

The copyright of this thesis vests in the author. No quotation from it or information derived from it is to be published without full acknowledgement of the source. The thesis is to be used for private study or non-commercial research purposes only.

Published by the University of Cape Town (UCT) in terms of the non-exclusive license granted to UCT by the author.

The Role of Ocean Dynamics within Tropical Atlantic Climate Variability

Natalie Burls



Thesis Presented for the Degree of
Doctor of Philosophy
in the Department of Oceanography

July 2010

Supervisors

Professor Chris Reason

Department of Oceanography, University of Cape Town, South Africa.

Dr Pierrick Penven

Institut de Recherche pour le Développement, Laboratoire de Physique des Océans (UMR 6523, CNRS, IFREMER, IRD, UBO), Centre IRD de Bretagne, France.

Declaration

I declare that this thesis is my own unaided work, both in concept and execution, and that apart from the normal guidance from my supervisors, I have received no assistance except as acknowledged. I declare that neither the substance nor any part of this thesis has been submitted in the past, or is being, or is to be submitted for a degree at this university or at any other university.

This work has been supported by resources and funding from the following institutions:

- The National Research Foundation (NRF)¹.
- The Postgraduate Funding Office at the University of Cape Town.
- The NRF/French National Centre for Scientific Research (CNRS)/French Embassy Doctoral Support Programme.
- The Centre for High Performance Computing (CHPC)

¹A list of all the acronyms used in this thesis is provided in Appendix E

Abstract

Capitalising on the vast knowledge gained in the tropical Pacific, tropical Atlantic variability is viewed from an energetics perspective and contrasted against the Pacific at both seasonal and inter-annual timescales. The character of remotely forced thermocline depth variability is evaluated using available potential energy which succinctly quantifies the response of the upper equatorial ocean to large scale wind forcing. To facilitate the Atlantic-Pacific comparison, available potential energy values are derived from the temperature and salinity fields of two global ocean models. The presence of a strong sea surface temperature-available potential energy relationship in the Atlantic, like that observed in the Pacific, suggests that anomalous vertical advection in the central-eastern basin due to remotely forced thermocline depth changes plays an equally important role in determining sea surface temperature variability in the Atlantic. However, unlike in the Pacific, the largest changes in available potential energy occur seasonally rather than inter-annually in the Atlantic.

Focusing in on the Atlantic, the Regional Ocean Modelling System is used to simulate conditions in the tropical Atlantic from 1980-2004. This simulation is conducted to facilitate the calculation of surface layer heat budgets as well as to diagnose the processes determining the evolution of available potential energy. The in-phase relationship between seasonal available potential energy and sea surface temperature fluctuations in the Atlantic is shown to be due to the fact that the large annual increase in the intensity of cooling due to vertical advection is not purely a locally forced phenomena. Instead, it is largely due to the horizontal redistribution of warm surface waters in response to large scale changes in seasonal wind forcing, a process that only plays a significant role inter-annually in the Pacific.

Furthermore, this seasonal upper ocean response appears to involve more than a passive response to seasonal wind forcing. The relationship between seasonal sea surface temperature, available potential energy, and zonal wind stress changes supports the notion that the seasonally excited growth of asymmetric conditions about the equator as the cold tongue develops is accompanied by a seasonally excited Bjerknes feedback. Seasonal available potential energy changes in the Atlantic are seen to be primarily forced by seasonal changes in the work done by the wind on the ocean. Distinctions between the seasonal sea surface temperature signature in the central-eastern Pacific and Atlantic are ascribed to the presence of this seasonally excited Bjerknes feedback as well as a delayed negative feedback mechanism in the form of ocean memory. In the Pacific, inter-annual available potential energy anomalies are grown by the Bjerknes feedback mechanism until the delayed response of the ocean affects the ability of the wind to do work on the ocean. The results of the energetics analysis conducted in this thesis indicate that a similar process, referred to as the thermocline mode, is seasonally excited in the Atlantic between April and October after which point seasonal forcing regains control.

As opposed to a dynamically distinct natural mode of variability operating on inter-annual timescales, seasonality in the occurrence of anomalous central-eastern basin sea surface temperature events correlated with thermocline depth and zonal wind anomalies suggests that inter-annual variability associated with the Atlantic's zonal mode may be best understood as modulations of the seasonally excited thermocline mode. The ocean memory mechanism associated with the zonal mode appears to operate on much shorter time scales than that associated with the El Niño Southern Oscillation, largely being associated with inter-annual modulations of the seasonally active delayed negative feedback response. Differences between the El Niño Southern Oscillation and the zonal mode can then be accounted for in terms of these distinctions.

Acknowledgements

I would like to thank my supervisors Chris Reason and Pierrick Penven for their guidance and kindness. I am grateful to Chris for all the support and direction he has given me over the past years, from my undergraduate years when he first offered me insight by recommending additional reading to the valuable advice he has given me surrounding this thesis. My thanks to Pierrick for all the knowledge he has imparted upon me through out this thesis and for hosting my research visits at the Laboratoire de Physique des Oceans (LPO) in France. I am particularly grateful to Pierrick for providing me with the supervision and confidence needed to tackle some of the more challenging aspects of this thesis.

I would also like to express my gratitude to George Philander for all the insightful and inspiring discussions that we shared in the three years since I first met him. The constructive feedback and advice George has given me at various stages has added clarity and helped in shaping this thesis.

I would like to take this opportunity to thank the NRF, the CNRS, the French Embassy and LPO for providing me with the opportunity to spend two research visits each of three months based at Ifremer under Pierrick's supervision. These research visits, during which I gained a large amount of insight into my research, were truly enriching experiences. I was exposed to a number of scientists with whom I was able to discuss various aspects of my research and believe that the quality of my PhD research was greatly improved as a direct result of these visits. Thanks to Jen and Issufo for the companionship on Rue Turenne and for making my stays in France so enjoyable.

To all the academic staff and postgraduate students in the department of oceanography thank you for fostering such a dynamic learning environment, especially to Bjorn for offering to proof read this thesis and Seb for the lively lunch breaks, interesting discussions and knowing my idiosyncrasies so well. I am grateful to all my friends, specifically Mary and Jess, for all the words of encouragement. To my family, Martin, Johanne and Patrick, thank you for providing me with such a strong foundation upon which to meet the world. Finally I would like to thank David, not only for all the love and support he has given throughout my undergraduate and postgraduate journey but also for showing me how to tackle various problems throughout the years and as a result to have confidence in my problem solving abilities.

Contents

Supervisors	i
Declaration	i
Abstract	ii
Acknowledgements	iii
List of Figures	vii
1 Introduction	1
1.1 Background	1
1.2 Objectives	5
2 Tropical Atlantic Climate Variability	12
2.1 Mean State	12
2.2 Seasonal Variability	16
2.3 Inter-annual Variability	23
2.4 Summary	29
3 Equatorial Variability in the Atlantic versus the Pacific	30
3.1 Introduction	30
3.2 Data and Methodology	32
3.2.1 Global Ocean Models	32
3.2.2 APE Estimation	35
3.3 Equatorial Atlantic vs Equatorial Pacific Analysis	43
3.3.1 Equatorial Atlantic vs Equatorial Pacific APE - SST Relationship	43
3.3.2 Equatorial Pacific APE - SST relationship on Seasonal and Inter-annual time scales	48

3.3.3	Equatorial Atlantic APE - SST relationship on Seasonal and Inter-annual time scales	51
3.3.4	Equatorial Atlantic vs Equatorial Pacific, APE - SST - Wind relationships	55
3.4	Conclusions	63
4	Regional Ocean Modelling System - Tropical Atlantic Simulation	64
4.1	The Regional Ocean Modelling System	64
4.2	Tropical Atlantic Configuration	65
4.3	Validation	66
4.4	APE Evaluation	71
4.5	Summary	72
5	Tropical Atlantic Seasonal Cycle	74
5.1	Introduction	74
5.2	Surface Layer Heat Budgets	75
5.2.1	Introduction	75
5.2.2	Methodology - Surface Layer Heat Budget Calculation and Estimating Ekman Vertical Velocities	77
5.2.2.1	Surface Layer Heat Budget Calculation	77
5.2.2.2	Estimating Ekman Vertical Velocities	77
5.2.3	Results	78
5.2.3.1	North-South Asymmetry	83
5.2.4	Conclusions	87
5.3	Seasonal Changes in Equatorial and Tropical Atlantic APE	88
5.3.1	Introduction	88
5.3.2	Methodology - Evaluating the APE Evolution Equation	88
5.3.2.1	Potential Energy Evolution Equation	89
5.3.2.2	Background Potential Energy Evolution Equation	91
5.3.2.3	Available Potential Energy Evolution Equation	93
5.3.2.4	Kinetic Energy Evolution Equation	98
5.3.2.5	Evaluating Climatological APE	98
5.3.3	Results	101
5.3.4	Conclusions	120

6	Inter-annual variability in the Equatorial Atlantic	122
6.1	Introduction	122
6.2	Anomalous APE-SST-Wind Relationship	124
6.3	Inter-annual Anomalies in Equatorial and Tropical Atlantic APE	132
6.4	Inter-annual modulation of the seasonally excited thermocline mode	139
6.4.1	Composite Analysis and Individual Events	147
6.4.2	Anomalous BPE	158
6.5	Conclusions	165
7	Summary of Conclusions and Outlook	166
	References	172
A	ORCA Figures	I
B	Surface Layer Heat Budget	IX
C	Ekman Transport	XII
D	KE Evolution Equation	XIV
E	List of Acronyms	XIX

List of Figures

- 1.1 Seasonal SST, wind stress, and rainfall fluctuations in the (a) western Pacific Ocean - 165°E, (b) central Pacific Ocean - 120°W, (c) eastern Pacific Ocean - 90°W, (d) western Atlantic Ocean - 30°W, (e) central Atlantic Ocean - 20°W, (f) eastern Atlantic Ocean - 0°W, (g) western Indian Ocean - 55°E, (h) central Indian Ocean - 70°E, and (i) eastern Indian Ocean - 90°E. Climatological SST, wind stress and rainfall values shown here were derived from the Comprehensive Ocean-Atmosphere Data Set (COADS) (DaSilva et al., 1994). The colour filled black contours represent SST values in °C as indicated by the corresponding colourbar. The white contours represent precipitation with a contour interval of 0.2cmday⁻¹. The black vectors represent seasonal changes in the zonal and meridional wind stress components. 6
- 1.2 Based on 1° resolution, Hadley OI SST data (Rayner et al., 2003) spanning 1958-2004, this figure shows the standard deviation of monthly, (a) total SST variations in the Pacific, (b) total SST variations in the Atlantic, (c) seasonal SST fluctuations in the Pacific, (d) seasonal SST fluctuations in the Atlantic, (e) inter-annual SST anomalies in the Pacific, and (f) inter-annual SST anomalies in the Atlantic. A contour interval of 0.25°C has been used. Regions of >1.5°C are shaded in figures a-d and >0.5°C in figures e and f. The Niño 3.4 (5°N-5°S 170°W-120°W) and Niño 3 (5°N-5°S 150°W-90°W) and the Atlantic Atl3 (3°S-3°N 20°W-0°E) regions have been demarcated. 7
- 1.3 A comparison between seasonal variability in Niño 3.4 (5°N-5°S 170°W-120°W) SST, Niño 3 (5°N-5°S 150°W-90°W) SST, and Atl3 (3°S-3°N 20°W-0°E) SST. Created using monthly Hadley OI SST data (Rayner et al., 2003). SST values are given in °C. 8
- 1.4 A comparison between variability in total versus climatological (a) Niño 3.4 (5°N-5°S 170°W-120°W) SST, (b) Niño 3 (5°N-5°S 150°W-90°W) SST, and (c) Atl3 (3°S-3°N 20°W-0°E) SST values. Created using monthly Hadley OI SST data (Rayner et al., 2003). SST values are given in °C and have been detrended. 9

- 2.1 The temporal mean of several atmospheric and oceanic variables within the tropical Atlantic: (a) mean tropical Atlantic SST, (b) mean wind stress with regions of negative wind stress curl shaded in grey and regions of positive curl in white, (c) mean precipitation, (d) the Sverdrup stream function based on the mean wind stress field depicted in (b), (e) mean dynamic topography from the CNES-CLS09 mean dynamic topography product (Rio et al., 2009) (www.aviso.oceanobs.com), and (f) mean surface currents derived from the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). The mean SST field, wind stress field and rainfall field were derived from the Comprehensive Ocean-Atmosphere Data Set (COADS) (DaSilva et al., 1994). In (d), the stream functions have been plotted with a contour interval of 2 Sverdrup. Counter-clockwise circulation is indicated by the solid contours and clockwise circulation by the dashed contours. 13
- 2.2 A Schematic of the mean features of the tropical Atlantic current system taken from Schouten et al. (2005). Surface currents indicated by the solid lines include the North Equatorial Current (NEC), North Equatorial Counter Current (NECC), Guinea Current (GC), Angola Current (AC), South Equatorial Current (SEC), North Brazil Current (NBC), Brazil Current (BC). Subsurface currents indicated by the dashed lines include the Equatorial Under Current (EUC), North and South Equatorial Under Currents ((N/S)EUC) and North Brazil Under Current (NBUc). Also shown is the position of the Angola–Benguela Front (ABF) and two NBC rings. 15
- 2.3 This figure depicts the seasonal cycle of (a) meridional wind stress and (b) zonal wind stress along the equator (3°S–3°N) in the Atlantic. COADS data (DaSilva et al., 1994) and a contour interval of 0.005Nm^{-2} has been used. The thick black line represents the 0Nm^{-2} contour line. 16
- 2.4 Taken from Schouten et al. (2005), this figure shows the standard deviation in cm of tropical Atlantic SSH measured between 1992 and 2001. Hatched regions indicate where the seasonal cycle contributes strongly ($> 50\%$) to the total variability. 17
- 2.5 Seasonal zonal surface current perturbations from the annual mean along the equator (1°S–1°N). This figure is based on the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). A contour interval of 0.1ms^{-1} has been used. The thick black line represents the 0ms^{-1} contour line. . . . 18
- 2.6 Taken from Ding et al. (2009), this figure depicts the seasonal cycle of SSH anomalies with respect to the annual mean: (a) at the equator, (b) at 4°N, and (c) at 4°S. Aviso (www.aviso.oceanobs.com) SSH data has been used. A contour interval of 1cm has been used in (a) and (c) while the -12,-7,-5,-3,-1,0,1,3,5,7,12 contours are shown in (b). 20
- 2.7 A monthly climatology derived from Pilot Research Moored Array in the Tropical Atlantic (PIRATA) (Servain et al., 1998) temperature profiles at three sites on the equator (a) 0°N35°W, (b) 0°N10°W and (c) 0°N0°E. 22

- 2.8 Taken from Kushnir et al. (2006), this figure depicts the dominant pattern of surface ocean-atmosphere variability in the tropical Atlantic region during (a) boreal spring (March-April) and (b) boreal summer (June-August). The black contours show the first Empirical Orthogonal Function (EOF) of the regional rainfall anomalies for each period from Global Precipitation Climatology Project data (1979-2001) (explaining 33% and 23% of the variance respectively). A contour interval of 0.5 mm day^{-1} has been used, positive contours are solid, negative contours are dashed, and the 0 contour is omitted. The coloured shading depicts the associated March-April and June-August SST anomalies (SST anomalies regressed against the rainfall EOF's principal component time series), with units given on the scale below in $^{\circ}\text{C}$, plus white contours every 0.2°C for clarity. Arrows depict the associated March-April and June-August surface wind anomalies (wind anomalies regressed against the rainfall EOF's principal component time series), with units in ms^{-1} (see arrow scale below each figure). 24
- 3.1 A comparison between inter-annual anomalies in (a) Niño 3.4, (b) Niño 3, and (c) Atl3 SST, based on observed Hadley OI SST data (Rayner et al., 2003) and simulated SODA and ORCA SST data. SST anomalies are shown in $^{\circ}\text{C}$. The correlation between observed and simulated anomalies in each SST index is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). 33
- 3.2 A comparison between climatological (a) Niño 3.4, (b) Niño 3, and (c) Atl3 SST, based on observed Hadley OI SST data (Rayner et al., 2003) and simulated SODA and ORCA SST data. SST values are shown in $^{\circ}\text{C}$ 34
- 3.3 A comparison between equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) climatological APE values obtain using method 1, given by Equation 3.3, and method 2, given by Equation 3.2. (a) APE values have been calculated using SODA data. (b) APE values have been calculated using ORCA data. 37
- 3.4 Zonal changes in the climatological depth of the 20°C isotherm along the equator (3°S - 3°N) for February, April, August, and September, based on (a) SODA data, and (b) ORCA data. 37
- 3.5 Cross-sections of temperature along the equator (3°S - 3°N), (a) April and (b) August climatology based on SODA data, (c) April and (d) August climatology based on ORCA data. . 38
- 3.6 Variability in the vertical density profile estimated by horizontally averaging the monthly mean density field within the chosen domain: (a) the equatorial Pacific (5°N - 5°S 130°E - 85°W), (b) the equatorial Atlantic (3°S - 3°N 60°W - 15°E), (c) the tropical Pacific (8°N - 8°S 130°E - 85°W), (d) the tropical Atlantic (8°S - 8°N 60°W - 15°E), (e) the extended tropical Pacific (15°N - 15°S 130°E - 85°W), and (f) the extended tropical Atlantic (15°S - 15°N 60°W - 15°E). Seasonal variability in the vertical density profile is overlayed on the total spread of vertical density profiles between 1958-2004 shown in cyan. Depth (m) is represented on the y-axis and density (kgm^{-3}) on the x-axis. Monthly density fields are derived from the monthly SODA temperature and salinity fields. 40
- 3.7 Seasonal changes in the average temperature of the upper equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) and the upper tropical Atlantic (8°S - 8°N 60°W - 15°E 0-400m) based on the WOA05 (Locarnini et al., 2006) climatological temperature field. 41

3.8	Seasonal changes in the average density of the upper equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) and the upper tropical Atlantic (8°S-8°N 60°W-15°E 0-400m). (a) Climatological density values have been calculated based on simulated temperature and salinity fields from the SODA model. (b) Climatological density values have been calculated based on simulated temperature and salinity fields from the ORCA model. (c) Climatological density values have been calculated based on WOA05 climatological temperature and salinity fields.	42
3.9	Seasonal changes in equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE obtained using Equation 3.3, APE and BPE obtained using Equation 3.2, as well as seasonal changes in the Mass of this domain. Mass, BPE and APE values have been normalised. Values shown here have been calculated using SODA data, the same figure only based on ORCA data, Figure A.2, can be found in Appendix A.	43
3.10	A comparison between total Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE variability. SST and APE values were calculated using SODA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.3, can be found in Appendix A.	44
3.11	A comparison between total Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE variability. SST and APE values were calculated using SODA data. The correlation between the Atl3 SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.4, can be found in Appendix A.	44
3.12	A comparison between total and climatological equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE values. APE values were calculated following Equation 3.2 using SODA data and have been detrended. The corresponding figure based on ORCA data, Figure A.5, can be found in Appendix A.	45
3.13	A comparison between total and climatological equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE values. APE values were calculated following Equation 3.2 using SODA data and have been detrended. The corresponding figure based on ORCA data, Figure A.6, can be found in Appendix A.	45
3.14	(a) Equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE versus tropical Pacific (15°N-15°S 130°E-85°W 0-400m) APE. (b) Equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE versus tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE. APE values were calculated using SODA data following Equation 3.2. The corresponding figure based on ORCA data, Figure A.7, can be found in Appendix A.	47

- 3.15 A comparison between inter-annual Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE anomalies. SST and APE anomalies were calculated using SODA data. The correlation between anomalies in each SST index and APE (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.8, can be found in Appendix A. 48
- 3.16 A comparison between climatological Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE. SST and APE values were calculated using SODA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.9, can be found in Appendix A. 49
- 3.17 This figure depicts the seasonal cycle of (a) meridional wind stress and (b) zonal wind stress along the equator (3°S-3°N) in the Pacific. COADS data (DaSilva et al., 1994) and a contour interval of 0.005Nm^{-2} has been used. The thick black line represents the 0Nm^{-2} contour line. 50
- 3.18 A comparison between climatological Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE. SST and APE values were calculated using SODA data. The correlation between the Atl3 SST index and equatorial Atlantic APE values (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.10, can be found in Appendix A. 51
- 3.19 A comparison between inter-annual Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE anomalies. SST and APE anomalies were calculated using SODA data. The correlation between SST and APE anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.11, can be found in Appendix A. 54
- 3.20 The seasonal dependence in correlations between Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE anomalies. SST and APE anomalies were calculated using SODA data. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.12, can be found in Appendix A. 54

- 3.21 Spatial maps, based on SODA re-analysis data, of the correlation between monthly zonal wind stress values and monthly Niño 3.4, Niño 3 and Atl3 time-series spanning 1958-2004. (a) Climatological zonal wind stress and climatological Niño 3.4 SST. (b) Anomalous zonal wind stress and anomalous Niño 3.4 SST. (c) Climatological zonal wind stress and climatological Niño 3 SST. (d) Anomalous zonal wind stress and anomalous Niño 3 SST. (e) Climatological zonal wind stress and climatological Atl3 SST. (f) Anomalous zonal wind stress and anomalous Atl3 SST. Correlation values plotted are those significance at the 95% level. 56
- 3.22 Spatial maps, based on SODA re-analysis data, of the correlation between monthly zonal wind stress values and monthly equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE spanning 1958-2004. (a) Climatological zonal wind stress and climatological APE, and (b) anomalous zonal wind stress and anomalous APE. Correlation values plotted are those significance at the 95% level. APE values were calculated following Equation 3.2. 57
- 3.23 A comparison between variability in total versus climatological, (a) western-central basin zonal wind stress in the Pacific (5°N-5°S 160°E-160°W), and (b) western basin zonal wind stress in the Atlantic (3°S-3°N 40°W-20°W). Created using SODA wind stress forcing data. Wind stress values are given in Nm^{-2} and have been detrended. 58
- 3.24 (a) A comparison between inter-annual Atl3 SST and western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies. The correlation between anomalies (r) is given together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). (b) The seasonal dependence in correlations between Atl3 SST and zonal wind stress anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST and zonal wind stress anomalies were calculated using SODA data and have been normalised and detrended. 60
- 3.25 (a) A comparison between inter-annual equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE and western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies. The correlation between anomalies (r) is given together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). (b) Seasonal dependence of correlation between APE and zonal wind stress anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. APE and zonal wind stress anomalies were calculated using SODA data and have been normalised and detrended. APE values were calculated following Equation 3.2. 61
- 3.26 Time lag correlations between anomalous (a) Atl3 SST and equatorial Atlantic APE (3°S-3°N 60°W-15°E 0-400m), (b) Atl3 SST and western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress, and (c) equatorial Atlantic APE and western equatorial Atlantic zonal wind stress. SST, APE and zonal wind stress anomalies were derived from 1958-2004 SODA data and have been detrended. APE values were calculated following Equation 3.2. 62

