI: 10.2151/jmsj.87.807.
- Padgett G. 2007. Monthly Global Tropical Cyclone Summary: February 2007. http://www.maybagyo.com/garyp_mgtcs/feb07sum.htm.
- Preston-Whyte RA and Tyson PD. 2000. The atmosphere and weather of southern Africa. Oxford University Press: Southern Africa.
- Reason CJC. 2007. Tropical Cyclone Dera, the unusual 2000/01 tropical cyclone season in the Southwest Indian Ocean associated rainfall anomalies over Southern Africa. *Meteorology and Atmospheric Physics* **97**: 181-188.

- Reason CJC, Hachigonta S and Phaladi RF. 2005. Interannual variability in rainy season characteristics over the Limpopo region of southern Africa. *International Journal of Climatology* **25**: 1835-1853. DOI: 10.1002/joc.1228.
- Reason CJC and Keibel A. 2004. Tropical Cyclone Eline and its unusual penetration over the Southern African mainland. *Weather and Forecasting* **19**: 789-805. DOI: 10.1175/1520-0434(2004)019.
- Reason CJC and Rouault M. 2002. ENSO-like decadal variability and South African rainfall. *Geophysical Research Letters* **29(13)**. DOI: 10.1029/2002GL014663.
- Richard Y, Fauchereau N, Pocard I, Rouault M and Trzaska S. 2001. 20th Century droughts in southern Africa: Spatial and temporal variability, teleconnections with oceanic and atmospheric conditions. *International Journal of Climatology* **21**: 873-885. DOI: 10.1002/joc.656.
- Royer J-F, Chauvin F, Timbal P, Araspin P and Grimal D. 1998. A GCM study of the impact of greenhouse gas increase on the frequency of occurrence of tropical cyclones. *Climate Change* **38**: 307-343.
- Sall SM, Sauvaeot H, Gaye AT, Viltard A and de Felice P. 2006. A cyclogenesis index for tropical Atlantic off the African coasts. *Atmospheric Research* **79**: 123-147.
- Sato N, Takahashi C, Seiki A, Yoneyama R and Ryuichi S. 2009. An evaluation of the reproducibility of the Madden-Julian Oscillation in the CMIP3 Multi-Models. *Journal of the Meteorological Society of Japan* **87**: 791-805. DOI: 10.2151/jmsj.87.791.
- Schulze RE. 1965. Climate of South Africa. 8. General Survey. Pretoria: Government Printer and Weather Bureau.
- Schulze RE. 1995. Hydrology and Agrohydrology: A text to accompany the ACRU 3.00 Agrohydrological Modelling System. Water Research Commission, Pretoria, RSA, WRC Report No. TT69/95 (ISBN 1 86845 136 4). 552 pp.
- Schulze RE. 2004. Modelling as a tool in integrated water resources management: Conceptual issues and case study applications. Water Research Commission, Pretoria, RSA. WRC Report No. 749/1/04 (ISBN 1-77005-143-0). 258 pp.
- Schulze RE. 2010. Methods 3: Modelling impacts of climate change on the hydrological system: Model requirements, selection of the ACRU model, its attributes and computations of major state variables and outputs. In: Schulze RE, Hewitson BC, Barichievy KR, Tadross MA, Kunz RP, Horan, MJC and Lumsden TG. *Methodological Approaches to Assessing Eco-Hydrological Responses to Climate Change in South Africa*. Water Research Commission, Pretoria, RSA. WRC Report No. 1562/1/10 (ISBN 978-1-4312-0050-4). Chapter 8, 75-88.
- Schulze RE. 2011. Summary of results and key findings from the 2011 study on climate change and the water related sector in South Africa. In: Schulze RE. *A 2011 Perspective on Climate Change and the South African Water Sector*. Water Research Commission, Pretoria, RSA, WRC Report No. 1843/2/11. Chapter 9.1, 257-274.