- 4.1 The standard deviation of monthly, (a) total TAtl-ROMS SST variations, (b) total Hadley OI SST variations, (c) seasonal TAtl-ROMS SST variations, (d) seasonal Hadley OI SST variations, (e) inter-annual TAtl-ROMS SST anomalies, and (f) inter-annual Hadley OI SST anomalies from 1980-2004. A contour interval of 0.25°C has been used, regions of $>1.5^{\circ}\text{C}$ are shaded in figures a-d and $>0.5^{\circ}\text{C}$ in figures e and f. The Atl3 (3°S - 3°N 20°W - 0°E) region has been demarcated. 67
- 4.2 A comparison between the Atl3 (3°S - 3°N 20°W - 0°E) index created using Hadley OI SST data versus ROMS-TAtl SST data, (a) climatological Atl3 SST and (b) inter-annual Atl3 SST anomalies. The correlation between inter-annual anomalies in each SST index (r) is give on figure b, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST values are given in $^{\circ}\text{C}$ and have been detrended in figure b. 68
- 4.3 Climatological cross-sections in (a) April and (b) August of temperature ($^{\circ}\text{C}$) along the equator (1°N - 1°S) based on WOA05 data (Locarnini et al., 2006) (solid black and colour contours) and overlaid with data from the ROMS-TAtl simulation (dashed black contours). 69
- 4.4 A monthly climatology of temperature profiles for three sites on the equator (a) 0°N , 35°W , (b) 0°N , 10°W and (c) 0°N , 0°E . (Black contours) PIRATA mooring data (Servain et al., 1998) (Red contours) ROMS-TAtl simulation. 69
- 4.5 A comparison between (a) ROMS-TAtl, and (b) observed climatological SSH anomalies north of the equator (3°N - 5°N); (c) ROMS-TAtl, and (d) observed climatological SSH anomalies along the equator (1°S - 1°N); (e) ROMS-TAtl and (f) observed climatological SSH anomalies south of the equator (3°S - 5°S). Observed climatological SSH anomalies where derived from $1/3^{\circ}$ Aviso data (www.aviso.oceanobs.com) spanning January 1993-December 2004 and therefore the ROMS-TAtl SSH climatology shown here was derived from simulated SSH over the corresponding period. 70
- 4.6 A comparison between (a) ROMS-TAtl, (b) observed, climatological, equatorial Atlantic, zonal surface currents averaged between 4°N - 4°S . Observed zonal surface currents plotted in b where derived from the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). This surface current climatology product is based on surface drifter observations spanning 28 October 1990 - 28 February 2004 and therefore the ROMS-TAtl zonal surface current climatology shown here was derived from simulated currents over the corresponding period. 71
- 4.7 Meridional sections of the mean simulated zonal velocity field (given in ms^{-1}) at (a) 35°W , (b) 23°W and (c) 10°W 72
- 4.8 A comparison between equatorial Atlantic (3°S - 3°N 60°W - 15°E) APE values calculated based on density values derived from ROMS-TAtl temperature and salinity output using a non-linear equation of state, and APE values calculated based on density values derived using a linear equation of state. APE values where evaluated following Equation 3.2 and have had the mean subtracted. 73

5.1	(a) Seasonal variations in SST, the average temperature of the Surface Mixed Layer (SML) (varying depth), and the average temperature of the SubSurface Layer (SSL) (fixed depth =70m) for the Atl3 region. Temperature values are given in °C. (b-c) Climatology of terms ($^{\circ}\text{Cs}^{-1}$) contributing to seasonal variations in the mean temperature of (b) the surface mixed layer, and (c) the subsurface layer. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.	79
5.2	Seasonal variations in the average depth of the surface mixed layer for the Atl3 region given in meters. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.	80
5.3	A comparison between climatological Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE. SST and APE values were calculated using ROMS-TAtl data and have been normalised. The correlation between the Atl3 SST and equatorial Atlantic APE (r) is given together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). APE values were calculated following Equation 3.2.	81
5.4	Seasonal variations in zonal and meridional ROM-TAtl wind stress forcing as well as local Ekman vertical velocities for the Atl3 region. Wind stress and Ekman vertical velocities values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.	82
5.5	Seasonal variations in the intensity of the vertical advection term for the Atl3 SubSurface Layer (SSL) heat budget compared with seasonal variation in Ekman vertical velocities for the Atl3 region and the rate of change of equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE. Note that, vertical advection values have been inverted to facilitate the comparison. Ekman vertical velocities, APE rate of change and vertical advection values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.	82
5.6	(a) Seasonal variations in SST ($^{\circ}\text{C}$) for the southern (0°N - 3°S 60°W - 15°E) and northern (3°N - 0°S 60°W - 15°E) Atl3 regions. (b) Same as (a) only SST values have been normalised for a closer comparison. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.	83
5.7	(a-b) Seasonal variations in zonal and meridional wind stress as well as local Ekman vertical velocities for (a) the southern Atl3 region, and (b) the northern Atl3 region. Wind stress and Ekman vertical velocities values have been normalised. (c-f) Climatology of terms ($^{\circ}\text{Cs}^{-1}$) contributing to seasonal variations in the mean temperature of, (c) the southern Atl3 region surface mixed layer, (d) the northern Atl3 region surface mixed layer, (e) the southern Atl3 region subsurface layer, and (f) the northern Atl3 region subsurface layer. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. . .	85

- 5.8 Seasonal variations in the intensity of the vertical advection term for (a) the southern Atl3 region and (b) the northern Atl3 region, subsurface layer heat budget, compared with seasonal variation in Ekman vertical velocities for (a) the southern Atl3 region and (b) the northern Atl3 region, and the rate of change of equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE. Note that, vertical advection values have been inverted to facilitate the comparison. Ekman vertical velocities, APE rate of change and vertical advection values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 86
- 5.9 The balance of terms for the PE evolution equation in the form given by Equation 5.14, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). dPE/dt represents the actual rate of change of PE where as *derived* dPE/dt represents the rate of change of the PE estimated by balancing the terms within Equation 5.14. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. . . 91
- 5.10 The balance of terms for the BPE evolution equation in the form given by Equation 5.19, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). $dBPE/dt$ represents the actual rate of change of BPE where as *derived* $dBPE/dt$ represents the rate of change of the BPE estimated by balancing the terms within Equation 5.19. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 93
- 5.11 The balance of terms for the APE evolution equation in the form given by Equation 5.20, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). $dAPE/dt$ represents the actual rate of change of APE where as *derived* $dAPE/dt$ represents the rate of change of the APE estimated by balancing the terms within Equation 5.20. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 94
- 5.12 Balance of terms for the APE evolution equation in the final form given by Equation 5.29, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). $dAPE/dt$ represents the actual rate of change of APE where as *derived* $dAPE/dt$ represents the rate of change of the APE estimated by balancing the terms within Equation 5.29. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 97
- 5.13 A comparison between climatological APE values and the two components which make up climatological APE values, cl_m and cl_p for (a) the equatorial Atlantic and (b) the tropical Atlantic. 100
- 5.14 The climatological balance of terms (Js^{-1}) for the APE evolution equation (Equation 5.29), assessed over (a) the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m), and (b) the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 102
- 5.15 The climatological balance of terms (Js^{-1}) for the KE evolution equation (Equation 5.30), assessed over (a) the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m), and (b) the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). A 14 day running mean has been applied to the time series to smooth out the high frequency variability. Note $\Phi_{apk_{cl}}$ values in this KE evolution equation balance are the inverse of $\Phi_{apk_{cl}}$ values given for the APE evolution equation (Figure 5.14). 103

- 5.16 A comparison between seasonal fluctuations in buoyancy power ($\Phi_{apk_{cl}}$) and wind power ($\Phi_{ww_{cl}}$) for (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m), and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). Buoyancy power ($\Phi_{apk_{cl}}$) and wind power ($\Phi_{ww_{cl}}$) values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 104
- 5.17 A comparison between seasonal fluctuations in the buoyancy power term ($\Phi_{apk_{cl}}$); the combined effect of physical processes operating within the domain, $\Phi_{ww_{cl}} + \Phi_{ss_{cl}}$; and the combined effect of terms associated with boundary processes, $\Phi_{ka_{cl}} + \Phi_{pw_{cl}}$ for (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). Values are given in Js^{-1} . A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 105
- 5.18 Equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) climatological APE and BPE changes versus tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) climatological APE and BPE changes. APE and BPE values were calculated based on output from the ROMS-TAtl simulation (1980-2004) following Equation 3.2. Values have been normalised to facilitate a better comparison. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 107
- 5.19 A comparison between seasonal fluctuations in the wind power term ($\Phi_{ww_{cl}}$), and its zonal ($\Phi_{ww_{cl}^x}$) and meridional ($\Phi_{ww_{cl}^y}$) components for the tropical Atlantic domain (8°S-8°N 60°W-15°E). Values are given in Js^{-1} . A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 108
- 5.20 A comparison between seasonal fluctuations in the zonal component of the wind power term from its temporal mean ($\iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS - \iint_{z=0} \overline{u_{cl}} \overline{\tau_{s_{cl}}^x} dS$), the mean perturbation terms, $\Phi_{ww_{cl}}^{mup\tau}$ and $\Phi_{ww_{cl}}^{m\tau pu}$ as well as the perturbation term, $\Phi_{ww_{cl}}^{pp}$ for the tropical Atlantic domain (8°S-8°N 60°W-15°E). Values are given in Js^{-1} . A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 109
- 5.21 A comparison between seasonal fluctuations in the mean perturbation term, $\Phi_{ww_{cl}}^{mup\tau}$, seasonal fluctuations in zonal wind stress integrated over the tropical Atlantic domain (8°S-8°N 60°W-15°E), and seasonal fluctuations in zonal wind stress integrated over 8°S-2°N 50°W-0°E. Values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. 110
- 5.22 The mean zonal surface velocity field (u) for the tropical Atlantic domain, based on output from the ROMS-TAtl simulation (1980-2004). Zonal surface velocities are given ms^{-1} . . . 110
- 5.23 Seasonal zonal surface current perturbations from the annual mean averaged between 4°S-4°N. This figure is based on output from the ROMS-TAtl simulation (1980-2004). Zonal surface velocity perturbations are given ms^{-1} 111
- 5.24 Monthly maps illustrating the spatial structure in the climatological wind power field (Wm^{-2}) over the tropical Atlantic domain (8°S-8°N 60°W-15°E). 113

5.25	Monthly maps illustrating the spatial structure in climatological zonal wind stress perturbations from the annual mean (Nm^{-2}) over the tropical Atlantic domain (8°S - 8°N 60°W - 15°E) between April and September.	114
5.26	Monthly maps illustrating the spatial structure in climatological zonal surface current perturbations from the annual mean (ms^{-1}) over the tropical Atlantic domain (8°S - 8°N 60°W - 15°E) between April and September. Figures in the left column (a, c, e, g, i and k) are based on output from the ROMS-TAtl simulation (1980-2004), while figures in the right column (b, d, f, h, j and l) are based on the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html).	115
5.27	(a) APE verse wind power phase diagram associated with the seasonal cycle in the tropical Atlantic domain (8°S - 8°N 60°W - 15°E). (b) A comparison between seasonal fluctuations in APE (APE_{cl}), the APE tendency term ($d\text{APE}_{cl}/dt$) and the wind power term ($\Phi_{ww_{cl}}$) for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E). Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability.	118
5.28	The climatological balance of terms (Js^{-1}) for the BPE evolution equation (Equation 5.19), assessed over (a) the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m) and (b) the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). A 14 day running mean has been applied to the time series to smooth out the high frequency variability.	119
6.1	(a) A comparison between inter-annual Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised. (b) The seasonal dependence in correlations between Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST and APE anomalies were calculated based on ROMS-TAtl output. APE values were calculated following Equation 3.2.	125
6.2	(a) A comparison between ROMS-TAtl, monthly, inter-annual anomalies in SST, the average temperature of the Surface Mixed Layer (SML) (varying depth) and the average temperature of the SubSurface Layer (SSL) (fixed depth =70m) for the Atl3 region. The correlation between variables (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). (b) The seasonal dependence in the correlation between inter-annual anomalies in SST and the average temperature of the subsurface layer for the Atl3 region. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey.	127
6.3	Climatological wind stress values for the Atl3 region. Climatological wind stress values were derived from the ROMS-TAtl simulation's wind stress forcing.	128

- 6.4 (a) A spatial map based on ROMS-TAtl data of the correlation between monthly anomalies in zonal wind stress values and Atl3 SST spanning 1980-2004. Correlation values plotted are those significance at the 95% level. (b) The seasonal dependence in correlation between ROMS-TAtl western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies and Atl3 SST anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. 129
- 6.5 (a) A spatial map based on ROMS-TAtl data of the correlation between monthly anomalies in zonal wind stress values and equatorial APE (3°S-3°N 60°W-15°E 0-400m) spanning 1980-2004. Correlation values plotted are those significance at the 95% level. (b) The seasonal dependence in correlation between ROMS-TAtl western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies and equatorial APE anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. 130
- 6.6 Time lag correlations between anomalous (a) Atl3 SST and equatorial Atlantic APE (3°S-3°N 60°W-15°E 0-400m), (b) Atl3 SST and western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress, and (c) equatorial Atlantic APE and western equatorial Atlantic zonal wind stress. SST, APE and zonal wind stress anomalies were derived from 1980-2004 ROMS-TAtl data. APE values were calculated following Equation 3.2. 131
- 6.7 A comparison between monthly, inter-annual anomalies in the buoyancy power term ($\Phi_{apk_{ano}}$ in Js^{-1}), and APE evolution ($d\text{APE}_{ano}/dt$ in Js^{-1}), assessed over (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). 133
- 6.8 A comparison between monthly inter-annual anomalies in the buoyancy power term ($\Phi_{apk_{ano}}$), and wind power ($\Phi_{ww_{ano}}$) term, for (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). Anomalous buoyancy power ($\Phi_{apk_{ano}}$) and wind power ($\Phi_{ww_{ano}}$) values have been normalised. . . . 135
- 6.9 (a) A comparison between inter-annual equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE anomalies and tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE anomalies. (b) Time lag correlations between equatorial and tropical Atlantic APE anomalies. (c) A comparison between inter-annual Atl3 SST and tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE anomalies. SST and APE anomalies were calculated based on ROMS-TAtl output. SST and APE values have been normalised. APE values were calculated following Equation 3.2. In a and c, the correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). 136

- 6.10 A comparison between monthly, inter-annual anomalies in APE evolution ($dAPE_{ano}/dt$) and wind power ($\Phi_{ww_{ano}}$) for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). Anomalous APE evolution ($dAPE_{ano}/dt$) and wind power ($\Phi_{ww_{ano}}$) values have been normalised. (b) The seasonal dependence in correlation between monthly inter-annual anomalies in APE evolution ($dAPE_{ano}/dt$) and wind power ($\Phi_{ww_{ano}}$) for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. 137
- 6.11 (a) A comparison between monthly, inter-annual anomalies in the wind power term ($\Phi_{ww_{ano}}$), and its zonal ($\Phi_{ww_{ano}^x}$) and meridional ($\Phi_{ww_{ano}^y}$) components for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). Values are given in Js^{-1} . (b) A comparison between monthly, inter-annual anomalies in the zonal component of the wind power term ($\Phi_{ww_{ano}^x}$), and APE evolution ($dAPE_{ano}/dt$) assessed over the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). Values have been normalised. 138
- 6.12 (a) Time lag correlations between inter-annual Atl3 SST anomalies and tropical Atlantic wind power anomalies ($\Phi_{ww_{ano}}$). (b) The seasonal dependence in correlations between monthly inter-annual Atl3 SST anomalies and tropical Atlantic wind power anomalies ($\Phi_{ww_{ano}}$). Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies and wind power anomalies were calculated based on ROMS-TAtl output and a three month running mean has been applied to the time series. 140
- 6.13 The seasonal dependence in monthly Atl3 SST auto correlation. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies were calculated based on ROMS-TAtl output. 142
- 6.14 A comparison between monthly anomalies in the zonal component of the wind power term, $\Phi_{ww_{ano}^x}$; the mean perturbation terms, $\Phi_{ww_{ano}}^{mup\tau}$ and $\Phi_{ww_{ano}}^{m\tau pu}$; and the perturbation term, $\Phi_{ww_{ano}}^{pp}$ for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). Values are given in Js^{-1} 143
- 6.15 (a) Time lag correlations between inter-annual Atl3 SST anomalies and the mean perturbation term $\Phi_{ww_{ano}}^{mup\tau}$. (b) The seasonal dependence in correlations between monthly inter-annual Atl3 SST anomalies and the mean perturbation term $\Phi_{ww_{ano}}^{mup\tau}$. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies and $\Phi_{ww_{ano}}^{mup\tau}$ were calculated based on ROMS-TAtl output and a three month running mean has been applied to the time series. 144

- 6.16 (a) Time lag correlations between inter-annual Atl3 SST anomalies and the mean perturbation term $\Phi_{wwano}^{m\tau pu}$. (b) The seasonal dependence in correlations between monthly inter-annual Atl3 SST anomalies and the mean perturbation term $\Phi_{wwano}^{m\tau pu}$. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies and $\Phi_{wwano}^{m\tau pu}$ were calculated based on ROMS-TAtl output and a three month running mean has been applied to the time series. 145
- 6.17 Monthly anomalies in ROMS-TAtl Atl3 SST plotted in relation to one standard deviation. Periods where simulated Atl3 SST anomalies correspond poorly with observed Hadley OI Atl3 SST anomalies are plotted in black. 148
- 6.18 Cold year (1983, 1986, 1992, 1994 and 2004) composites of (a) anomalous Atl3 SST, equatorial Atlantic APE (3°S-3°N 60°W-15°E 0-400m), tropical Atlantic APE (8°S-8°N 60°W-15°E 0-400m) and tropical Atlantic BPE (8°S-8°N 60°W-15°E 0-400m); (b) total versus climatological values of Atl3 SST, tropical Atlantic APE and tropical Atlantic BPE; (c) total versus climatological values for wind work (Φ_{ww}) and advection of the density field (Φ_{bpa}); and (d) the relative contribution of mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ to anomalous wind power. In (a)-(c) values have been normalised while in (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability. 149
- 6.19 Cold year (1983, 1986, 1992, 1994 and 2004) composite monthly maps illustrating the spatial structure in wind power anomalies (Wm^{-2}) over the tropical Atlantic domain (8°S-8°N 60°W-15°E) between April and September are shown in the left column (a, c, e, g, i, k and m). These are compared with the spatial structure in climatological wind power anomalies from the annual mean in the right column (b, d, f, h, j, l and n). 150
- 6.20 Cold year (1983, 1986, 1992, 1994 and 2004) composite of tropical Atlantic APE versus wind power phase diagram. The phase diagram based on composited total APE and wind power values is shown in red while the seasonal phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability. 151
- 6.21 (a) 1992 and (b) 2004 anomalous Atl3 SST, equatorial Atlantic APE (3°S-3°N 60°W-15°E 0-400m), tropical Atlantic APE (8°S-8°N 60°W-15°E 0-400m) and tropical Atlantic BPE (8°S-8°N 60°W-15°E 0-400m). (c) The 1992 and (b) the 2004 relative contribution of mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ to anomalous wind power. In (a) and (b) values have been normalised while in (c) and (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability. 152
- 6.22 (a) 1992 and, (b) 2004, total versus climatological values for Atl3 SST and equatorial Atlantic APE. Values have been normalised. A 14 day running mean has been applied to each time series to smooth out the high frequency variability. 153

- 6.23 Tropical Atlantic APE versus wind power phase diagram for (a) 1992 and (b) 2004. The phase diagram based on total APE and wind power values is shown in red while the climatological phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability. 153
- 6.24 Warm year (1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999) composites of (a) anomalous Atl3 SST, equatorial Atlantic APE (3°S - 3°N 60°W - 15°E 0-400m), tropical Atlantic APE (8°S - 8°N 60°W - 15°E 0-400m) and tropical Atlantic BPE (8°S - 8°N 60°W - 15°E 0-400m); (b) total versus climatological values of Atl3 SST, tropical Atlantic APE and tropical Atlantic BPE; (c) total versus climatological values for wind work (Φ_{ww}) and advection of the density field (Φ_{bpa}); (d) the relative contribution of mean perturbation terms Φ_{wwano}^{mupr} and Φ_{wwano}^{mtpu} to anomalous wind power. In (a)-(c) values have been normalised while in (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability. 155
- 6.25 Warm year (1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999) composite monthly maps illustrating the spatial structure in wind power anomalies (Wm^{-2}) over the tropical Atlantic domain (8°S - 8°N 60°W - 15°E) between April and September are shown in the left column (a, c, e, g, i, k and m). These are compared with the spatial structure in climatological wind power anomalies from the annual mean in the right column (b, d, f, h, j, l and n). 156
- 6.26 Warm year (1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999) composite of tropical Atlantic APE versus wind power phase diagram. The phase diagram based on composited total APE and wind power values is shown in red while the climatological phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability. 157
- 6.27 (a) 1988 and (b) 1991 anomalous Atl3 SST, equatorial Atlantic APE (3°S - 3°N 60°W - 15°E 0-400m), tropical Atlantic APE (8°S - 8°N 60°W - 15°E 0-400m) and tropical Atlantic BPE (8°S - 8°N 60°W - 15°E 0-400m). (c) The 1988 and (b) the 1991 relative contribution of mean perturbation terms Φ_{wwano}^{mupr} and Φ_{wwano}^{mtpu} to anomalous wind power. In (a) and (b) values have been normalised while in (c) and (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability. 159
- 6.28 (a) 1988 and, (b) 1991, total versus climatological values for Atl3 SST and equatorial Atlantic APE. Values have been normalised. A 14 day running mean has been applied to each time series to smooth out the high frequency variability. 160
- 6.29 Tropical Atlantic APE versus wind power phase diagram for (a) 1988 and (b) 1991. The phase diagram based on total APE and wind power values is shown in red while the climatological phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability. 160

- 6.30 A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) BPE anomalies and tropical Atlantic (8°S - 8°N 60°W - 15°E 0-400m) BPE anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). BPE anomalies were calculated based on ROMS-TAtl output following Equation 3.2 and have been normalised. 161
- 6.31 A comparison between monthly inter-annual anomalies in BPE evolution ($\text{dBPE}_{\text{ano}}/\text{dt}$ in Js^{-1}) and BPE evolution driven by fluctuations in the advection of the density field ($\Phi_{\text{bpa}_{\text{ano}}}$ in Js^{-1}), assessed over (a) the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m) and (b) the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). 162
- 6.32 (a) A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE and BPE anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). APE and BPE values have been normalised. (b) Time lag correlations between equatorial APE and BPE anomalies. (c) The seasonal dependence in correlations between equatorial Atlantic APE and BPE anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. APE and BPE anomalies were calculated based on ROMS-TAtl output following Equation 3.2. 163
- 6.33 (a) A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) BPE and Atl3 SST anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and BPE values have been normalised. (b) Time lag correlations between equatorial Atlantic BPE and Atl3 SST anomalies. (c) The seasonal dependence in correlations between equatorial Atlantic BPE and Atl3 SST anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. BPE anomalies were calculated based on ROMS-TAtl output following Equation 3.2. 164
- A.1 Variability in the vertical density profile estimated by horizontally averaging the monthly mean density field within the chosen domain: (a) the equatorial Pacific (5°N - 5°S 130°E - 85°W), (b) the equatorial Atlantic (3°S - 3°N 60°W - 15°E), (c) the tropical Pacific (8°N - 8°S 130°E - 85°W), (d) the tropical Atlantic (8°S - 8°N 60°W - 15°E), (e) the extended tropical Pacific (15°N - 15°S 130°E - 85°W), and (f) the extended tropical Atlantic (15°S - 15°N 60°W - 15°E). Seasonal variability in the vertical density profile is overlayed on the total spread of vertical density profiles between 1958-2004 shown in cyan. Depth (m) is represented on the y-axis and density (kgm^{-3}) on the x-axis. Monthly density fields are derived from the monthly ORCA temperature and salinity fields. II

A.2	Seasonal changes in equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE obtained using Equation 3.3, APE and BPE obtained using Equation 3.2, as well as seasonal changes in the Mass of this domain. Mass, BPE and APE values have been normalised. Values calculated using ORCA data.	III
A.3	A comparison between total Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE variability. SST and APE values were calculated using ORCA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended.	IV
A.4	A comparison between total Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE variability. SST and APE values were calculated using ORCA data. The correlation between the Atl3 SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended. . . .	IV
A.5	A comparison between total and climatological equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE values. APE values were calculated using ORCA data and have been detrended.	V
A.6	A comparison between total and climatological equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE values. APE values were calculated using ORCA data and have been detrended.	V
A.7	(a) Equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE versus tropical Pacific (15°N-15°S 130°E-85°W 0-400m) APE. (b) Equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE versus tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE. APE values where calculated using ORCA data.	VI
A.8	A comparison between inter-annual Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE anomalies. SST and APE anomalies where calculated using ORCA data. The correlation between anomalies in each SST index and APE (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended.	VI
A.9	A comparison between climatological Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE. SST and APE values where calculated using ORCA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised.	VII
A.10	A comparison between climatological Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE. SST and APE values where calculated using ORCA data. The correlation between the Atl3 SST index and equatorial Atlantic APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised.	VII

- A.11 A comparison between inter-annual Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. SST and APE anomalies were calculated using ORCA data. The correlation between anomalies in each SST index and APE (r) is given on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended. VIII
- A.12 The seasonal dependence in correlations between Atl3 SST and equatorial Atlantic APE anomalies. SST and APE anomalies were calculated using ORCA data. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. VIII
- D.1 Balance of terms for the KE evolution equation in the form given by Equation D.5, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). dKE/dt represents the actual rate of change of KE whereas *derived* dKE/dt represents the rate of change of the KE estimated by balancing the terms within Equation D.5. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. XVI
- D.2 Balance of terms for the KE evolution equation in the form given by Equation D.9, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). dKE/dt represents the actual rate of change of KE whereas *derived* dKE/dt represents the rate of change of the KE estimated by balancing the terms within Equation D.9. A 14 day running mean has been applied to the time series to smooth out the high frequency variability. XVIII

Chapter 1

Introduction

1.1 Background

Despite the dominant semi-annual signal in solar forcing at the top of the atmosphere, atmospheric and oceanic conditions in the central-eastern equatorial Pacific and Atlantic basins exhibit a distinct annual cycle (Mitchell and Wallace, 1992). In these regions, atmospheric and oceanic conditions observed at the time of the two equinoxes strongly contrast each other. This seemingly inconsistent behaviour is due to the integral role ocean-atmosphere interactions play in determining the seasonal cycle within these regions (Mitchell and Wallace, 1992).

Tropical oceans are particularly susceptible to ocean-atmosphere interactions. Firstly, atmospheric circulation exhibits a strong dependence on underlying Sea Surface Temperature (SST) in the tropics (Rowell, 1998) as deep atmospheric convection is driven by the warmer ($>27^{\circ}\text{C}$) SSTs which reside in the tropics. Secondly, while extra-tropical SSTs are predominately determined by the magnitude of the surface heat flux, with ocean dynamics tending to play a relatively small role, the oceanic response to wind forcing has a much stronger influence on SST in the tropics (Kraus, 1977; Philander, 1990). This is primarily because the depth of the surface mixed layer is not determined locally over most of the tropics due to the presence of a sharp, shallow thermocline that separates cold sub-thermocline waters from a thin layer of warm surface waters. The depth of this thermocline is remotely forced by large scale oceanic adjustment in response to changes in atmospheric forcing involving oceanic Kelvin and Rossby waves (Philander, 1990).

At the inter-annual time scales, global climate variability is primarily associated with the El Niño Southern Oscillation (ENSO) (Wallace et al., 1998). While ENSO is the strongest mode of coupled climate variability in the tropics and has far reaching impacts, there are other tropical modes of coupled climate variability in the Indian and Atlantic Oceans which have received considerably less attention. Interaction between the tropical oceans and their overlying atmosphere is seen to give rise to clear signals of coupled climate variability at several time scales (seasonal, inter-annual, and decadal as outlined in the review paper of Chang et al., 2006a). The upper equatorial ocean is influenced by the atmosphere via thermodynamic and/or dynamic processes. In response, the ocean acts to reinforce, in the case of a positive feedback, or dampen, in the case of negative feedback, atmospheric behaviour through changes in its upper ocean heat content and SST field.

From a practical point of view, when forecasting seasonal anomalies in atmospheric fields such as temperature and rainfall, one of the greatest limitations lies in providing atmospheric models with accurate future variations in SST. Given the correct evolution of global SSTs as boundary conditions, atmospheric models are able to produce a realistic and statistically significant atmospheric response (Trenberth et al., 1998). Numerical models of the ocean on the other hand can reproduce SST evolution with reasonable accuracy when forced with observed wind stress and surface heat flux forcing from the atmosphere. Although it should be noted that this skill is largely determined by the restoring effect in the bulk formulation of surface heat fluxes (Seager et al., 1988). It is the interaction between these two fluid layers that is an important source of feedbacks, which lead to the growth of empirical errors in coupled ocean-atmosphere models, and limit our ability to predict both future SST fields and the resultant atmospheric response.

As SST changes represent variability at the interface between each component of the coupled ocean-atmosphere system, one can gain great insight into the mechanisms driving coupled variability by understanding the primary processes controlling the evolution of SST. The type of ocean-atmosphere feedback mechanism behind coupled variability may be identified according to the primary physical process governing SST changes. Feedback mechanisms associated with coupled equatorial variability are seen to be either thermodynamic or dynamic in nature. If SST changes associated with coupled variability are driven primarily by changes in the intensity of the surface heat flux forcing, then a thermodynamic feedback mechanism is behind the growth of coupled anomalies. On the other hand, a dynamic feedback mechanism is responsible if advection and mixing processes are behind the SST changes, as changes in upper ocean circulation in response to wind forcing plays an active role in driving SST variability.

A thermodynamic feedback mechanism inherent to the tropics is the Wind-Evaporation-SST (WES) feedback mechanism (Xie and Philander, 1994; Chang et al., 2006a). As the name suggests, anomalous latent heating is responsible for the growth of SST anomalies attributed to the WES feedback mechanism. The WES feedback is initiated by the presence of a small anomalous cross-equatorial SST gradient. The atmospheric response to this anomalous SST pattern gives rise to anomalous cross-equatorial surface winds. The Coriolis force deflects the meridional flow such that it acts to increase climatological zonal wind speeds in the colder hemisphere enhancing evaporative cooling, and decrease climatological zonal wind speeds in the warmer hemisphere decreasing evaporative cooling. The meridional atmospheric boundary layer flow from the cooler to the warmer hemisphere therefore acts to reinforce the anomalous SST gradient through anomalous latent heating.

Another thermodynamic feedback mechanism active in the tropics is the Cloud-SST feedback (Wang and Enfield, 2001; Okumura et al., 2001; Tanimoto and Xie, 2002). Two kinds of Cloud-SST feedback mechanisms of potential importance in the tropical Atlantic are identified in the literature. Firstly, there is a feedback mechanism between SST anomalies in the western basin's warm pool region and cloud induced longwave radiation anomalies. Warm (cold) SST anomalies decrease (increase) Sea Level Pressure (SLP), thereby increasing (decreasing) atmospheric convection and cloudiness. This increased (decreased) cloudiness decreases (increases) the loss of longwave radiation, reinforcing the warm (cold) SST anomaly (Wang and Enfield, 2001). Secondly, there is a feedback mechanism involving low-level stratiform clouds in the subtropics. Cold (warm) SST anomalies increase (decrease) the static stability of the atmospheric boundary layer enhancing (suppressing) stratus cloud formation which limits the amount of solar radiation reaching the earth's surface by reflecting it back up into space, reinforcing the cold (warm) SST anomaly (Klein and Hartmann, 1993; Okumura et al., 2001; Tanimoto and Xie, 2002).

Equatorial regions support two distinct types of dynamic feedback mechanisms known as the 'Bjerknes'/'thermocline' feedback and the 'Ekman'/'surface-layer' feedback mechanisms (Neelin et al., 1998). Both of these feedback mechanisms involve a circular argument whereby SST changes are both the cause and result of wind perturbations. These feedbacks arise from the fact that equatorial oceans have a rather special mean state which facilitates the coupling between subsurface and surface ocean processes.

This mean state, in the case of the Pacific and Atlantic ocean basins, is characterised by a zonal sea level and corresponding thermocline slope across the basin. The thermocline shoals in the east, as the mean easterly zonal wind stress forcing is balanced at the first order by a zonal pressure gradient. Another key aspect of the mean state is the change in sign of Coriolis forcing across the equator and the associated Ekman divergence which drives equatorial upwelling. This equatorial upwelling, superimposed on the zonally sloping thermocline, in turn results in the zonal SST gradient which supports the atmospheric Walker circulation that gives rise to the mean equatorial easterly trade winds.

As a result of these mean state characteristics, the dynamic ocean response to small perturbations in wind forcing is capable of forcing central-eastern basin equatorial SST changes big enough to feedback and significantly effect atmospheric circulation. The shallow thermocline in the east means that SST is strongly modulated by the vertical advection of colder subsurface water into the surface mixed layer. The extent of cooling associated with this vertical advection depends on either the strength of vertical velocities induced by local Ekman pumping, or on remotely forced changes in the thermocline depth induced by a basin wide adjustment to wind stress fluctuations.

The Bjerknes feedback (Bjerknes, 1969), involves a circular argument whereby anomalously warm (cool) SSTs in the eastern equatorial basin modify the basin's Walker circulation causing a decrease (increase) in the easterly trade winds. This perturbation to the zonal wind stress forcing triggers oceanic equatorial Kelvin waves that act to zonally redistribute warm water along the equator decreasing (increasing) the east-west slope in the thermocline. The resulting thermocline depth changes in the east then modulate the strength of vertical advection such that the original SST anomaly is enhanced completing the positive feedback loop.

The ocean-atmosphere feedback mechanism referred to as the Ekman feedback, involves a purely local dynamical ocean response to changes in the wind forcing. Like the Bjerknes feedback, SST anomalies modify the strength of equatorial winds, which then act to reinforce the original SST anomaly. However, unlike the Bjerknes feedback mechanism, thermocline depth variations associated with the Ekman feedback mechanism are driven by the divergence of surface Ekman currents rather than remotely forced geostrophic current anomalies. Upper ocean dynamics influence SST through anomalous advection driven by the divergence of surface Ekman currents and associated thermocline depth changes are in balance with the local wind stress forcing (Chang and Philander, 1994).

Idealised, natural modes of oscillation arising from dynamical interactions between the equatorial oceans and their overlying atmosphere can be considered in terms of the oceanic processes and associated dynamical feedbacks governing SST fluctuations (Neelin et al., 1998). Classified in this manner, three main types of idealised modes have previously been suggested, namely the "delayed oscillator" (Hirst, 1986; Suarez and Schopf, 1988; Battisti and Hirst, 1989), the "SST-modes" (Neelin, 1991) and the "recharge oscillator" (Jin, 1997).

The delayed oscillator Central-eastern basin SST fluctuations attributed to the idealised “delayed oscillator” mode are driven by equatorial upwelling acting on thermocline displacements. The Bjerknes feedback leads to the growth of anomalies associated with this mode. The lagged negative feedback which leads to the periodic nature of this mode is due to the reflection of off-equatorial Rossby waves generated during the initial phase at the western edge of the basin. These waves then propagate to the east along the equator as Kelvin waves, initiating the phase reversal (Battisti and Hirst, 1989). Thermocline displacements are driven by equatorial ocean wave dynamics, and so the delayed oscillator mode has a time scale which depends on the processes of oceanic adjustment (Hirst, 1986; Suarez and Schopf, 1988). The ocean provides the slow adjustment time scale as well as the memory for transition from one phase to the next (Neelin et al., 1998).

SST-modes A westward propagating, equatorially symmetric mode of coupled ocean-atmosphere variability referred to as an SST-mode, is forced by the local Ekman feedback mechanism (Neelin et al., 1998; Neelin, 1991). This equatorially symmetric SST-mode together with equatorially antisymmetric SST-modes influences the seasonal cycle in the eastern part of both the Pacific and Atlantic where the thermocline is shallow (Chang and Philander, 1994; Carton and Zhou, 1997). Forced by the local Ekman feedback mechanism, these coupled unstable modes do not involve remotely forced changes in thermocline depth, thus rendering the time scale of equatorial wave dynamics unimportant. The oscillatory time scale is determined by local advective processes as SST changes are determined by the divergence of surface Ekman currents and associated thermocline depth changes (Chang and Philander, 1994). The growth rate of these modes is inversely proportional to the mean thermocline depth and therefore the existence of these modes depends on a shallow thermocline (Chang and Philander, 1994).

The recharge oscillator Like the delayed oscillator, the recharge oscillator is an idealisation of the coupled ocean-atmosphere mode with which the ENSO is associated (Jin, 1997). As with the delayed oscillator, the recharge oscillator paradigm ascribes the growth of instabilities to the Bjerknes feedback mechanism. However, the recharge oscillator idealisation differs from the delayed oscillator idealisation in its approximation of the phase transition mechanisms. As described above, the phase-transition mechanism in the case of the delayed oscillator idealisation is attributed to the delay associated with planetary wave propagation and western boundary wave reflection. In the case of the recharge oscillator idealisation, the ocean wave propagation process is not at the centre of the delayed negative feedback, but rather regarded as a part of the oceanic adjustment in which heat and mass are re-distributed under changing wind forcing. The recharge-discharge of equatorial zonal mean heat content is viewed as the phase-transition mechanism. The basin wide ocean adjustment, which is viewed as being in quasi-Sverdrup balance with anomalous trade wind forcing, results in an anomalously shallow (deep) equatorial thermocline after warm (cold) events. This shallowing (deepening) of the mean equatorial thermocline while the zonal equatorial thermocline slope remains suppressed (enhanced) in response to anomalously weak (strong) equatorial trades, provides the negative feedback/phase transition mechanism that allows the thermocline to shoal (deepen) in the east. As the regulator of eastern basin SST, this delayed reversal of eastern basin thermocline depth acts to reverse the SST anomaly (Jin, 1997).

1.2 Objectives

This thesis focuses on local coupled variability in the equatorial Atlantic. It aims to advance our understanding of the mechanisms behind tropical Atlantic climate variability and, in particular, the role of ocean dynamics within observed coupled variability in the equatorial Atlantic. Relative to the Pacific, the role of the ocean within equatorial Atlantic climate variability is poorly understood and particularly with respect to the role of ocean memory (Chang et al., 2006a). In the Pacific, energetics analysis of inter-annual variability has elucidated the mechanisms through which ocean-atmosphere interactions give rise to ENSO (Goddard and Philander, 2000; Fedorov et al., 2003; Fedorov, 2007). In this thesis, an assessment of the energetics of equatorial Atlantic variability, as well as of the physical processes driving SST variability is performed. This analysis offers insight into the role of the ocean within equatorial Atlantic ocean-atmosphere interactions at both seasonal and inter-annual time scales.

Figure 1.1 depicts the observed seasonal variability in SST, wind stress and precipitation at three locations (western, central and eastern basin) within the Pacific, Atlantic and Indian oceans. Apparent in Figure 1.1 is the significant seasonal cooling observed in the central-eastern equatorial Pacific (Figures 1.1 a-c) and Atlantic (Figures 1.1 d-f) which is absent from the Indian Ocean (Figures 1.1 g-i). This seasonal cooling is indicative of the key role dynamic ocean processes play in driving seasonal SST changes in the central-eastern equatorial Atlantic (Foltz et al., 2003; Peter et al., 2006; Yu et al., 2006) and the central-eastern equatorial Pacific (Hayes et al., 1991; Chang, 1993, 1994; Swenson and Hansen, 1999; Wang and McPhaden, 1999) where seasonal SST variations cannot be explained by surface heat flux variations alone. Overlaid on seasonal SST changes, precipitation and wind stress changes display the covariability observed within seasonal oceanic and atmospheric conditions in both the central-eastern Pacific and Atlantic basins, evidence that ocean-atmosphere interactions are integral in determining the seasonal cycle within these regions.

Figure 1.2 compares the standard deviation of monthly, total¹ SST variations, seasonal SST fluctuations and inter-annual SST anomalies spanning 1958-2004 for the tropical Pacific and Atlantic basins. As shown in Figure 1.2b, the Atl3 (3°S-3°N 20°W-0°E) region in the central-eastern Atlantic, like the Niño3.4 (5°N-5°S 170°W-120°W) and Niño3 (5°N-5°S 150°W-90°W) boxes in the Pacific (Figure 1.2a), represent the region of greatest equatorial SST variance.

Although seasonal SST variability in the eastern-central Atlantic, like seasonal SST variability in the central-eastern Pacific, has been attributed to seasonally excited SST-modes and their associated Ekman feedback (Chang and Philander, 1994), the seasonal cycle in Atl3 SST is somewhat distinct from that of the Niño3.4 and Niño 3 boxes. These distinctions are illustrated in Figure 1.3, which compares seasonal variability in the Niño 3.4, Niño 3 and Atl3 SST indices. The annual decline of SST in both basins, associated with the development of the equatorial cold tongue in April is evident. However, seasonal cooling in the Atlantic is steeper and greater than in the Pacific. Furthermore, cool conditions do not persist as long in the Atlantic as they do in the Pacific. This comparison between the seasonal SST signatures within each basin raises the following question: What distinct processes occurring in the Atlantic may account for the observed differences in seasonal central-eastern basin SST signature between the Pacific and the Atlantic? Based on the relationship between SST, thermocline depth and surface zonal currents, Ding et al. (2009) suggest that the mechanism behind seasonal SST changes in the Atlantic is more consistent with the presence of

¹“total” refers to the combined seasonal and inter-annual signals

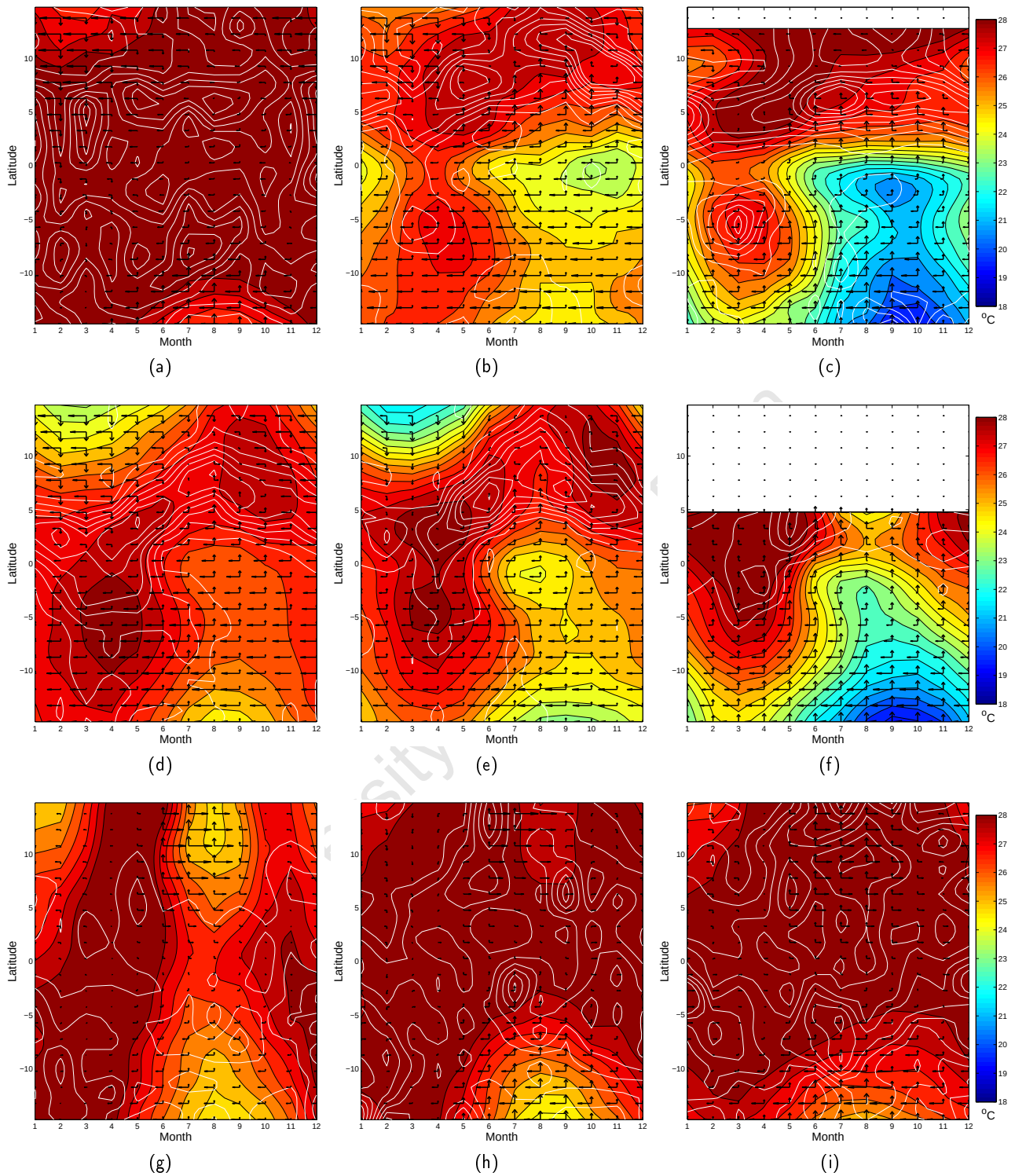


Figure 1.1: Seasonal SST, wind stress, and rainfall fluctuations in the (a) western Pacific Ocean - 165°E, (b) central Pacific Ocean - 120°W, (c) eastern Pacific Ocean - 90°W, (d) western Atlantic Ocean - 30°W, (e) central Atlantic Ocean - 20°W, (f) eastern Atlantic Ocean - 0°W, (g) western Indian Ocean - 55°E, (h) central Indian Ocean - 70°E, and (i) eastern Indian Ocean - 90°E. Climatological SST, wind stress and rainfall values shown here were derived from the Comprehensive Ocean-Atmosphere Data Set (COADS) (DaSilva et al., 1994). The colour filled black contours represent SST values in °C as indicated by the corresponding colourbar. The white contours represent precipitation with a contour interval of 0.2cmday^{-1} . The black vectors represent seasonal changes in the zonal and meridional wind stress components.