- Schulze RE and Horan MJC. 2010. Methods 1: Delineation of South Africa, Lesotho and Swaziland into Quinary Catchments. In: Schulze RE, Hewitson BC, Barichievy KR, Tadross MA, Kunz RP, Horan MJC and Lumsden TG. *Methodological Approaches to Assessing Eco-Hydrological Responses to Climate Change in South Africa*. Water Research Commission, Pretoria, RSA. WRC Report No. 1562/1/10 (ISBN 978-1-4312-0050-4). Chapter 6, 55-62.

- Schulze RE, Horan MJC, Kunz RP, Lumsden TG and Knoesen DM. 2010. Methods 2: Development of the Southern African Quinary Catchments Database. In: Schulze RE, Hewitson BC, Barichievy KR, Tadross MA, Kunz RP, Horan MJC and Lumsden TG. *Methodological Approaches to Assessing Eco-Hydrological Responses to Climate Change in South Africa*. Water Research Commission, Pretoria, RSA. WRC Report No. 1562/1/10 (ISBN 978-1-4312-0050-4). Chapter 7, 63-74.
- Schulze RE and Knoesen DM. 2011. Climate change and design streamflows: A 2011 perspective. In: Schulze RE. *A 2011 Perspective on Climate Change and the South African Water Sector*. Water Research Commission, Pretoria, RSA, WRC Report No. 1843/2/11. Chapter 7.3, 213-219.
- Schulze RE, Lumsden TG, Horan MJC, Warburton M. and Maharaj M. 2005. An assessment of impacts of climate change on agrohydrological responses over Southern Africa. In: Schulze RE. *Climate Change and Water Resources in Southern Africa: Studies on Scenarios, Impacts, Vulnerabilities and Adaptation*. Water Research Commission, Pretoria, RSA, WRC Report No. 1430/1/05. Chapter 9, 141-189.
- Singh OP, Khan TMA and Rahman MDS. 2001. Probable reasons for enhanced cyclogenesis in the Bay of Bengal during July-August of ENSO years. *Global Plan* Chapter 29: 135-147. DOI: 10.1016/S0921-8181(00)00090-4.
- Sturaro G. 2003. A closer look at the climatological discontinuities present in the NCEP/NCAR reanalysis temperature due to the introduction of satellite data. *Climate Dynamics* **21**: 309-316.
- Sugi M, Murakami H and Yoshimura J. 2009. A reduction in global tropical cyclone frequency due to global warming. *Sola* **5**: 164-167. DOI: 10.2151/sola.2009-042.
- Sugi M and Noda A. 2002. Influence of global warming on tropical cyclone climatology: An experiment with the JMA global model. *Journal of the Meteorological Society of Japan* **80**: 249-272.
- Takatsuka M. 2001. An application of Self-Organizing Map and Interactive 3-D Visualization to Geospatial Data. *The Proceedings of the 6th International Conference on GeoComputation*, September, Brisbane, Australia.
- Taljaard JJ. 1985. Cut-off lows in the South African region. *South African Weather Bureau Technical Paper* **14**: 153 pp.
- Taylor KE. 2001. Summarizing multiple aspects of model performance in a single diagram. *Journal of Geophysical Research* **106**: 7183-7192.
- Tennant WJ. 2004. Considerations when using pre-1979 NCEP/NCAR reanalysis in the southern hemisphere. *Geophysical Research Letters* **31**: L11112. DOI: 10.1029/2004GLO197501.
- Tsutsui T. 2002. Implications of anthropogenic climate change for tropical cyclone activity: A case study with the NCAR CCM2. *Journal of the Meteorological Society of Japan* **80**: 45-65.
- Tyson PD. 1981. Atmospheric circulation variations and the occurrence of extended wet and dry spells over southern Africa. *Journal of Climatology* **1**: 115-130.

- Tyson PD, Sturman AP, Fitzharris BB, Mason SJ and Owens IF. 1997. Circulation changes and teleconnections between glacial advances on the west coast of New Zealand and extended spells of drought years in South Africa. *International Journal of Climatology* **17**: 1499–1512. doi: 10.1002/(SICI)1097-0088(19971130)17:14<1499::AID-JOC207>3.0.CO;2-O.