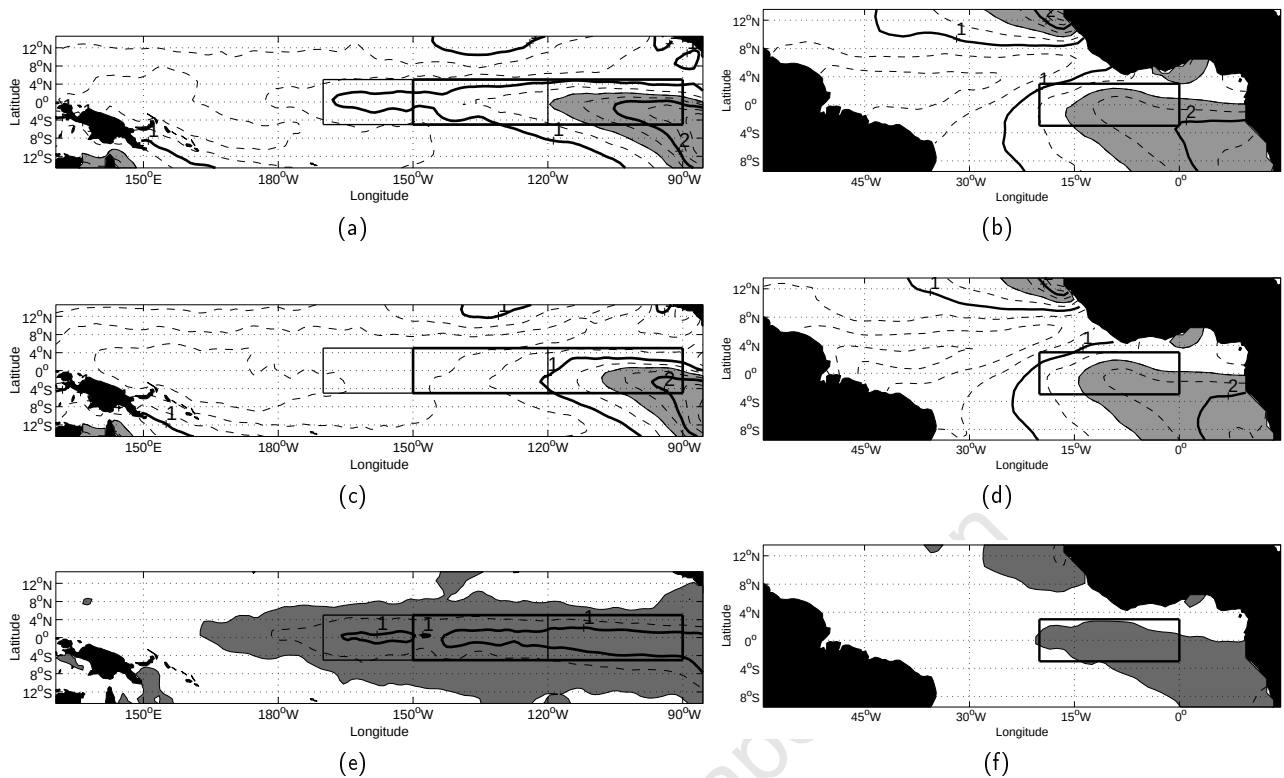


Figure 1.2: Based on 1° resolution, Hadley OI SST data (Rayner et al., 2003) spanning 1958-2004, this figure shows the standard deviation of monthly, (a) total SST variations in the Pacific, (b) total SST variations in the Atlantic, (c) seasonal SST fluctuations in the Pacific, (d) seasonal SST fluctuations in the Atlantic, (e) inter-annual SST anomalies in the Pacific, and (f) inter-annual SST anomalies in the Atlantic. A contour interval of 0.25°C has been used. Regions of $>1.5^{\circ}\text{C}$ are shaded in figures a-d and $>0.5^{\circ}\text{C}$ in figures e and f. The Niño 3.4 (5°N-5°S 170°W-120°W) and Niño 3 (5°N-5°S 150°W-90°W) and the Atlantic Atl3 (3°S-3°N 20°W-0°E) regions have been demarcated.

a “thermocline mode” than an SST-mode. This suggestion is substantiated by the results obtained in this thesis.

Furthermore, when comparing SST variability in the equatorial Atlantic with that of the equatorial Pacific it is apparent that while the inter-annual ENSO signal dominates the variability in the Pacific, the annual cycle dominates in the Atlantic (Vauclair and du Penhoat, 2001; Xie and Carton, 2004). This difference between the two basins is best illustrated by Figure 1.4, a comparison between variability in spatially averaged SST over the Pacific Niño 3.4 (5°N-5°S 170°W-120°W) and Niño 3 (5°N-5°S 150°W-90°W) regions, and the Atlantic Atl3 region (3°S-3°N 20°W-0°E).

The nature of SST variability in the central-eastern equatorial Atlantic is strikingly different from that of the Pacific. In the central Pacific (Figure 1.4a), seasonal SST variations of the order of 1°C are overshadowed by larger inter-annual excursions. This situation remains true even in the eastern Pacific (Figure 1.4b), where seasonal oscillations in SST are amplified by the shallowness of the thermocline. Seasonal SST changes in the central-eastern Atlantic on the other hand are substantially larger in magnitude than any inter-annual SST anomalies (Figure 1.4c). To what may we attribute the noticeable difference in the time-scale of the largest SST fluctuations between the two basins? The dominant mechanism responsible for the large inter-annual changes in central-eastern basin SST in the equatorial Pacific is well established (Chang

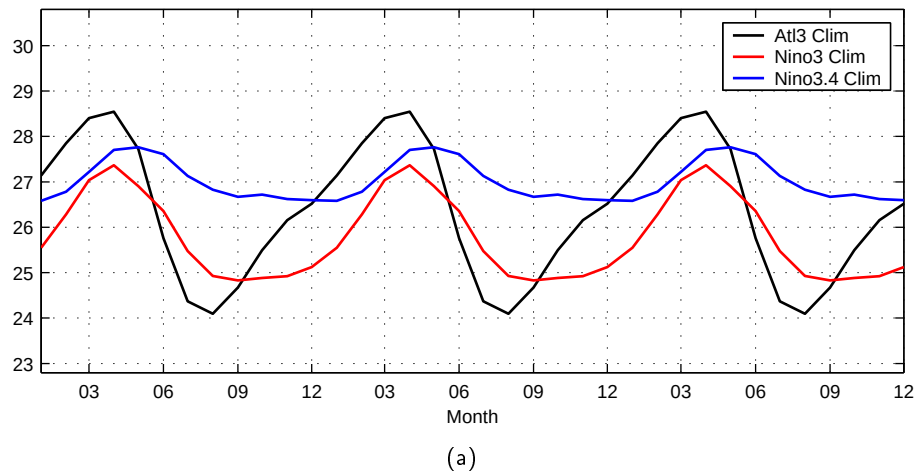


Figure 1.3: A comparison between seasonal variability in Niño 3.4 (5°N-5°S 170°W-120°W) SST, Niño 3 (5°N-5°S 150°W-90°W) SST, and Atl3 (3°S-3°N 20°W-0°E) SST. Created using monthly Hadley OI SST data (Rayner et al., 2003). SST values are given in °C.

et al., 2006a). Remotely forced thermocline depth perturbations, associated with the Bjerknes feedback mechanism, modulate the vertical advection of cool, sub-thermocline waters into the surface mixed layer. Does this mean that the Bjerknes feedback mechanism is acting primarily on seasonal time scales in the Atlantic and hence the largest SST variability occurs seasonally?

Given the supremacy of seasonal SST variability in the central-eastern equatorial Atlantic (Figure 1.4), one wonders whether the seasonal cycle in the Atlantic is so dominant that it is able to strongly influence the evolution of its inter-annual variability? Figure 1.4 suggests that unlike in the Pacific, where inter-annual SST variability is characterised by the presence of a natural coupled mode operating on inter-annual time scales (ENSO), inter-annual SST variability in the eastern Atlantic is driven by a modulation of seasonally active coupled variability. Can inter-annual SST variability in the eastern equatorial Atlantic be explained in terms of a modulation of seasonally active coupled variability?

The key questions raised above are listed below:

- What distinct processes occurring in the Atlantic may account for the observed differences in the seasonal central-eastern basin SST signature between the Pacific and the Atlantic?
- Why do inter-annual SST fluctuations dominate over seasonal fluctuations in the central-eastern Pacific while the opposite is true in the Atlantic? Is it because, unlike in the Pacific, the Bjerknes mechanism acts primarily on seasonal time scales in the Atlantic?
- Can inter-annual SST variability in the central-eastern equatorial Atlantic be explained in terms of a modulation of seasonally active coupled variability?
- What is the role of ocean memory within central-eastern equatorial Atlantic SST variability?

As highlighted in the previous section, the ocean-atmosphere feedback mechanism responsible for coupled variability can be identified according to the primary physical process governing associated SST changes; SSTs link the oceanic and atmospheric components of the coupled system. The approach taken in this thesis is therefore one which concentrates on diagnosing the role of dynamic ocean-atmosphere interaction

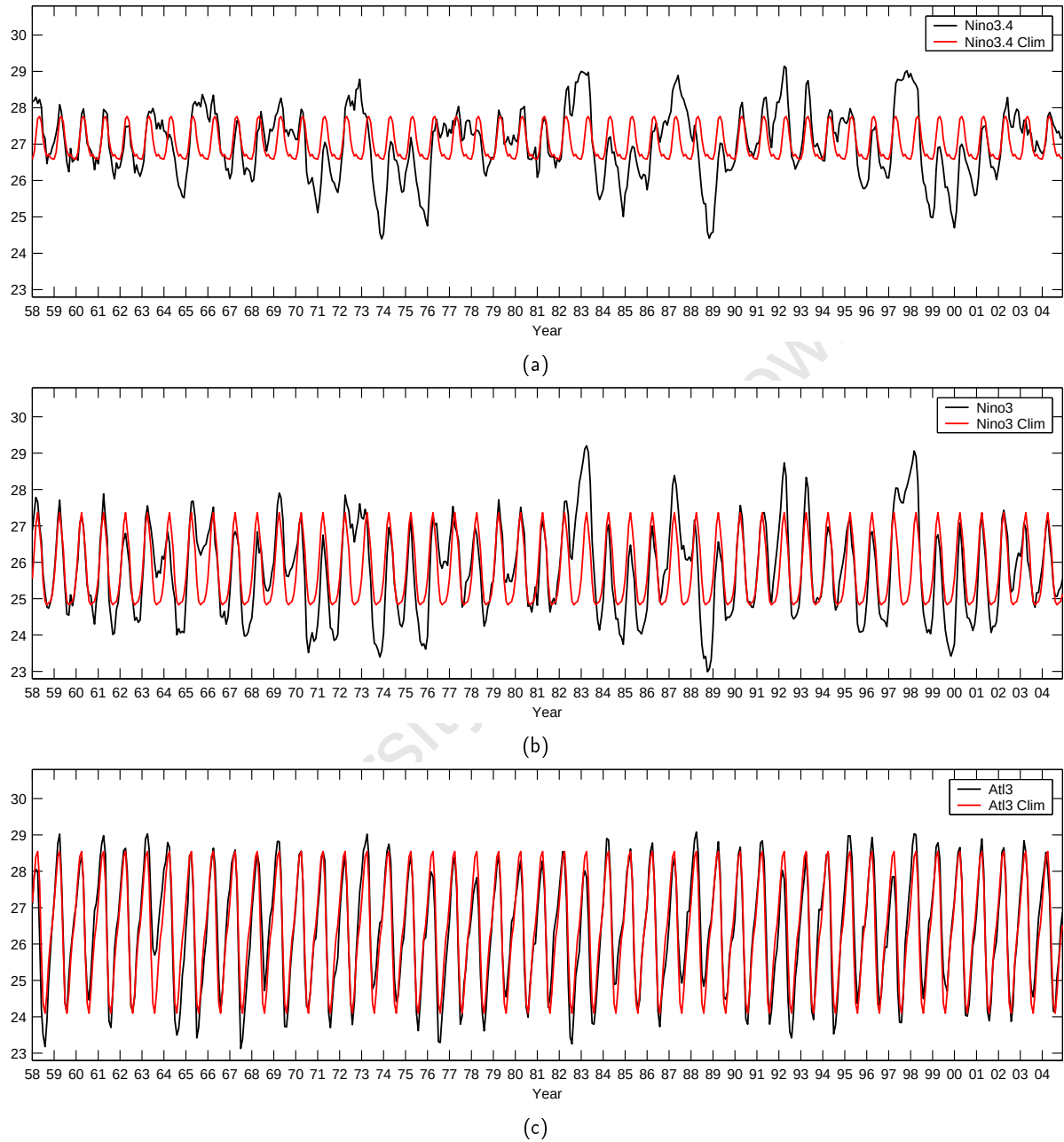


Figure 1.4: A comparison between variability in total versus climatological (a) Niño 3.4 (5°N-5°S 170°W-120°W) SST, (b) Niño 3 (5°N-5°S 150°W-90°W) SST, and (c) Atl3 (3°S-3°N 20°W-0°E) SST values. Created using monthly Hadley OI SST data (Rayner et al., 2003). SST values are given in °C and have been detrended.

within equatorial Atlantic coupled variability from a perspective that focuses on the mechanisms governing SST evolution.

Establishing dynamic coupled variability involves establishing the presence of both the ocean's influence on the atmosphere and the atmosphere's influence on the ocean. Within the analysis conducted in this thesis the response of the atmosphere to SST changes is assessed simply in terms of surface wind changes in response to altered atmospheric circulation. Whereas, the dynamic response of the ocean to changes in wind stress forcing and the subsequent impact on SST is studied in detail. This oceanic focus is motivated by the fact that it is ocean dynamics, as the slow time scale and source of oscillatory behaviour, that usually provides predictability in the tropical coupled ocean-atmosphere system (ocean dynamics provide the mechanism to reverse SST tendencies).

In the following chapter (Chapter 2), a review of our current understanding of mean, seasonal and inter-annual conditions within the tropical Atlantic is provided. Thereafter, Chapter 3 broadly addresses the first three questions posed above by capitalising on the vast knowledge gained in the tropical Pacific and contrasting the nature of variability in a number of pertinent variables in the Atlantic against the Pacific. This comparison reveals the uniqueness of the equatorial Atlantic. Chapter 3 is aimed at establishing the relationship between remotely forced ocean dynamics and central-eastern basin equatorial Atlantic SST variability at both the seasonal and inter-annual time scales. This assessment of equatorial Atlantic variability sheds light on the role of upper ocean variability, suggesting that the dominant upper ocean response to wind variations which operates at inter-annual time scales in the tropical Pacific and plays a crucial role in the deterministic nature of ENSO, is in fact occurring predominately at seasonal time scales in the Atlantic.

Inferences made based on the relationships observed in Chapter 3 are considered further in Chapters 5 and 6. To facilitate the diagnostics undertaken in Chapters 5 and 6, a simulation of oceanic conditions within the tropical Atlantic between 1980 and 2004 has been conducted using the Regional Ocean Modelling System (ROMS). The configuration and validation of this inter-annual simulation is detailed in Chapter 4.

Chapter 5 pursues the concept that the oceanic processes associated with inter-annual SST variability in the Pacific, play an active role in dictating seasonal SST variability in the equatorial Atlantic. This chapter focuses purely on seasonal variability in the tropical Atlantic. The analysis conducted in Chapter 5 is divided into two parts. The first part investigates the relative contribution of each surface layer process to seasonal SST variability in the central-eastern equatorial Atlantic and is dedicated to establishing the relative role of remotely forced ocean dynamics in determining seasonal SST variability in this region. The second part of this chapter investigates the mechanisms responsible for seasonal thermocline depth changes and the role of ocean memory. Thermocline depth variability in the equatorial basin is concisely evaluated by equatorial Atlantic Available Potential Energy (APE) changes. The mechanisms responsible for seasonal thermocline depth changes are consequently established by addressing the question - What are the mechanisms responsible for seasonal variability in equatorial Atlantic APE?

In Chapter 6, equatorial Atlantic inter-annual variability is assessed within the context of the seasonally active mechanisms identified in Chapter 5. The mechanisms responsible for inter-annual thermocline depth changes associated with coupled variability and the role of ocean memory is assessed from an energetics perspective in this chapter.

Finally, conclusions based on the analysis conducted within this thesis are summarised in Chapter 7. Distinctions between the seasonal SST signature in the central-eastern Pacific and the central-eastern Atlantic

are ascribed to the presence, in the Atlantic, of a seasonally excited Bjerknes feedback as well as a delayed negative feedback mechanism which are absent from the Pacific. The sharper, enhanced seasonal decline of central-eastern basin SST observed in the Atlantic in comparison to the Pacific is attributed to the fact that the seasonally excited growth of asymmetric conditions about the equator is accompanied by a seasonally excited Bjerknes feedback as the thermocline shoals seasonally in the Atlantic. Furthermore, the fact that cool conditions do not persist as long as they do in the Pacific is ascribed to seasonal warming that is in phase with a deepening of the thermocline in the east from August. This seasonal decay in the seasonally excited Bjerknes feedback and deepening of the thermocline in the east is seen to be the result of a delayed, negative, feedback mechanism similar to the negative, feedback mechanism operating inter-annually in the Pacific; transient induced changes in surface currents, associated with the delayed response of the tropical Atlantic to seasonal changes in wind stress forcing, affect the ability of the wind to do work on the ocean. Between April-September a circular relationship between APE and the work done on the ocean by the wind suggests that a seasonally excited thermocline mode² of coupled variability driven by the seasonally excited Bjerknes feedback and its subsequent decay, plays an active role in the tropical Atlantic's seasonal cycle. The difference in the time-scale of the largest SST fluctuations between the Pacific and Atlantic basins is therefore attributed to the fact in the Pacific the Bjerknes feedback and its associated delayed, negative feedback mechanism operates on inter-annual time-scales whereas in the Atlantic it operates seasonally. Evidence is offered supporting the notion that inter-annual Atl3 variability associated with the zonal mode may be best understood as the result of the modulation either in phase or amplitude of this seasonally active process as opposed to a dynamically distinct natural mode of variability operating on inter-annual time scales as is the case with ENSO. As elaborated on in the final chapter (Chapter 7), differences between ENSO and the zonal mode mentioned in Chapter 2 can then be accounted for in terms of this distinction.

²As suggested by Ding et al. (2009) the term 'thermocline mode' is used to refer to the seasonally excited mode identified within the Atlantic with similar properties as the delayed oscillator mode operating inter-annually in the Pacific. A distinct term is used despite the similarity as unlike the delayed oscillator mode this mode occurs amidst seasonal forcing and is therefore different from the delayed oscillator idealisation which describes a free mode of the coupled system.

Chapter 2

Tropical Atlantic Climate Variability

2.1 Mean State

Figure 2.1 depicts the temporal mean of several atmospheric and oceanic variables within the tropical Atlantic: SST, wind stress, wind stress curl and its associated Sverdrup stream function, precipitation, dynamic topography, and surface currents. Evident in Figure 2.1 is a distinct asymmetry in mean atmospheric and oceanic conditions relative to the equator. Notable features in the tropical Atlantic's asymmetric mean state illustrated in Figure 2.1 include the cross-equatorial SST gradient between the warmer SSTs to the north and cooler SSTs in the south (Figure 2.1a), the southerly cross-equatorial component in the surface wind field (Figure 2.1b), a band of high precipitation to the north of the equator corresponding to the mean position of the Inter Tropical Convergence Zone (ITCZ) (Figure 2.1c), and an eastward oceanic current to the north of the equator that has no counterpart south of the equator (Figure 2.1f). Despite the fact that the mean position of maximum heating by the sun is on the equator, the warmest SSTs, over which the ITCZ resides, lie to the north of the equator on average (Figures 2.1a and 2.1c). These asymmetric conditions are indicative of the key role ocean-atmosphere interaction plays in determining the mean state of the tropical Atlantic.

From an atmospheric perspective, the above mentioned asymmetries in its mean state can be understood in terms of its response to the asymmetric SST pattern seen in Figure 2.1a. The establishment of this asymmetric SST pattern on the other hand can be explained in terms of the ocean's response to the southerly cross-equatorial surface winds that result from the asymmetric atmospheric conditions. Southerly cross-equatorial winds support upwelling and an elevated thermocline south of the equator and downwelling and a deep thermocline north of the equator thereby establishing the cross-equatorial SST gradient (Mitchell and Wallace, 1992; Philander et al., 1996). It is this circular argument whereby the wind influences the ocean and the ocean influences the wind that is behind the unstable ocean-atmosphere interaction which gives rise to these asymmetric climatic conditions (Mitchell and Wallace, 1992; Chang and Philander, 1994; Philander et al., 1996). This unstable ocean-atmosphere interaction converts equatorially symmetric conditions into asymmetric conditions by growing an initial perturbation in the symmetric state (Chang and Philander, 1994; Philander et al., 1996). In theory this asymmetrical state could form in either hemisphere, in reality asymmetrical geographic features such as asymmetry in the land mass distribution are presumed responsible for the ITCZ mainly lying in the northern hemisphere (Xie and Philander, 1994; Philander et al., 1996).

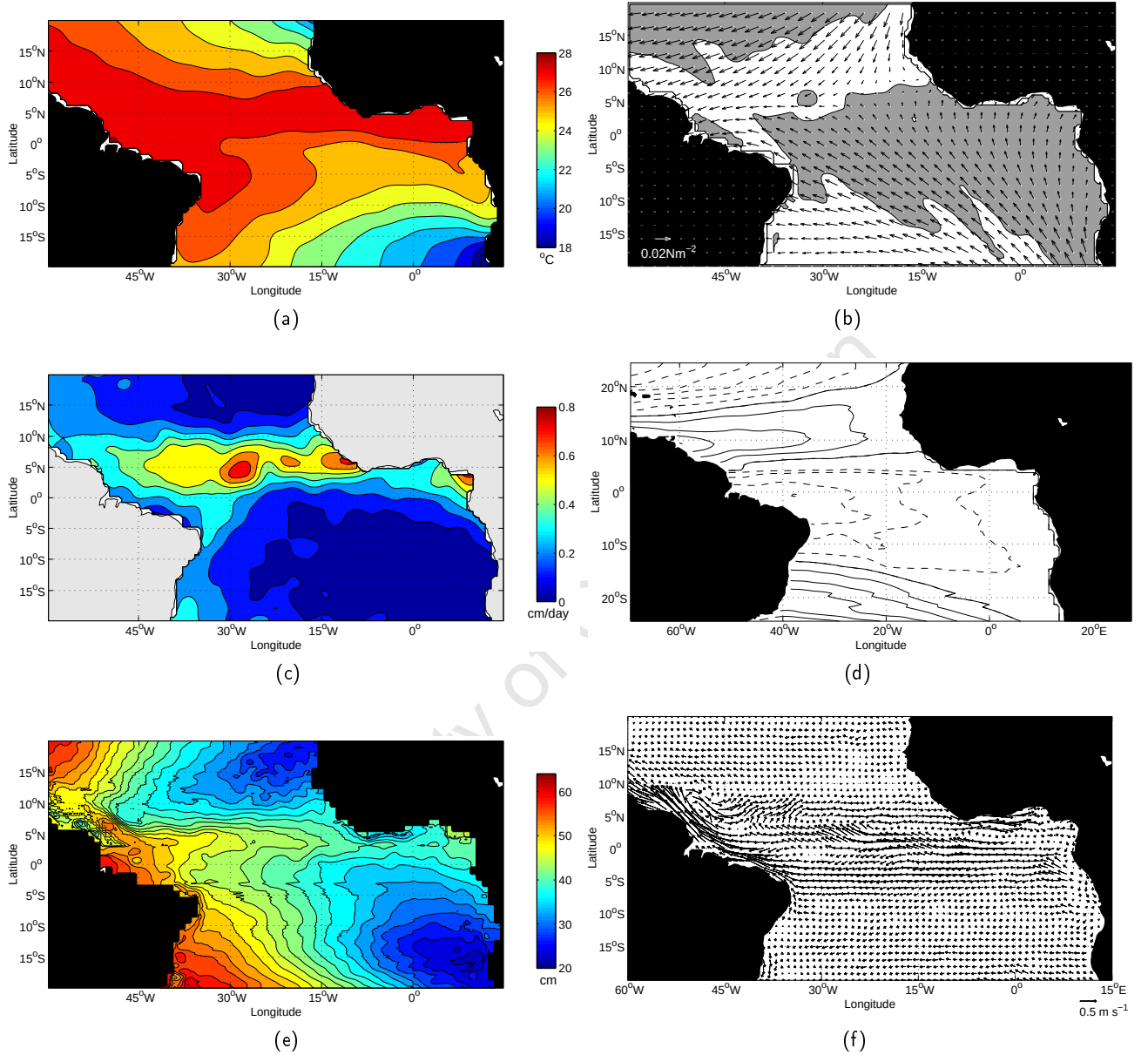


Figure 2.1: The temporal mean of several atmospheric and oceanic variables within the tropical Atlantic: (a) mean tropical Atlantic SST, (b) mean wind stress with regions of negative wind stress curl shaded in grey and regions of positive curl in white, (c) mean precipitation, (d) the Sverdrup stream function based on the mean wind stress field depicted in (b), (e) mean dynamic topography from the CNES-CLS09 mean dynamic topography product (Rio et al., 2009) (www.avisioceanobs.com), and (f) mean surface currents derived from the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). The mean SST field, wind stress field and rainfall field were derived from the Comprehensive Ocean-Atmosphere Data Set (COADS) (DaSilva et al., 1994). In (d), the stream functions have been plotted with a contour interval of 2 Sverdrup. Counter-clockwise circulation is indicated by the solid contours and clockwise circulation by the dashed contours.

The upper tropical Atlantic ocean can essentially be described as a two layer system, with cool deep waters separated from a shallow layer of warm surface waters by a sharp thermocline. The basin wide mean depth of this thermocline is determined by diabatic processes which dictate the amount of warm surface water residing above the thermocline in each basin. Boccaletti et al. (2004) argue that the primary diabatic process determining the depth of a basin's thermocline is the constraint of a balanced heat budget between the primary regions of heat loss within warm subtropical western boundary current regions and heat gained within the equatorial and coastal upwelling regions. The amount of heat gain at the equator depends strongly on the depth of the thermocline, it is large when the thermocline is shallow and is small when the thermocline is deep. Connected to the warm subtropical western boundary current regions by circulation systems referred to as the SubTropical Cells (STCs) (McCreary and Lu, 1994), the equatorial thermocline reaches a depth such that its heat gain is in equilibrium with their heat loss (Boccaletti et al., 2004). This idealisation may be an adequate approximation for the Pacific basin, however the situation in the Atlantic is complicated by another significant source of heat gain in the form of Agulhas leakage (Weijer et al., 2002; Byrne et al., 2006). Furthermore meridional circulation is not symmetrical about the equator as a net northward transport of warm surface waters across the equator forms the upper branch of the Meridional Over-turning Circulation (MOC) (Fratantoni et al., 2000).

The mean spatial distribution of warm surface water and hence the local depth of the thermocline is on the other hand maintained by the mean oceanic response to mechanical wind forcing. Mean SSH shown in Figure 2.1e provides an indication of the amount of warm surface water residing in a region and serves as a proxy for thermocline depth - regions of high SSH are associated with convergence of surface currents and a deep thermocline while regions of low SSH are associated with the divergence of surface currents and a shallow thermocline.

The mean state of the tropical Atlantic is characterised by a zonally sloping thermocline which shoals in the east, as mean easterly winds advect warm surface waters westward (Figures 2.1b and 2.1e). This zonal gradient in the depth of the thermocline results in a zonal SST gradient (Figure 2.1a), which in turn forces the zonal SLP gradient that is responsible for the mean easterly winds. Once again a circular argument exists, whereby the mean state of the ocean is characterised by the wind and the mean state of the wind is characterised by the ocean, which is indicative of the key role ocean-atmosphere interaction plays in determining the mean state of the tropical Atlantic.

The mean upper ocean response of the tropical Atlantic to wind forcing results in the current system depicted in Figure 2.2 (taken from Schouten et al. (2005)). Salient features of the tropical Atlantic current system illustrated in Figure 2.2 include the South Equatorial Current (SEC), the North Brazil Current (NBC), the Equatorial Under Current (EUC), the North Equatorial Counter Current (NECC) and the North Equatorial Current (NEC). Bounded by the westward flowing north and south equatorial currents, the tropical Atlantic circulation comprises of two wind driven gyres roughly in Sverdrup balance with the mean wind stress curl (Figure 2.1d, Fratantoni et al., 2000). Negative wind stress curl, situated roughly between 10°S and 5°N (Figure 2.1b), drives the clockwise circulation of the Equatorial Gyre (Figures 2.1d, 2.1f and 2.2), while a band of positive wind stress curl to the north (Figure 2.1b) drives the anti-clockwise circulation of the Tropical Gyre (Figures 2.1d, 2.1f and 2.2). This tropical Atlantic circulation is bordered to the north and the south by the anti-cyclonic circulation of the north Atlantic and south Atlantic sub-tropical gyres.

The mean wind driven depletion and accumulation of warm surface waters over the tropical Atlantic results

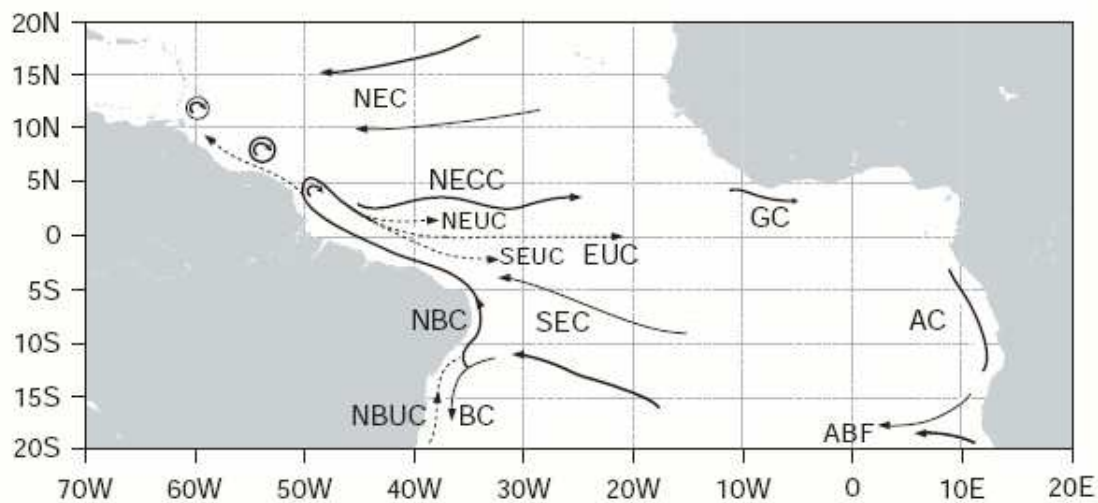


Figure 2.2: A Schematic of the mean features of the tropical Atlantic current system taken from Schouten et al. (2005). Surface currents indicated by the solid lines include the North Equatorial Current (NEC), North Equatorial Counter Current (NECC), Guinea Current (GC), Angola Current (AC), South Equatorial Current (SEC), North Brazil Current (NBC), Brazil Current (BC). Subsurface currents indicated by the dashed lines include the Equatorial Under Current (EUC), North and South Equatorial Under Currents ((N/S)EUC) and North Brazil Under Current (NBUC). Also shown is the position of the Angola–Benguela Front (ABF) and two NBC rings.

in a series of zonal ridges and troughs in SSH superimposed on the rise in SSH from the east to the west (Figure 2.1e, Katz, 1981; Merle and Arnault, 1985). Driven by the band of positive wind stress curl to the north of the mean position of the ITCZ (Figures 2.1b and 2.1c), the cyclonic Tropical Gyre separates the Equatorial Gyre from the North Atlantic sub-tropical gyre. Ekman divergence associated with the positive wind stress curl is responsible for the formation of a trough in SSH which extends zonally across the basin in the region of 10°N (Figure 2.1e, Katz, 1981; Merle and Arnault, 1985). Situated in the middle of the Tropical Gyre, this sea level trough marks the southern boundary of the NEC and the northern boundary of the NECC (Merle and Arnault, 1985). The geostrophic component of the NEC is driven by the pressure gradient established between this trough and higher SSH values to the north (20°N), while the NECC is induced by the meridional pressure gradient between this trough and a zonal ridge in SSH which lies to the south (3°N) (Figure 2.1e, Merle and Arnault, 1985).

The retroflecting NBC extending into the NECC forms the border between the tropical and equatorial wind driven gyres in the region of 5–8°N (Fratantoni et al., 2000). Driven by the band of negative wind stress curl to the south of the mean position of the ITCZ (Figures 2.1b and 2.1c), the Equatorial Gyre includes branches of the westward flowing SEC which, on reaching the coast of Brazil (north of 12°S–14°S), feed into the NBC (Lumpkin and Garzoli, 2005). The NBC current flows northwards along the coast of Brazil, crossing the equator at which point a portion feeds into the EUC, the remainder continues northwestwards, retroflecting in the region of 7°N to feed the NECC. The NECC flows eastward across the Atlantic meeting the Guinea Current (GC) in the Gulf of Guinea. The Equatorial Gyre is then closed in the south by the Angola Current (AC) which flows southward until the Angola-Benguela Front (ABF) (Figure 2.2).

Superimposed upon the clockwise flow of the equatorial gyre and the anti-clockwise flow of the tropical gyre is a net northward flow which forms part of the upper branch of the MOC (Fratantoni et al., 2000).

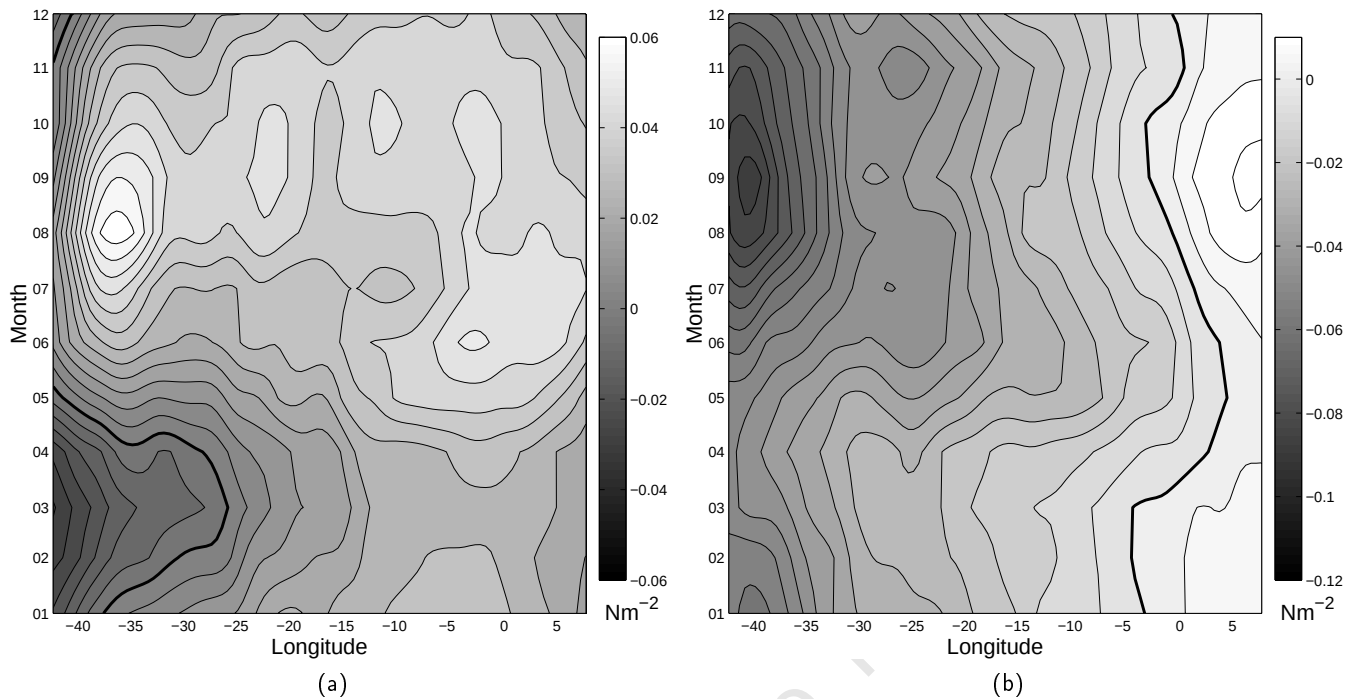


Figure 2.3: This figure depicts the seasonal cycle of (a) meridional wind stress and (b) zonal wind stress along the equator (3°S–3°N) in the Atlantic. COADS data (DaSilva et al., 1994) and a contour interval of 0.005Nm^{-2} has been used. The thick black line represents the 0Nm^{-2} contour line.

2.2 Seasonal Variability

The seasonal cycle is the largest atmosphere-ocean signal in the tropical Atlantic (Weingartner and Weisberg, 1991) and is determined by the response of the coupled land-ocean-atmosphere system to periodic solar forcing. Despite the dominant semi-annual signal in insolation at the top of the atmosphere, equatorial Atlantic atmospheric and oceanic conditions exhibit a distinct annual cycle (Mitchell and Wallace, 1992; Okumura and Xie, 2004). As highlighted by Mitchell and Wallace (1992), atmospheric and oceanic conditions observed at the time of the two equinoxes strongly contrast each other. In March the ITCZ, where the trade winds converge, is closest to the equator as equatorial SSTs are at their warmest and cross-equatorial atmospheric flow is at its weakest. On the other hand in September the equatorial cold tongue is well developed with equatorial SSTs having reached their annual minimum in August. At this time the warmest SSTs together with the ITCZ reside well north of the equator supporting a strong cross-equatorial component in the surface wind field as the southeast trades penetrate into the northern hemisphere. The tropical Atlantic's coupled ocean-atmosphere system oscillates between conditions which are relatively symmetric about the equator in boreal spring (March–April) to those which are the most asymmetric in boreal autumn (September–October) as the ITCZ annually migrates between its southern-most position near the equator in March and April to its northern-most position beyond 8°N during August and September (Mitchell and Wallace, 1992). The associated seasonal cycle in meridional and zonal wind stress along the equator is shown in Figure 2.3.

Arising amidst predominantly semi-annual solar forcing, the dominant annual oscillation in the tropical Atlantic's coupled ocean-atmosphere system described above is seen to be the result of seasonally excited

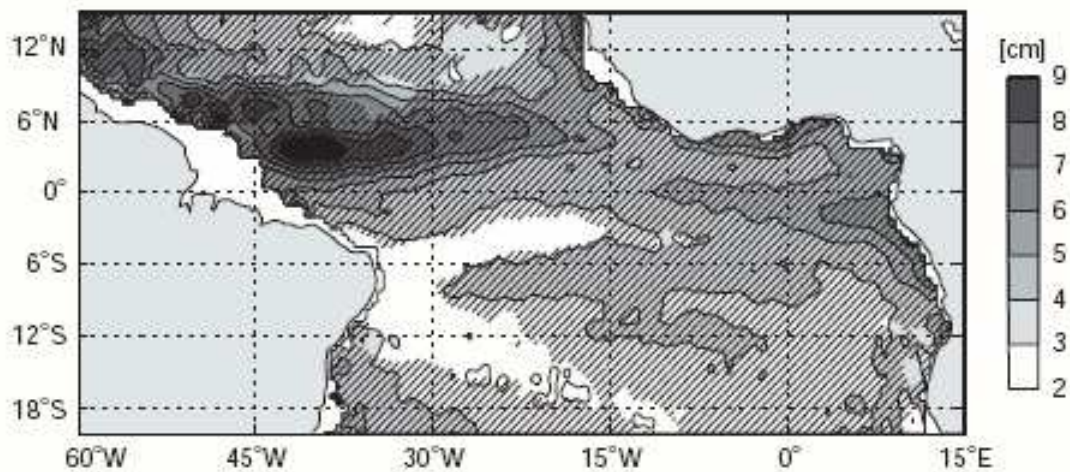


Figure 2.4: Taken from Schouten et al. (2005), this figure shows the standard deviation in cm of tropical Atlantic SSH measured between 1992 and 2001. Hatched regions indicate where the seasonal cycle contributes strongly ($> 50\%$) to the total variability.

ocean-atmosphere instabilities. The annual reestablishment of the equatorial cold tongue commences in April/May when equatorial SSTs start to cool. The initiation of this cooling is attributed to the onset of the West African Monsoon as continental convection moves from the equator into the northern hemisphere accelerating southerly winds over the Gulf of Guinea (Figure 2.3a Mitchell and Wallace, 1992; Okumura and Xie, 2004). These cross-equatorial southerly winds drive upwelling south of the equator cooling the eastern equatorial Atlantic. Chang and Philander (1994) have demonstrated how asymmetric SST-modes act to amplify a perturbation from conditions which are symmetric about the equator into asymmetric conditions. A zero frequency, zero zonal wave number asymmetric SST-mode with a rapid growth rate is accredited with the sudden seasonal growth in asymmetric conditions about the equator in the eastern Pacific and Atlantic, as an unstable cross-equatorial Ekman feedback mechanism leads to the growth of the equatorial cold tongue-ITCZ complex (Chang and Philander, 1994). Asymmetric conditions peak in August/September when the ITCZ is furthest north (Figure 2.3a). Once established, the equatorial cold tongue-ITCZ complex sustains itself even as the region of maximum insolation and continental convection starts to move southwards again. Only towards the end of the southern hemisphere summer does insolation manage to warm south equatorial SST enough to weaken the cold tongue-ITCZ complex allowing the ITCZ to migrate southwards back towards the equator (Chang and Philander, 1994).

Local heat content changes in the equatorial Atlantic are far greater than those which can be explained by seasonal variations in surface heat gain alone (Merle, 1980a; Schouten et al., 2005). Instead, local heat content changes are predominantly forced by the dynamic response of the upper ocean to seasonal winds as warm surface waters are redistributed by seasonal changes in the equatorial currents. The seasonal cycle dominates upper ocean variability in the tropical Atlantic (Merle and Arnault, 1985; Vauclair and du Penhoat, 2001; Schouten et al., 2005). Driven primarily by the dynamic response of the tropical Atlantic to seasonal changes in mechanical wind forcing, large seasonal changes in the distribution of warm surface waters, and hence the local depth of the thermocline, are seen to occur (Merle, 1980a). SSH variations encapsulate this upper ocean response to seasonal changes in regional atmospheric forcing, with SSH generally mirroring seasonal thermocline depth changes in the tropics. Using 10 years of altimeter data, Schouten et al. (2005)

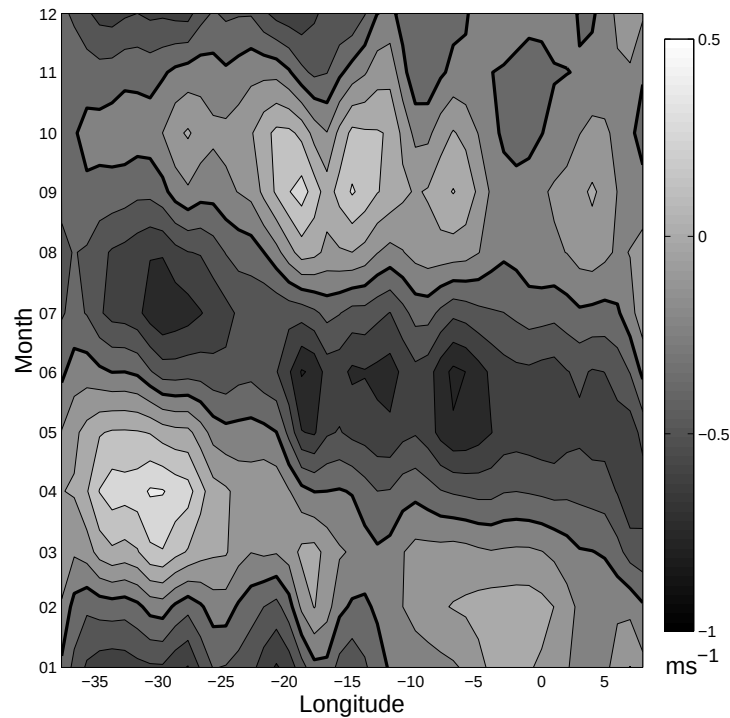


Figure 2.5: Seasonal zonal surface current perturbations from the annual mean along the equator (1°S - 1°N). This figure is based on the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). A contour interval of 0.1 ms^{-1} has been used. The thick black line represents the 0 ms^{-1} contour line.

examined seasonal variability in tropical Atlantic SSH. Between 1992-2001, seasonal variations in tropical Atlantic SSH accounted for nearly 60% of the total variance over most of the tropical Atlantic with peaks of over 80% in some regions (Schouten et al., 2005).

Figure 2.4 taken from Schouten et al. (2005) depicts the standard deviation of tropical Atlantic SSH measured between 1992 and 2001, with the regions where the seasonal contribution is strong ($> 50\%$) marked with hatches. Two distinct regions of high seasonal SSH variability are evident in Figure 2.4, one in the north west tropical Atlantic (NEC/NECC ITCZ region) and the other in the south east equatorial Atlantic (cold tongue region). Maximum seasonal variability occurs in the north west tropical Atlantic under the mean position of the ITCZ's SSH ridge extending zonally in the region of 3°N - 5°N as well as along the equatorial trough (0°S - 3°S) in the east (Gulf of Guinea) extending southward near the African coast (Figure 2.1e). These two regions are separated by an area of comparatively low variability in the central equatorial basin (20°W - 30°W). August-October, when the ITCZ is furthest north, is the period of largest contrast; seasonally SSH values reach their minimum in the south east equatorial Atlantic while they reach their maximum in the north west tropical Atlantic (Merle and Arnault, 1985). Both meridional and zonal SSH gradients are at a maximum during this time, and a minimum during boreal spring, when conditions are the most homogeneous.

Seasonal variability in wind stress forcing over the tropical Atlantic is driven primarily by the annual meridional migration of the ITCZ, where the north east and south east trades converge. The zonal slope in the equatorial Atlantic thermocline is at a minimum in March/April when seasonally the ITCZ is at its southernmost position and zonal winds along the equator are weak (Figure 2.3b). In April/May, the ITCZ starts to

migrate northwards and the south east trades intensify near the equator (Figure 2.3b). The increased westward zonal wind stress appears first in the eastern equatorial Atlantic and moves westward (Figure 2.3b). This increase in zonal wind stress in May causes the westward surface flow associated with the SEC to increase, displacing warm thermocline water from the eastern to the western side of the basin as the thermocline shoals in the east and deepens in the west (Philander and Pacanowski, 1986a). This seasonal acceleration of westward zonal surface currents, which also propagates westward, can be seen in Figure 2.5 which shows the seasonal cycle of observed zonal surface currents along the equator (1°S - 1°N).

Zonal wind stress along the equator continues to increase into June, at which point its behaviour in the east becomes disparate from the west. West of 30°W , where the seasonal variance in zonal winds is the greatest, variability is dominated by an annual harmonic which is in-phase with the annual north-south migration of the ITCZ, peaking in September-October (Figure 2.3b). East of 30°W , a semi-annual signal is dominant with zonal wind stress component peaking in June/July and then again in November/December (Philander and Pacanowski, 1986a) (Figure 2.3b). The meridional component on the other hand exhibits a dominant annual cycle in both the east and the west, increasing as the ITCZ moves north and decreasing as it nears the equator (Figure 2.3a).