- Tyson PD, Cooper GRJ and McCarthy TS. 2002. Millennial to multi-decadal variability in the climate of southern Africa. *International Journal of Climatology* **22**: 1105-1117. DOI: 10.1002/joc.787.
- Van Ulden AP and Van Oldenborgh GJ. 2006. Large-scale atmospheric circulation biases and changes in global climate model simulations and their importance for climate change in Central Europe. *Atmospheric Chemistry and Physics* **6**: 863-881.
- Vitart F, Anderson L and Stern WF. 1999. Impact of large-scale circulation on tropical storm frequency, intensity and location, simulated by an ensemble of GCM integrations. *Journal of Climate* **12**: 3237-3254. DOI: 10.1175/1520-0442(1999)012<3237:IOLSCO>2.0.CO;2.
- Vitart F, Anderson A and Stockdale T. 2003. Seasonal forecasting of tropical cyclone landfall over Mozambique. *Journal of Climate* **16**: 3932-3945. DOI: 10.1175/1520-0442(2003)016<3932:SFOTCL>2.0.CO.
- Vogel C and O'Brien K. 2003. Climate forecasts in southern Africa. Chapter 1. In: O'Brien K and Vogel C (eds). *Coping with climate variability: The use of seasonal forecasts in southern Africa*. Ashgate: Aldershot, UK.
- Walsh KJE, Fiorino M, Landsea CW and McInnes KL. 2007. Objectively determined resolution-dependent threshold criteria for the detection of tropical cyclones in climate models and reanalysis. *Journal of Climate* **20**: 2307-2314.
- Walsh KJE and Katzfey JJ. 2000. The impact of climate change on the poleward movement of tropical cyclone-like vortices in a regional climate model. *Journal of Climate* **13**: 1116-1132.
- Walsh KJE, Nguyen K-C and McGregor JL. 2004. Fine resolution regional climate model simulations of the impact of climate change on tropical cyclones near Australia. *Climate Dynamics* **22**: 47-56.
- Walsh KJE and Ryan BF. 2000. Tropical cyclone intensity increase near Australia as a result of climate change. *Journal of Climate* **13**: 3029-3036.
- Walsh KJE and Watterson IG. 1997. Tropical cyclone-like vortices in a limited area model: Comparison with observed climatology. *Journal of Climate* **10**: 2240-2259. DOI: 10.1175/JCLI4074.1.
- Wang B and Chan JCL. 2002. Pacific-East Asian Teleconnection. Part II: How ENSO regulates tropical storm activity over the western North Pacific. *Journal of Climate* **15**: 1643-1658.
- Warburton ML, Schulze RE and Jewitt GPW. 2010. Confirmation of ACRU model results for applications in land use and climate change studies. *Hydrology and Earth System Sciences* **14**: 2399-2414.
- Webster PJ, Holland GJ, Curry JA and Chang H-R. 2005. Changes in tropical cyclone number, duration and intensity in a warming environment. *Science* **309**: 1844-1846. DOI: 10.1126/science.1116448.

- Wilks DS. 2006. Statistical Methods in the Atmospheric Sciences, 2nd Edition. Academic Press, San Diego, CA, USA.
- Xie P and Arkin PA. 1997. A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bulletin of the American Meteorological Society* **78**(11): 2539-2558.
- Yoshimura J and Sugi M. 2005. Tropical cyclone climatology in a high-resolution AGCM – impacts of SST warming and CO₂ increase. *SOLA* **1**: 133-136. DOI: 10.2151/sola.2005-035.
- Yu J, Wang Y and Hamilton K. 2010. Response of tropical cyclone potential to a global warming scenario in the IPCC AR4 CGCMs. *Journal of Climate* **23**: 1254-1373. DOI: 10.1175/2009JCLI2843.1.
- Zeng Z, Wang Y and Wu C-C. 2007. Environmental dynamical control of tropical cyclone intensity – an observational study. *Monthly Weather Review* **135**: 38-59. DOI: 10.1175/MWR3278.1.