Looking at Hovmoeller plots of seasonal SSH changes at the equator (0°N , Figure 2.6a) and off the equator (4°N and 4°S , Figures 2.6b and 2.6c) taken from Ding et al. (2009), variability in the west is clearly dominated by an annual signal, while in the central-eastern basin a semi-annual cycle is superimposed on the annual cycle. On the equator (0°N , Figure 2.6a) seasonal SSH/thermocline depth perturbations are seen to propagate eastward while off the equator (4°N and 4°S , Figures 2.6b and 2.6c) seasonal anomalies propagate westward (Ding et al., 2009). Furthermore a distinct semi-annual signal and westward propagation of seasonal anomalies is seen in equatorial zonal surface currents (not shown, see Ding et al. (2009) Figure 1h).

A strong in-phase relationship exists between seasonal variability in local zonal wind stress forcing and the local upper ocean response in the western equatorial Atlantic. SSH/Thermocline depth fluctuations exhibit a prominent annual cycle, as the thermocline deepens while winds intensify from May to September and shoals as the winds relax from October to April (Figures 2.3b and 2.6a, Merle, 1980a; Philander and Pacanowski, 1986a). The local zonal pressure gradient is seen to be approximately in balance with local wind forcing in the western equatorial Atlantic, with about a 10 day lag between wind changes and local slope changes (Katz, 1987). Philander and Pacanowski (1986a) attribute this near equilibrium response to seasonal wind changes to the fact that the time scale of zonal wind stress variability in the west is comparable to that of the adjustment time scale for the equatorial Atlantic, estimated to be of the order of 150 days (Cane, 1979).

In fact a local oceanic response that is in-phase with local zonal wind fluctuations is observed over most of the basin, with the exception of east of the Greenwich, where the correlation between local winds and oceanic behaviour is poor (Philander and Pacanowski, 1986a; Schouten et al., 2005). East of 30°W a semi-annual component in-phase with the semi-annual signal in zonal wind stress forcing is evident in thermocline depth variability (Figures 2.3b and 2.6a). As zonal wind stress in the east weakens in July/August, the shoaling of the thermocline in the east abates, and it begins to deepen during September and October (Figure 2.7). The local thermocline depth response, in-phase with zonal wind stress variations in both the west and the east during September-October, is made possible by cross-latitudinal fluxes which allow the thermocline to

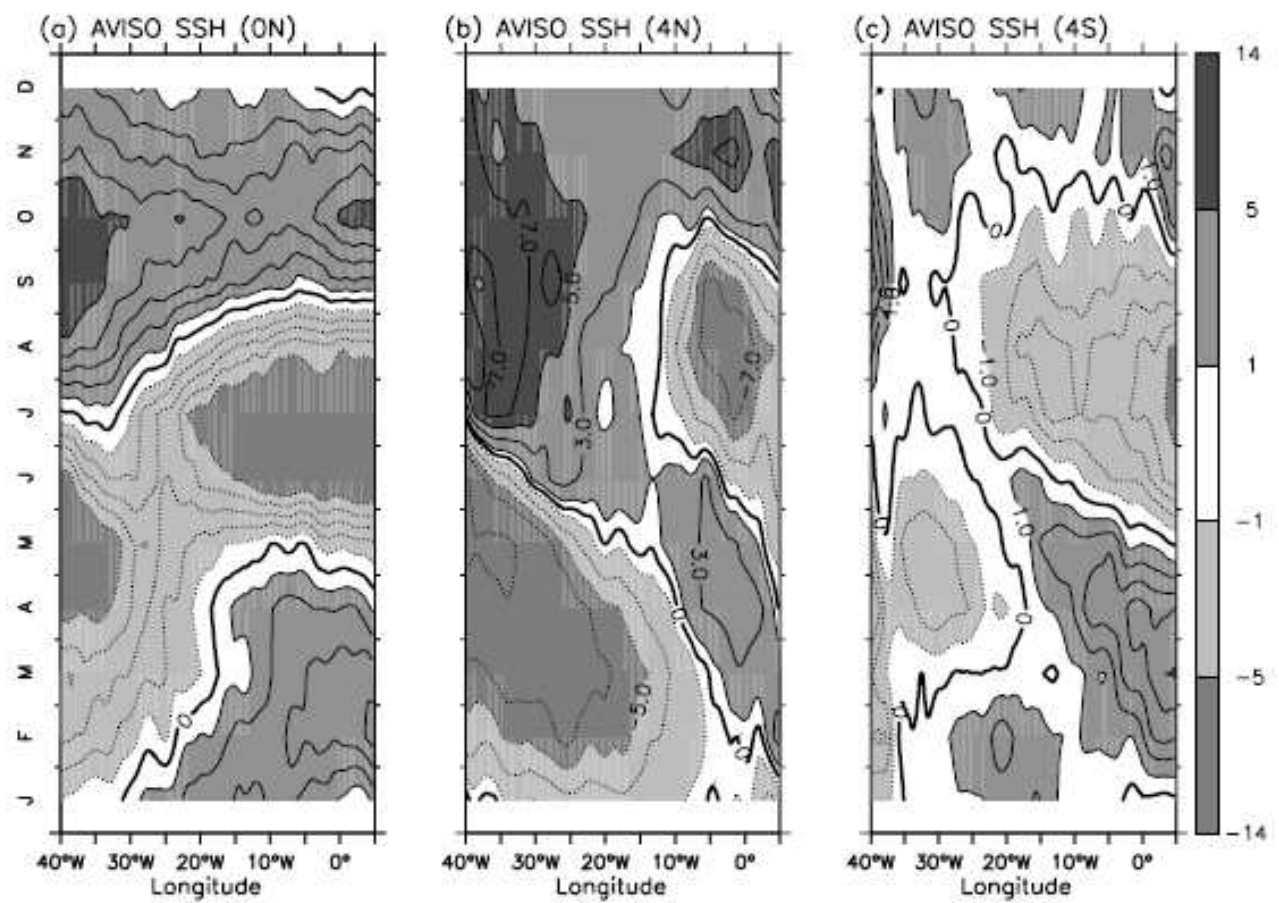


Figure 2.6: Taken from Ding et al. (2009), this figure depicts the seasonal cycle of SSH anomalies with respect to the annual mean: (a) at the equator, (b) at 4°N, and (c) at 4°S. Aviso (www.aviso.oceanobs.com) SSH data has been used. A contour interval of 1cm has been used in (a) and (c) while the -12,-7,-5,-3,-1,0,1,3,5,7,12 contours are shown in (b).

deepen all along the equator during this time as the zonal winds intensity in the west and weaken in the east (Philander and Pacanowski, 1986a). The zonal wind stress component in the east intensifies once more in November/December, causing a second although some what weaker equatorial upwelling event in the east. The thermocline is seen to shoal slightly during this time (Figure 2.7).

Philander and Pacanowski (1986a), suggest that the semi-annual oceanic response in the central-eastern basin is partially driven by a delayed response of the ocean to the sudden intensification of the trades in May-June. East of 30°W, the sudden onset of zonal wind stress forcing in May, due to its semi-annual nature, occurs at a much shorter time scale than the adjustment time scale of the equatorial Atlantic and is therefore seen to excite transients with a phase similar to this semi-annual wind forcing (Philander and Pacanowski, 1986a). The impact of these seasonally excited transients on equatorial surface currents results in the deceleration of westward surface flow in September-October and the subsequent increase in November-December, after which point these transients decay (Philander and Pacanowski, 1986a).

Ding et al. (2009), in a study that employs both observations and a hierarchy of ocean models, have further investigated the importance of transients within the seasonal upper ocean dynamics of the equatorial Atlantic suggested by previous authors (Philander and Pacanowski, 1986a; Schouten et al., 2005). Ding et al. (2009) conclude that the seasonal cycle in equatorial Atlantic SSH/thermocline depth can largely be explained by linear dynamics with a linear solution that essentially consists of the four gravest baroclinic modes. Similarly linear theory also captures the seasonal cycle in equatorial zonal surface currents however the inclusion of non-linear terms significantly improves the solution correcting the strength, phase and zonal extent of the seasonal variability (du Penhoat and Treguier, 1985; Ding et al., 2009). On the equator, the eastward propagation of seasonal SSH/thermocline depth anomalies is primarily attributed to Kelvin waves (Schouten et al., 2005; Ding et al., 2009). The simulated Kelvin wave contribution to the seasonal cycle in equatorial SSH captures the eastward propagation of negative SSH anomalies between January and August, as well as the some what faster eastward propagation of positive SSH anomalies between September and October (Figure 2.6a, Ding et al., 2009). While westward propagation of surface zonal current anomalies on the equator is attributed to first meridional mode Rossby waves as the simulated Rossby wave contribution is seen to out-way the Kelvin wave contribution (Ding et al., 2009). The westward propagation of zonal wind forcing is also thought to reinforce this westward propagation (Ding et al., 2009). Off the equator, the westward propagation of seasonal SSH/thermocline depth perturbations (Figures 2.6b and 2.6c) is attributed to first meridional mode Rossby waves (Ding et al., 2009).

Unlike the Pacific (Yu and McPhaden, 1999), boundary (coastal) reflections are seen to play an important role in seasonal SSH variability in the tropical Atlantic (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009). Seasonally, the contribution of Kelvin and Rossby waves reflected at the eastern and western boundaries to simulated SSH variability is equal to directly forced waves (Ding et al., 2009).

Seasonal equatorial SSH and zonal surface currents variability in the tropical Atlantic possesses characteristics of a basin mode. Firstly, a pronounced semi-annual cycle in equatorial SSH and zonal surface currents even though the semi-annual cycle in surface winds is relatively weak in comparison with annual forcing, and secondly, the in quadrature (90° out of phase) relationship between the zonal SSH gradient and zonal surface currents from March to August. Ding et al. (2009) propose that the semi-annual component in zonal winds excites the second baroclinic basin mode, thus forcing the prominent semi-annual component in SSH and surface zonal currents despite being relatively weak. The second baroclinic basin mode for the

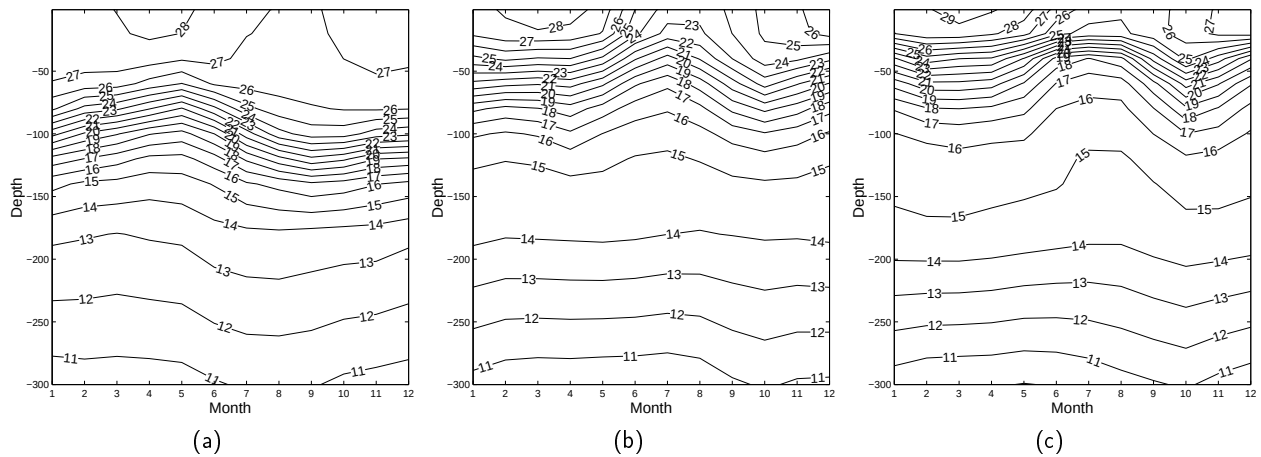


Figure 2.7: A monthly climatology derived from Pilot Research Moored Array in the Tropical Atlantic (PIRATA) (Servain et al., 1998) temperature profiles at three sites on the equator (a) $0^{\circ}\text{N}35^{\circ}\text{W}$, (b) $0^{\circ}\text{N}10^{\circ}\text{W}$ and (c) $0^{\circ}\text{N}0^{\circ}\text{E}$.

Atlantic has a period of 220 days approximately semi-annual (Cane and Moore, 1981; Ding et al., 2009).

Bunge and Clarke (2009) argue that the eastward propagating seasonal signal in equatorial (2°N - 2°S , 42°W - 10°W) Atlantic SSH/ 20°C isotherm depth can be explained by the superposition of two independent modes of variability instead of equatorial wave propagation; firstly, a zonally asymmetric mode representing changes in the slope of the equatorial thermocline which are approximately in balance with those in the equatorial zonal wind stress, and secondly, a zonally symmetric mode representing zonal mean variability in the amount of warm water above the thermocline (2°N - 2°S), that is driven by variability in the cross-latitude flux of warm surface waters forced by off equatorial wind stress curl. Akin to the equatorial heat content recharge-discharge mechanism of Jin (1997), these two modes are seen to be in quadrature (Bunge and Clarke, 2009).

The dynamic response of the upper ocean to changes in seasonal wind forcing involves not only large seasonal changes in the distribution of warm surface waters within the tropical domain (8°N - 8°S), but also in the cross-latitude flux of warm surface waters into/out of this domain as well as the formation of warm surface waters within the domain. Significant seasonal changes in the global upper ocean heat content of the Atlantic region between 8°N - 8°S are seen to occur due to seasonal changes in warm water formation and escape process (Lee and Csanady, 1999). Merle (1980a) was amongst the first to point out the large seasonal changes in the heat content of the upper Atlantic within the zonal band between 6°N - 6°S , with the mean depth of the equatorial thermocline reaching a maximum in October-November and minimum in June-July. The region imports heat from July to October, but for the rest of the year it exports heat resulting in a net annual export of heat to the north (Merle, 1980a).

As the entrainment rate of water from the thermocline layer into the surface mixed layer (warm water formation) is too small to measure, statements made below are based on the model results of Lee and Csanady (1999). Entrainment occurs in the central to eastern basin where the thermocline is relatively shallow. Seasonally the entrainment rate on the equator ceases between January-April and is significant from May-December. In May, the entrainment rate suddenly increases in the eastern basin and quickly extends across the basin to 25°W , peaking in June due to enhanced wind and current shear. Thereafter, the

entrainment rate decreases and retreats to the east before ceasing in January/February (Lee and Csanady, 1999).

The escape of warm surface waters to the north is seasonally regulated by what is referred to as the heat capacitor mechanism (Philander and Pacanowski, 1986b). The escape of heat across 8°N is regulated by seasonal changes in the NBC/NECC current system in response to changes in wind stress forcing associated with the annual migration of the ITCZ. The escape of warm water takes place via two main pathways. Firstly, north-west transport along the South American coast via the NBC-Guiana current system from December-May which, reduces considerably in June as the NECC intensifies and the NBC veers offshore to feed the NECC. By August northward transport from the NBC to the Guiana current almost vanishes. From October the NECC weakens causing the northward escape of warm water via this route to increase again, reaching a maximum in February (Philander and Pacanowski, 1986a; Johns et al., 1998; Lee and Csanady, 1999). The second escape route is via the interior of the tropical Gyre which transports warm water to the NEC between December-May. This northward transport is blocked during June-September as increased wind stress curl causes the mixed layer to deepen south of the NECC (8°N) and shoal to the north (Philander and Pacanowski, 1986b; Lee and Csanady, 1999).

While large seasonal changes are seen in heat loss across 8°N , seasonal variations across 15°N are modest. The zonal bands between 5°S - 8°N and 8°N - 15°N act like out of phase capacitors (Philander and Pacanowski, 1986b). From July-September heat is stored in the 5°S - 8°N band deepening the thermocline. The export of heat across 8°N is minimal at this time and the thermocline between 8°N - 15°N shoals sustaining heat loss across 15°N . Then during the months of December-May when heat transport across 8°N is significant, it transfers heat from the 5°S - 8°N band replenishing the 8°N - 15°N band as the thermocline deepens there during this time (Philander and Pacanowski, 1986b).

2.3 Inter-annual Variability

Unlike in the Pacific where inter-annual coupled variability is predominantly driven by dynamic air-sea interactions, both dynamic and thermodynamic local air-sea interactions give rise to inter-annual coupled variability in the Atlantic (Chang et al., 2006a). In the Pacific, the majority of inter-annual coupled climate variability is associated with a single mode focused on the equator, whereas in the tropical Atlantic, two distinct modes of inter-annual ocean-atmosphere variability have been identified in the literature, namely the “meridional mode” and the “zonal mode” (Chang et al., 2006a).

Figure 2.8 taken from Kushnir et al. (2006) shows the anomalous SST, wind and rainfall patterns associated with the meridional and zonal modes. Like ENSO, the zonal mode is equatorially focused and attributed to dynamic air-sea inter-actions, while the meridional mode is associated with off-equatorial tropical SST variability and is attributed to thermodynamic air-sea interactions (Chang et al., 2006a).

Although upper ocean dynamics are believed to play a role in the decay of meridional mode events (Barreiro et al., 2005), the growth of meridional mode events is attributed to a thermodynamic feedback mechanism. As illustrated by Figure 2.8a, the meridional mode is defined by an anomalous cross-equatorial SST gradient (Servain et al., 1999). This anomalous SST pattern gives rise to cross-equatorial winds which in-turn reinforce the anomalous SST gradient through the thermodynamic WES feedback. A meridional displacement of the ITCZ from its climatological position towards the warmer hemisphere and rainfall anomalies

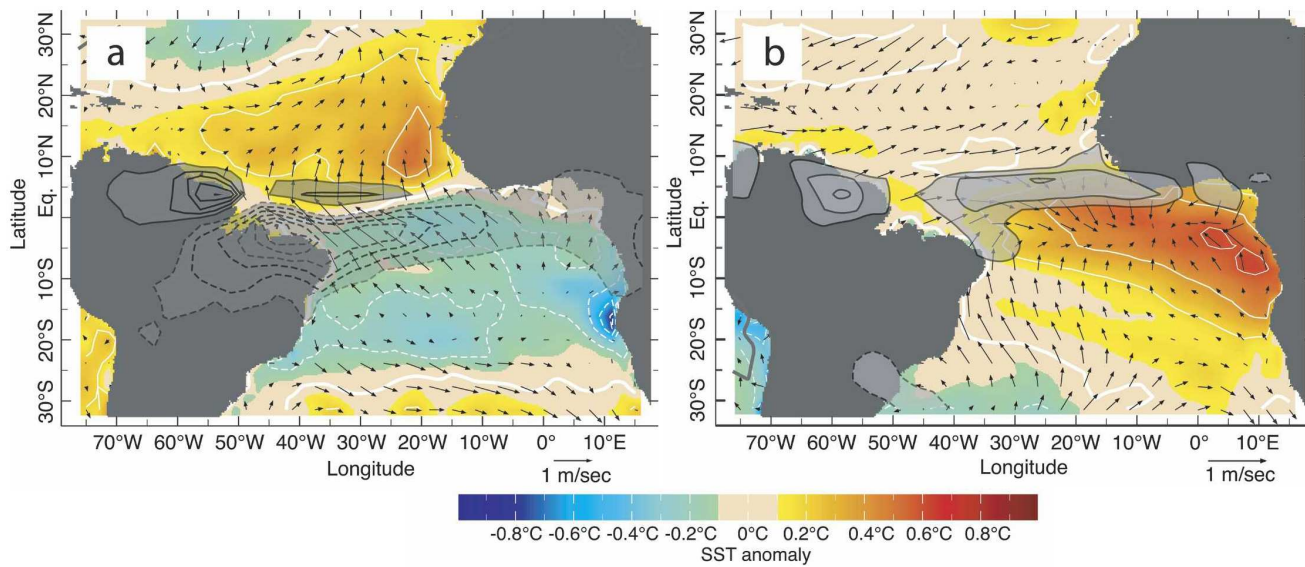


Figure 2.8: Taken from Kushnir et al. (2006), this figure depicts the dominant pattern of surface ocean-atmosphere variability in the tropical Atlantic region during (a) boreal spring (March-April) and (b) boreal summer (June-August). The black contours show the first Empirical Orthogonal Function (EOF) of the regional rainfall anomalies for each period from Global Precipitation Climatology Project data (1979-2001) (explaining 33% and 23% of the variance respectively). A contour interval of 0.5 mm day^{-1} has been used, positive contours are solid, negative contours are dashed, and the 0 contour is omitted. The coloured shading depicts the associated March-April and June-August SST anomalies (SST anomalies regressed against the rainfall EOF's principal component time series), with units given on the scale below in $^{\circ}\text{C}$, plus white contours every 0.2°C for clarity. Arrows depict the associated March-April and June-August surface wind anomalies (wind anomalies regressed against the rainfall EOF's principal component time series), with units in ms^{-1} (see arrow scale below each figure).

over northeast Brazil are seen to be strongly correlated with the SST anomaly pattern associated with the meridional mode (Wallace et al., 1998). Meridional mode events are seen to peak in boreal spring (Figure 2.8a).

Inter-annual variability in central-eastern equatorial Atlantic SSTs is primarily attributed to the zonal mode (Figure 2.8b). The zonal mode is often referred to as the “Atlantic Niño” mode because its spatial signature resembles the Pacific’s ENSO (Merle, 1980b; Hisard, 1980; Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas et al., 2000). As illustrated by Figure 2.8b, the zonal mode, like ENSO, displays a SST-wind relationship whereby a decrease (increase) in the equatorial trades is associated with positive (negative) SST anomalies in the eastern equatorial Atlantic (Merle, 1980b; Servain et al., 1982; Zebiak, 1993; Ruiz-Barradas et al., 2000; Kushnir et al., 2006). The leading EOF for anomalous precipitation during boreal summer is shown in Figure 2.8b with maximum anomalous precipitation seen along the northern coastline bordering the Gulf of Guinea (Kushnir et al., 2006).

As this study is aimed at establishing the role of upper ocean dynamics within coupled tropical Atlantic climate variability, the focus here falls on the zonal mode rather than the meridional mode, as ocean dynamics are credited with the development of zonal mode events.

Using available observations, Zebiak (1993) compared the relationship between wind stress and SST anomalies in the equatorial Atlantic with that observed in the equatorial Pacific. To facilitate this comparison, Zebiak (1993) defined the Atl3 index (3°S - 3°N 20°W - 0°E) based on the spatial structure of observed SST variance. Correlations between equatorial wind stress and the Atl3 index revealed a common SST-wind relationship to that observed in the Pacific. Zebiak (1993) found this to be strong evidence that an equatorial coupled mode driven by dynamical processes similar to ENSO exists in the Atlantic. A more recent analysis (Ruiz-Barradas et al., 2000), of not only the relationship between observed equatorial SST and wind stress but also oceanic heat content and atmospheric convection, corroborates the suggestion that dynamical processes similar to ENSO are acting in the Atlantic. Anomalous warm SSTs in the eastern equatorial Atlantic are seen to be associated with oceanic heat content anomalies (Vauclair and du Penhoat, 2001; Vauclair et al., 2004) as well as anomalously strong convection near the equator (Carton and Huang, 1994). Analogous to the Pacific, the Bjerknes feedback mechanism (Bjerknes, 1969) is believed to be behind the growth of the zonal mode events (Carton and Huang, 1994; Keenlyside and Latif, 2007; Chang et al., 2000).

While the spatial signature of SST and wind variations associated with the equatorial zonal mode clearly resembles that of ENSO, the observational analysis of Zebiak (1993) also revealed several important differences between the nature of the zonal mode and ENSO. The most apparent is that the anomalous wind-SST relationship observed in the Atlantic is weaker and less consistent (Zebiak, 1993; Ruiz-Barradas et al., 2000). The analysis of Zebiak (1993) revealed a correlation of 0.7 between inter-annual zonal wind anomalies and SST anomalies in the Pacific and only 0.4 in the Atlantic. While ENSO accounts for the majority of inter-annual SST variability in the Pacific, the zonal mode explains a smaller percentage of the observed variability. Regression values obtained by Keenlyside and Latif (2007) based on observational data suggest that the wind anomaly associated with a 1°C SST anomaly is less in the Atlantic than in the Pacific and while this relationship explains 20% of the variance in the Pacific it only explains 10% in the Atlantic. The relationship between eastern basin SST anomalies and western basin wind anomalies appears to be weaker in the Atlantic. Furthermore, the relationship between inter-annual SST anomalies and thermocline depth anomalies in the equatorial Pacific is far more robust than in the equatorial Atlantic (Zebiak, 1993;

Vauclair and du Penhoat, 2001; Keenlyside and Latif, 2007). Not all inter-annual SST anomalies in the eastern-central equatorial Atlantic are associated with thermocline depth anomalies (Carton and Huang, 1994). While some warm SST events in the eastern tropical Atlantic have been attributed to anomalous vertical temperature gradients due to a change in the east-west slope of the thermocline, others have not exhibited this El Niño like characteristic.

As a result of the relative weakness and inconsistency of variability affiliated with the zonal mode, confirming the presence of the Bjerknes feedback has been more difficult than in the Pacific (Chang et al., 2006a). The evolution of anomalous SST events following the Bjerknes feedback mechanism is less apparent in the Atlantic. Nevertheless, a recent study by Keenlyside and Latif (2007), based on observational evidence, offers strong evidence that all three elements of the Bjerknes feedback are active inter-annually in the equatorial Atlantic: 1) an atmospheric response to eastern equatorial SST anomalies that results in wind anomalies to the west, 2) the creation of heat content anomalies in the east by wind anomalies in the west, 3) the translation of thermocline depth anomalies in the central-eastern equatorial basin into SST anomalies. Keenlyside and Latif (2007) confirm that the Bjerknes feedback explains less of the inter-annual variability than in the Pacific.

This difference between the zonal mode and ENSO has been attributed to a larger contribution from external forcing (Chang et al., 2000) and disparate local air-sea interactions in the tropical Atlantic (Zebiak, 1993; Chang et al., 2006a). Whereas inter-annual SST anomalies in the Pacific are predominantly forced by the dynamic Bjerknes feedback mechanism associated with ENSO, inter-annual SST anomalies in the equatorial Atlantic may be the result of local thermodynamic (WES Feedback) or local dynamical (Ekman Feedback) feedbacks as well as the remotely forced Bjerknes feedback (Chang et al., 2006a).

This complexity has made diagnosing the role of ocean dynamics and in particular the role of ocean memory within zonal mode events somewhat more difficult than in the case of the Pacific. Although it is of equal importance as the Bjerknes mechanism, the delayed negative ocean feedback has received much less attention and its role within zonal mode events is poorly understood (Chang et al., 2006a; Keenlyside and Latif, 2007).

Another notable difference between the nature of the zonal mode and ENSO is that the Atlantic has more high frequency variability (shorter events) with larger contributions at seasonal time scales (Zebiak, 1993; Latif and Grötzner, 2000). With no significant spectral peak at inter-annual time scales, spectral characteristics are hardly discernible from red noise. The zonal mode is therefore described as stable (damped) and noise driven (Latif and Barnett, 1995; Nobre et al., 2003; Illig and Dewitte, 2006; Kushnir et al., 2006; Keenlyside and Latif, 2007).

Furthermore, the zonal mode and ENSO exhibit a very different relationship with the seasonal cycle, SST anomalies associated with ENSO events peak in boreal winter, while those associated with the zonal mode peak in boreal summer (Zebiak, 1993; Latif and Grötzner, 2000; Keenlyside and Latif, 2007).

As highlighted by Chang et al. (2006a), literature surrounding the remote influence, via atmospheric teleconnections, of ENSO on the zonal mode is inconsistent. While some observationally based analyses finds equatorial Atlantic SST variability associated with the zonal mode to be poorly related to ENSO (Zebiak, 1993; Enfield and Mayer, 1997), other studies suggest that some zonal mode events are linked to ENSO events (Delecluse et al., 1994; Carton and Huang, 1994; Latif and Barnett, 1995). Chang et al. (2006a) suggest that one of the reasons the zonal mode might be less influenced by the remote influence of ENSO

- as well as the North Atlantic Oscillation (NAO) - than the meridional mode is because the zonal mode peaks in boreal summer while variability associated with ENSO and the NAO peaks in boreal winter.

The most robust influence of ENSO on tropical Atlantic SST is seen in the northern tropical Atlantic and is attributed to the “tropospheric temperature mechanism” which is most effective between later boreal winter when ENSO events peak and early spring (Chiang and Sobel, 2002; Chiang and Lintner, 2005; Chang et al., 2006b). During an El Niño event anomalous warming in the eastern equatorial Pacific is associated with increased atmospheric heating and a tropospheric warming signal that propagates eastward to the tropical Atlantic in the form of an equatorial Kelvin wave. This tropospheric warming acts to promote stable atmospheric conditions and reduced evaporation at the ocean surface resulting in warm SST anomalies (Chang et al., 2006b).

The response of the equatorial and southern tropical Atlantic to remote ENSO forcing is less statistically significant than that of the northern tropical Atlantic (Chang et al., 2006b). The South Atlantic subtropical anticyclone is generally weaker than usual during and following the mature phase of El Niño events (Colberg et al., 2004). Chang et al. (2006b) suggest that the cause of the fragile relationship between ENSO and equatorial Atlantic SST variability is due to observed destructive interference by local coupled variability. Whether an El Niño event results in anomalously cool or anomalously warm SST conditions in the central-eastern equatorial Atlantic depends on the relative contribution of warming by the “tropospheric temperature mechanism” versus local dynamic ocean–atmosphere feedback induced cooling (Chang et al., 2006b). The local dynamic ocean–atmosphere feedback induced cooling appears to be due to the appearance of anomalous winds in the western equatorial Atlantic in response to some El Niño events (Latif and Grötzner, 2000; Chang et al., 2006b). The reason for the appearance of anomalously strong trades in the western Atlantic in response to some El Niño events and not others is unclear. Chang et al. (2006b) suggest that the development of anomalously strong trades in response to El Niño events may depend on the nature of SST conditions prior to the El Niño event, stochasticity of the atmosphere, and/or variability in the structure and duration of the atmospheric heating associated with individual El Niño events.

Okumura and Xie (2006) have highlighted the presence of a second mode of equatorially focused coupled variability associated with inter-annual variability in the second, some what smaller, seasonal cooling which occurs in the central equatorial Atlantic when easterly winds intensify between November and December. The easterly intensification of equatorial easterlies in November increases upwelling and shoals the thermocline slightly in the Gulf of Guinea (Figure 2.7). This seasonal equatorial cooling event lasts only slightly more than a month and is confined to a much smaller region than that of the equatorial cooling associated with the seasonal development of the equatorial cold tongue in boreal summer. High-resolution climatological data is needed to resolve this SST cooling which is confined to the central equatorial Atlantic (2°S – 2°N , 14° – 22°W).

This second mode of coupled equatorial Atlantic variability is referred as the November-December zonal mode and the associated warm events are referred to as Atlantic Niño II events (Okumura and Xie, 2006). Okumura and Xie (2006) identify Atlantic Niño II events according to the Atlantic Niño II index which is defined as November-December SST anomalies averaged over the central equatorial Atlantic (3°S – 3°N , 5° – 15°W). Anomalous oceanic warming is associated with an increase in equatorial precipitation and relaxed easterly trades. Okumura and Xie (2006) suggest that the Bjerknes feedback is associated with the generation of November-December zonal mode events based on lag correlations between SST, SSH and zonal wind.

Although anomalous November-December central equatorial Atlantic SST events are moderate, especially in comparison to the boreal summer central equatorial Atlantic SST events, the results of Okumura and Xie (2006) suggest that the atmosphere is sensitive to these small changes. November-December zonal mode events are short in comparison to ENSO and boreal summer zonal mode events. Furthermore, November-December zonal mode events are not necessarily an extension of boreal summer zonal mode events. Although in some years November-December SST anomalies in the central equatorial Atlantic are of the same sign as those which occurred the previous summer in other years they are the reverse. Between 1981 and 2004, variability in November-December zonal mode is statistically independent from both that of the preceding summer's zonal mode and the Pacific ENSO (Okumura and Xie, 2006).

Okumura and Xie (2006) highlight that the November-December peak in inter-annual SST associated with the November-December zonal mode is not evident in some reconstructed SST products. Furthermore they point out that prior to 1980, the November-December peak is hardly distinguishable from the decaying signal of the summer Atlantic Niño, which they speculate could be the result of insufficient observations, or long-term climate change, or a combination of both these factors. Okumura and Xie (2006) also suggest that the impact of the Atlantic Niño II extends beyond the equatorial Atlantic.

Along the west coast of southern Africa, episodes of anomalously warm and cold sea surface temperature, referred to as Benguela Niños and Benguela Niñas respectively, represent an important mode of variability in the Benguela upwelling system (Shannon et al., 1986; Florenchie et al., 2004). These events, which occur in boreal spring in the region of the ABF, affect rainfall variability within the region (Rouault et al., 2003) as well as impact local fish populations (Boyer et al., 2001). The occurrence of a Benguela Niños/Niñas is often linked to the occurrence of a zonal mode event (Reason et al., 2006; Lübbecke et al., 2010). Like zonal mode Atl3 SST events, Benguela Niño/Niña SST events - occurring in the Angola-Benguela Area (ABA 10°S-20°S and 8°E -15°E as defined by Florenchie et al. (2003)) - appear to be generated not locally but by zonal wind stress anomalies in the western equatorial Atlantic (Florenchie et al., 2003, 2004; Rouault et al., 2007). Benguela Niños/Niña events are seen to be the surface expression of subsurface temperature anomalies that are generated by wind stress fluctuations in the western equatorial Atlantic and propagate as equatorially and coastally trapped Kelvin waves eventually outcropping on reaching the ABFZ (Florenchie et al., 2003, 2004; Rouault et al., 2007; Lübbecke et al., 2010). An interesting aspect of ABA SST events is that while they are largely linked to zonal mode Atl3 SST events due to coherent forcing, anomalous ABA SST events occurring in boreal spring tend to precede boreal summer Atl3 SST events (Rouault et al., 2009; Lübbecke et al., 2010). Recently reasons for this seemingly inconsistent behaviour have been explored by Lübbecke et al. (2010). They find the cause of the time-lag between ABA and Atl3 SST anomalies to be related to differences in the depth of the thermocline between each region as well as differences in the seasonal phase-locking of the inter-annual SST variations. Anomalous Atl3 events peak in boreal summer as the influence of remotely forced thermocline depth variations on Atl3 SST is strongest in June/July when seasonally the thermocline in the Atl3 region is at its shallowest. In the ABA region, SST anomalies peak when seasonally the ABFZ is at its southern most position and inter-annual variability in the strength of Kelvin wave variability is at its highest. Lübbecke et al. (2010) refer to combined zonal mode-Benguela Niño warm events as eastern tropical Atlantic Niños which start with a weakening of the south-east trades linked to fluctuations in the strength of the South Atlantic subtropical anticyclone. They suggest that relaxed trades excite eastward equatorial Kelvin waves which first influence SSTs in the ABA region where the thermocline outcrops in March/April and only later in June/July does the seasonal shallowing of the

thermocline allow subsurface anomalies to be reflected in the Atl3 SST field.

Seasonally, the decay of meridional mode events in April coincides with the onset period of zonal mode events (Chang et al., 2006a). Based on observational indices from 1980-1997, Servain et al. (1999) show that there is a significant correlation between these two climatic modes of variability. They propose that both modes are associated with anomalous latitudinal shifts in the position of the ITCZ, with the strongest relationship observed between the meridional mode and anomalies in the position of the ITCZ. Using a oceanic general circulation model, results obtained by Servain et al. (2000) corroborate the observationally based results obtained by Servain et al. (1999). They suggest that the zonal and meridional modes have the same physics as annual variability and are related by the fact that they are both a result of anomalous fluctuations in the trade wind system associated with latitudinal displacements of the ITCZ. Servain et al. (2000) propose that dynamic ocean processes as opposed to thermodynamic are the principal cause of climate variability within the tropical Atlantic. However, Chang et al. (2006a) draw attention to two potential caveats in the results obtained by Servain et al. (1999, 2000). Firstly, that the strong relationship between the meridional mode and the zonal mode obtained may have only occurred within the 1980s and 1990s. Results obtained by Murtugudde et al. (2001), based on a simulation of oceanic conditions between 1949-2000 forced with the NCEP re-analysis winds, suggest that the correlation between the zonal and meridional modes was much lower prior to the 1980s. Secondly, the relationship between inter-annual SST anomalies and thermocline depth anomalies in the equatorial Atlantic is far less robust than in the Pacific; therefore, the strong relationship observed between anomalies in the cross-equatorial SST gradient (the index used for the meridional mode) and anomalies in the zonal slope of 20°C isotherm at the equator (the index used for the zonal mode) does not necessarily mean that there is a strong relationship between the meridional mode and equatorial SST anomalies. Chang et al. (2006a) highlight the fact that much remains to be investigated in terms of the relationship between the two modes, particularly their relationship during boreal spring and summer months when they coexist.

2.4 Summary

In this chapter the current understanding of mean, seasonal and inter-annual conditions within the tropical Atlantic has been reviewed. As highlighted in Chapter 1 several questions surrounding seasonal and inter-annual variability within the tropical Atlantic remain. In the following chapters (Chapters 3, 5 and 6) these questions are tackled.

Chapter 3

Equatorial Variability in the Atlantic versus the Pacific

3.1 Introduction

In the Pacific a wealth of studies have led to a comprehensive understanding of the role of dynamic ocean processes in driving SST variability at both the seasonal and inter-annual time scales. As detailed in section 3.3.2, a clear distinction exists in the central-eastern equatorial Pacific between the mechanisms which drive inter-annual SST variability and those controlling seasonal SST changes. Seasonal SST variability in the eastern-central Pacific is attributed to seasonally excited SST-modes, and their associated Ekman feedback, while inter-annual variability is primarily attributed to the delayed oscillator mechanism and the associated Bjerknes feedback. The same can not be said for the equatorial Atlantic. Relative to the Pacific, the role of the ocean within equatorial Atlantic SST variability is poorly understood (Chang et al., 2006a).

In Chapter 1, several questions were raised involving the role of the ocean within equatorial Atlantic SST variability:

- What distinct processes occurring in the Atlantic may account for the observed differences in the seasonal central-eastern basin SST signature between the Pacific and the Atlantic?
- Why do inter-annual SST fluctuations dominate over seasonal fluctuations in the central-eastern Pacific while the opposite is true in the Atlantic? Is it because, unlike in the Pacific, the Bjerknes mechanism acts primarily on seasonal time scales in the Atlantic?
- Can inter-annual SST variability in the central-eastern equatorial Atlantic be explained in terms of a modulation of seasonally active coupled variability and, if so, what factors are responsible for the modulation of seasonally active processes?
- What is the role of ocean memory within central-eastern equatorial Atlantic SST variability?

In this chapter the first three questions are broadly addressed by establishing the relationship between remotely forced ocean dynamics and central-eastern basin equatorial Atlantic SST variability at both the seasonal and inter-annual time scales. This objective is achieved by capitalising on the insight gained in the

Pacific and contrasting the nature of variability in key indicators (SST, thermocline depth and zonal wind stress) in the Pacific against the Atlantic.

The character of remotely forced thermocline depth variability in the equatorial basin is evaluated here using APE, as APE succinctly quantifies the response of the upper equatorial ocean to large scale wind forcing. The APE within a chosen oceanic volume is the amount of Potential Energy (PE) available to the region through the adiabatic horizontal redistribution of mass (Lorenz, 1955), APE therefore concisely evaluates the generation and evolution of thermocline perturbations due to the horizontal redistribution of warm surface waters in response to changes in wind forcing.

Inter-annual SST fluctuations in the central-eastern equatorial Pacific are highly anti-correlated with the amount of gravitational APE in the equatorial Pacific (Goddard and Philander, 2000). This inverse relationship arises from the fact that APE effectively represents changes in east-west slope of the equatorial thermocline. During La Niña conditions the east-west slope of the thermocline is at a maximum together with the amount of APE in the equatorial Pacific, while during El Niño conditions, APE and the slope of the thermocline are at a minimum. Thus, the equatorial Pacific ocean gains APE during the transition from El Niño to La Niña and loses APE during the transition from La Niña to El Niño. Inter-annual SST anomalies in the Pacific can therefore be regarded as surface expressions of upper ocean energy changes (Goddard and Philander, 2000).

Results obtained in this chapter reveal that as in the Pacific a strong SST-APE relationship is seen in the Atlantic. In both basins, SST fluctuations in the eastern-central equatorial basin are highly anti-correlated with equatorial APE. The biggest distinction between these two basins exists in the period of dominant APE fluctuations. While the largest changes in APE occur inter-annually in the Pacific, they occur annually in the Atlantic.

The strong correlation between large seasonal fluctuations in APE and central-eastern Atlantic SST revealed implies that remotely forced thermocline variability, a process which only plays a significant role inter-annually in the Pacific, plays a key role in seasonal SST variability in the Atlantic. This seasonal upper ocean response appears to involve more than a passive response to seasonal wind forcing and evidence is offered in this chapter to support the hypothesis that the Bjerknes feedback mechanism is excited seasonally and plays an active role in the development of the Atlantic cold tongue between April and August.

Furthermore, results presented in this chapter, as well as the Chapter 6, argue that inter-annual SST variability associated with the zonal mode is best understood as the modulation either in phase or amplitude of this seasonally active process as opposed to a dynamically distinct natural mode of variability operating on inter-annual time scales as is the case with ENSO. This analysis suggests that anomalous SST events in the Atl3 region forced by the Bjerknes feedback mechanism occur predominantly during boreal summer as a modulation of an already seasonally active Bjerknes feedback. Differences between ENSO and the Zonal mode mentioned in Chapter 2 can then be accounted for in terms of this distinction as elaborated on in the final chapter (Chapter 7).

In the following section (Section 3.2), the data employed in the analysis undertaken in this chapter as well as the method chosen to evaluate APE are detailed. Thereafter Section 3.3 deals with the results of the analysis wherein the relation between equatorial Atlantic APE, central-eastern basin SST and equatorial wind stress at both seasonal and inter-annual time scales is evaluated and compared with the Pacific. Finally Section 3.4, summarises the conclusion drawn based on the analysis outlined in Section 3.3.

3.2 Data and Methodology

3.2.1 Global Ocean Models

Density values based on monthly temperature and salinity fields from the $1/2^\circ$ Simple Ocean Data Assimilation (SODA) re-analysis product (Carton et al., 2000; Carton and Giese, 2008), and the $1/4^\circ$ ORCA025-G70 inter-annual global ocean simulation (Molines et al., 2006), both spanning 47 years (1958-2004), have been used to assess seasonal and inter-annual variability in both the equatorial Atlantic and equatorial Pacific APE. The reason for basing the analysis on not one but two global ocean datasets is twofold; firstly, to establish the robustness of results across different ocean models, and secondly, to compare results obtained based on a simulated versus an assimilated model as energy is not strictly conserved in assimilated models. A non-linear equation of state (Jackett and McDougall, 1995) has been used when deriving density values from SODA and ORCA temperature and salinity data.

Figure 3.1 depicts the relationship between inter-annual anomalies in the Niño 3, Niño 3.4, and Atl3 SST based on observed Hadley OI SST data, SODA SST data and ORCA SST data. Both the SODA and ORCA datasets capture a large amount of the observed inter-annual variability, with correlation coefficients of $r=0.92$ ($r^*=0.25$), $r=0.92$ ($r^*=0.25$) and $r=0.86$ ($r^*=0.19$) between observed Hadley OI and SODA Niño 3, Niño 3.4, and Atl3 SST respectively (Figure 3.1) and correlation coefficients of $r=0.97$ ($r^*=0.25$), $r=0.97$ ($r^*=0.25$) and $r=0.87$ ($r^*=0.19$) between observed Hadley OI and ORCA Niño 3, Niño 3.4, and Atl3 SST respectively (Figure 3.1).

Figure 3.2 depicts the relationship between climatological Niño 3, Niño 3.4, and Atl3 SST based on observed Hadley OI SST data, SODA SST data and ORCA SST data. Observed seasonal fluctuations in Niño 3, Niño 3.4, and Atl3 SST are well represented in the SODA and ORCA datasets, except that a warm bias of 1°C is seen between observed Hadley OI Atl3 SST and simulated ORCA Atl3 SST. Although the unassimilated ORCA output reproduces both seasonal and inter-annual fluctuation in observed central-eastern Pacific and Atlantic SST as well as the assimilated SODA output, it shows a warm bias particularly in the central-eastern Atlantic.

It should be noted that in the case of a forced global ocean simulation, SST is not the best variable with which to assess model skill. SST in the model is implicitly restored to observed SST through the bulk formulation of surface fluxes and therefore is not a completely independent variable. The SST comparisons made here serve however to provide confirmation that these datasets have reproduced observed SST variability adequately enough to support the analysis conducted within this section.

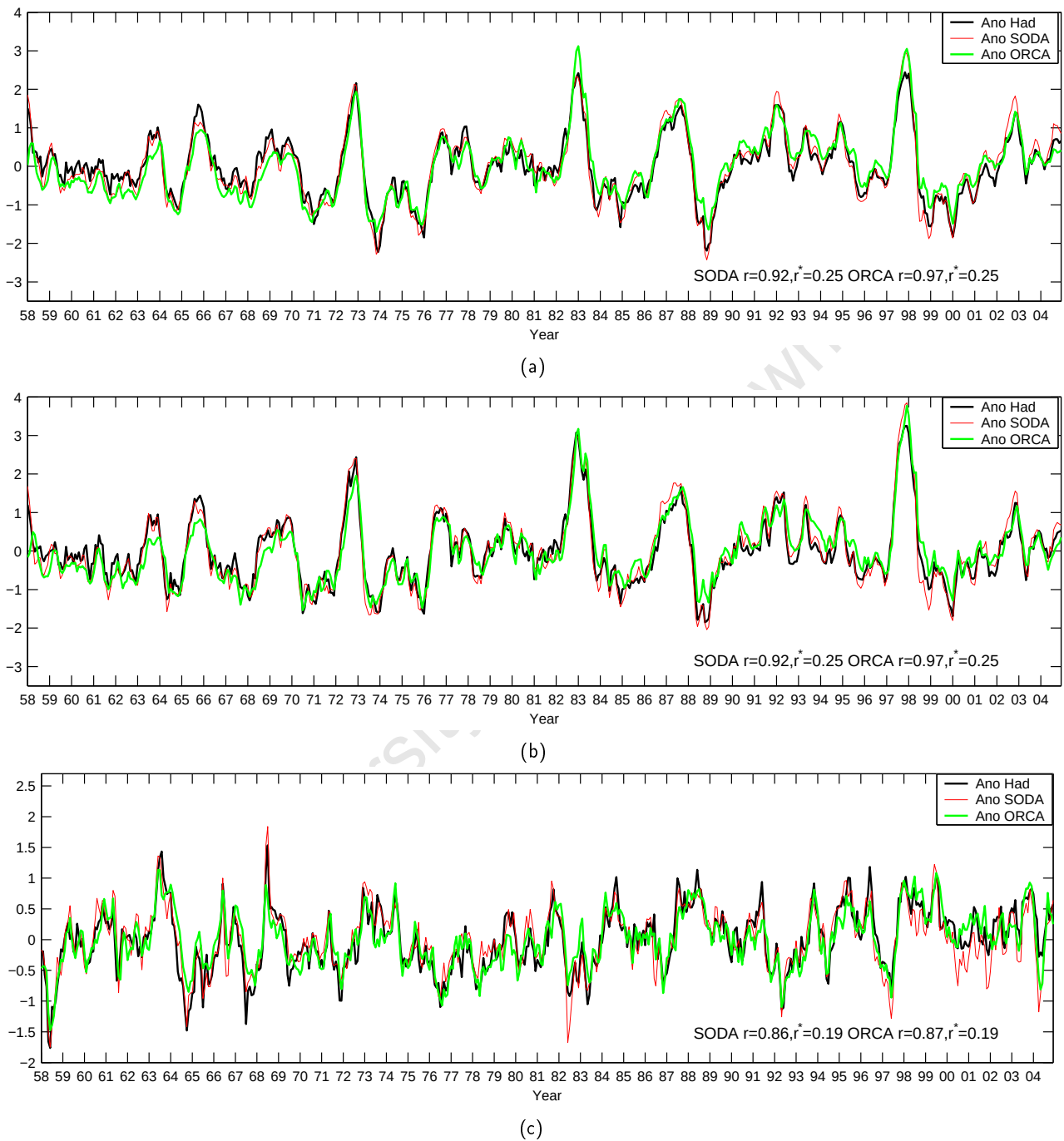


Figure 3.1: A comparison between inter-annual anomalies in (a) Niño 3.4, (b) Niño 3, and (c) Atl3 SST, based on observed Hadley OI SST data (Rayner et al., 2003) and simulated SODA and ORCA SST data. SST anomalies are shown in °C. The correlation between observed and simulated anomalies in each SST index is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*).

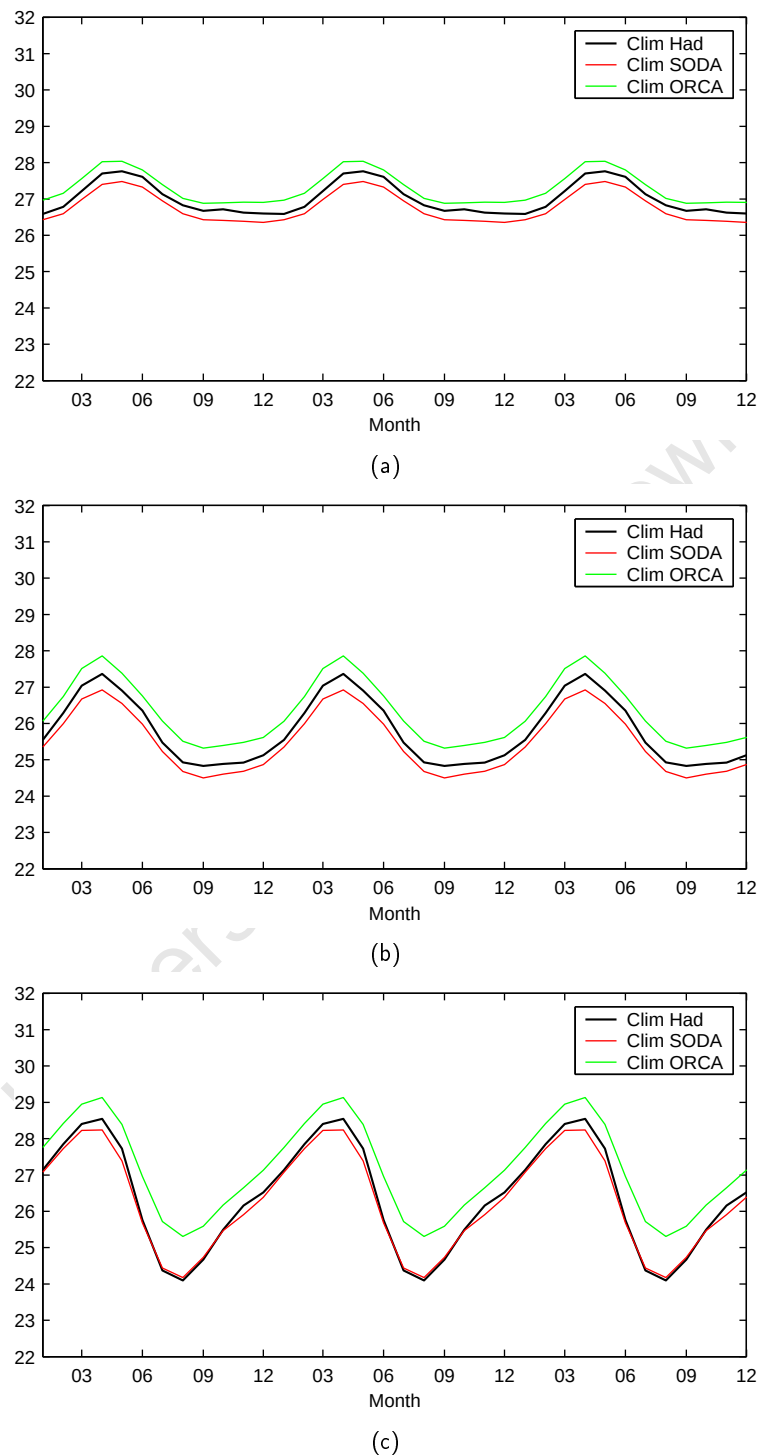


Figure 3.2: A comparison between climatological (a) Niño 3.4, (b) Niño 3, and (c) AtI3 SST, based on observed Hadley OI SST data (Rayner et al., 2003) and simulated SODA and ORCA SST data. SST values are shown in °C.

3.2.2 APE Estimation

Formalised by Lorenz (1955), the available potential energy of a fluid volume is defined as the difference in potential energy between its physical state and a rest state of minimum potential energy.

$$APE = \iiint g\rho z dV - \iiint g\rho_r z_* dV \quad (3.1)$$

where $\iiint g\rho z dV$ is the physical PE of the fluid volume and $\iiint g\rho_r z_* dV$ the minimum PE attainable within the fluid volume: a rest state in which all density surfaces are level. For the fluid parcel at position $(\vec{x}, t)$, $z_*(\vec{x}, t)$ is its vertical position and $\rho_r(\vec{x}, t)$ its density in the reference state of minimum PE. Where $\vec{x}$ is the three dimensional position vector $\vec{x} = x, y, z$.

In the oceanic context the fluid is assumed incompressible and so the effect of pressure on density during the levelling process is neglected ($\rho_r = \rho$) (Oort et al., 1989; Huang, 1998). Therefore:

$$\begin{aligned} APE &= \underbrace{\iiint g\rho z dV}_{PE} - \underbrace{\iiint g\rho z_* dV}_{BPE} \\ &= \iiint g\rho(z - z_*) dV \end{aligned} \quad (3.2)$$

where g is the acceleration due to gravity, $\iiint g\rho z dV$ is the physical PE of the fluid volume and $\iiint g\rho z_* dV$ the minimum PE attainable through the reversible adiabatic redistribution of mass, referred to as the Background Potential Energy (BPE). The vertical position of each water parcel in the rest state, $z_*(\vec{x}, t)$, is determined following Huang (1998). $V(\vec{x}, t)$ is the volume beneath the centre of the fluid parcel, $\rho(\vec{x}, t)$, once moved to its rest state position according to $V(\vec{x}, t) = \iiint H(\rho(\vec{x}', t) - \rho(\vec{x}, t)) dV'$, where H is a Heaviside step function, satisfying $H(y) = 0$ for $y < 0$, $H(y) = 0.5$ for $y = 0$ and $H(y) = 1$ for $y > 0$ (Winters et al., 1995; Huang, 1998). The corresponding vertical co-ordinate $z_*(\vec{x}, t) = z(V(\vec{x}, t))$ is then found based on the fact that the volume above the bottom of the chosen domain is a function of z (Huang, 1998).

Previous studies of equatorial energetics in the Pacific (Goddard and Philander, 2000; Fedorov et al., 2003; Fedorov, 2007) have used an approximation for APE (Equation 3.2) that assumes the background rest state is stably stratified and that density perturbations relative to the reference density profile are small in amplitude. APE can then be described by its leading order Taylor-expanded term (Reid et al., 1981; Oort et al., 1989):

$$\begin{aligned} APE_{approx} &= - \iiint \frac{g}{2 \frac{d\rho^*}{dz}} \tilde{\rho}^2 dV \\ &= \iiint \frac{1}{2} \frac{\tilde{\rho}^2}{N^2} dV \end{aligned} \quad (3.3)$$

where $N^2 = -\frac{d\rho^*/dz}{\rho^*}$ the depth-dependent stability factor, $\rho^*(z)$ is the vertical hydrostatically balanced density profile associated with the minimum PE rest state/BPE ($\rho^*(z) = \rho(z^*)$), and $\tilde{\rho}(\vec{x}, t)$ perturbations

from this reference state hence $\rho(\vec{x}, t) = \rho^*(z) + \tilde{\rho}(\vec{x}, t)$. A second assumption made in the previous literature (Goddard and Philander, 2000; Fedorov et al., 2003; Fedorov, 2007) when employing this definition for APE (Equation 3.3) is that the background rest state (the BPE of the fluid volume) is fixed in time: $\rho^*(z)$ as opposed to $\rho^*(z, t)$. The reason for making this assumption appears to be to facilitate the derivation of an evolution equation for APE as defined by Equation 3.3 (Goddard and Philander, 2000, Appendix). This assumption is only valid if changes in the mass and stratification of the chosen volume are negligible over the time scales upon which one is assessing APE changes. However, as outlined in the following paragraphs, the assumption that BPE changes are small does not appear to hold when evaluating APE changes within the equatorial Atlantic on seasonal to inter-annual time scales. As a result we chose not to use the approximation of APE given by Equation 3.3 as done in the previous literature (Goddard and Philander, 2000; Fedorov et al., 2003; Fedorov, 2007), but rather to evaluate APE according to Equation 3.2 for which one can take into account BPE changes as well as obtain a meaningful evolution equation as outline in Section 5.3.2.

The domain over which equatorial Atlantic APE is evaluated within this analysis has been chosen as 3°S-3°N 60°W-15°E 0-400m. The narrower meridional extent chosen in comparison to that of 5°S-5°N used in the Pacific by Goddard and Philander (2000), as well as within this analysis, is due to the narrower meridional extent of the Atl3 index in comparison to the Niño3 and Niño3.4 indices. The narrower meridional extent of the Atl3 index was chosen by Zebiak (1993) as SST variability associated with the zonal mode is more tightly focused on the equator (Figure 1.2). As discussed in Section 3.3.1, APE evaluated over the 3°S-3°N 60°W-15°E 0-400m domain correlates better with Atl3 SST at zero lag than APE evaluated over the 5°S-5°N 60°W-15°E 0-400m, 8°S-8°N 60°W-15°E 0-400m and 15°S-15°N 60°W-15°E 0-400m domains.

Climatological, equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE values calculated based on SODA and ORCA data are depicted in Figure 3.3. Labelled as method 1 are the APE values which have been calculated according to the APE approximation given by Equation 3.3 assuming a reference state that is constant in time. According to these estimates APE is at its annual minimum in February based on SODA data and in February/March based on ORCA data. One would therefore expect that the east-west slope of the equatorial thermocline reaches its climatological minimum in February. However when comparing east-west cross-sections of temperature along the equator during different months of the year, it is clear that the zonal slope of the thermocline is not at its climatological minimum in February but in April. This is evident in Figure 3.4 which depicts zonal changes in the climatological depth along the equator (3°S-3°N) of the 20°C isotherm (a typical proxy for the mid-thermocline in the tropics). The average zonal slope of the 20°C isotherm during the respective months is also given in Figure 3.4. The average slope of the 20°C isotherm reaches its minimum in both the SODA and ORCA climatologies in April. Moreover, APE estimates calculated according to method 1 peak in September (Figure 3.3) yet the zonal slope of thermocline is seen to be at a maximum in August (Figure 3.4).

On the other hand, climatological APE values for the equatorial Atlantic calculated according to Equation 3.2 with a reference state that is time dependent (method 2), are also depicted in Figure 3.3. Consistent with Figure 3.4, climatological APE values calculated according to method 2 reach a minimum in April when seasonally the east-west slope of the equatorial thermocline is at a minimum (Figure 3.4), and a maximum in August when the zonal slope of the equatorial thermocline is at a maximum. The contrast in the vertical temperature structure along the equator between April and August is shown in Figure 3.5.

When evaluating the nature of seasonal equatorial Atlantic APE changes, seasonal changes in the BPE of

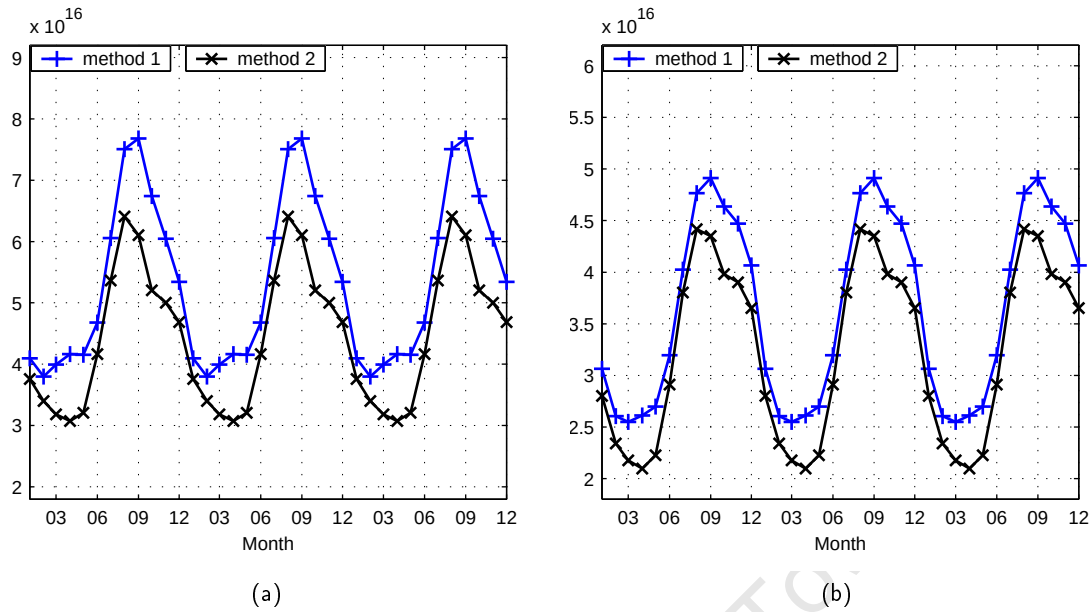


Figure 3.3: A comparison between equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) climatological APE values obtain using method 1, given by Equation 3.3, and method 2, given by Equation 3.2. (a) APE values have been calculated using SODA data. (b) APE values have been calculated using ORCA data.

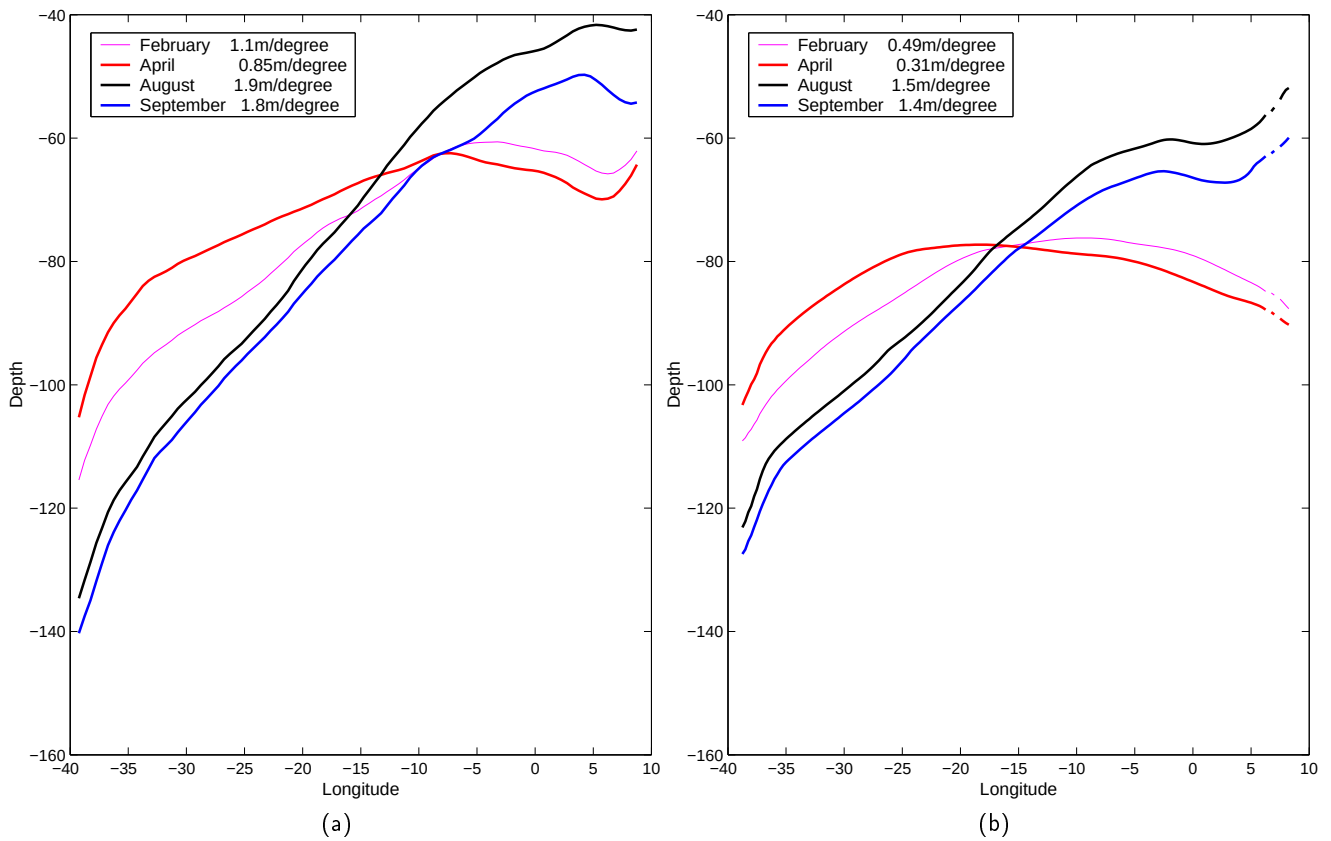


Figure 3.4: Zonal changes in the climatological depth of the 20°C isotherm along the equator (3°S - 3°N) for February, April, August, and September, based on (a) SODA data, and (b) ORCA data.

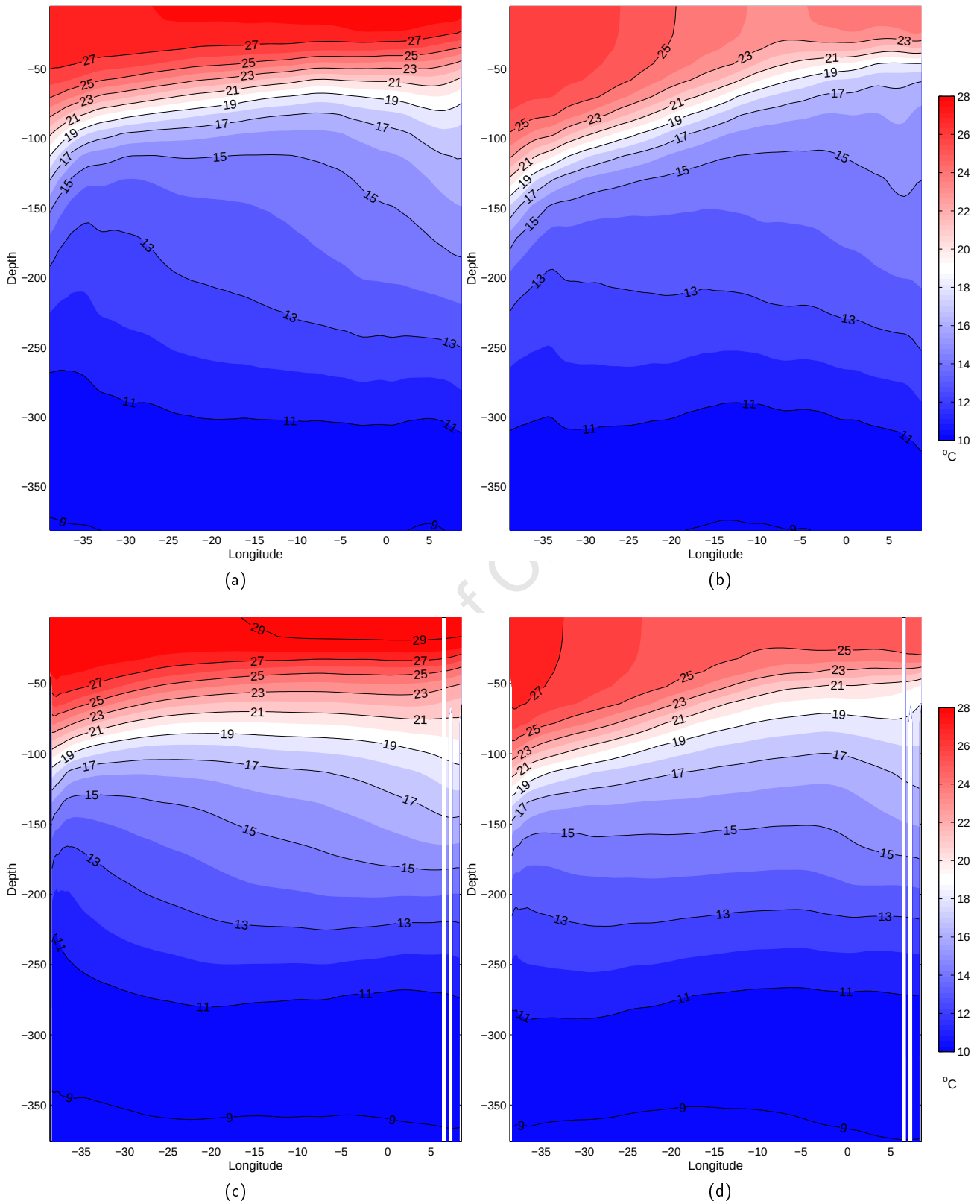


Figure 3.5: Cross-sections of temperature along the equator (3°S-3°N), (a) April and (b) August climatology based on SODA data, (c) April and (d) August climatology based on ORCA data.

equatorial Atlantic appear to affect the accuracy of method 1 which assumes a constant BPE. Figure 3.6 depicts the variability in the mean vertical density profile of the equatorial Atlantic (3°S - 3°N 60°W - 15°E), equatorial Pacific (5°N - 5°S 130°E - 85°W), tropical Atlantic (8°S - 8°N 60°W - 15°E), tropical Pacific (8°N - 8°S 130°E - 85°W), extended tropical Atlantic (15°S - 15°N 60°W - 15°E) and extended tropical Pacific (15°N - 15°S 130°E - 85°W) based on SODA data. The comparable figure created based on ORCA data, Figure A.1, can be found in Appendix A¹. The variability in the background rest state seen in these figures is neglected when calculating APE according to method 1. Within the equatorial zone and the 8°S - 8°N tropical zone, seasonal variability in the background rest state of the Atlantic considerably outweighs that observed in the Pacific. In the Pacific, the largest changes in the background rest state of the equatorial zone and the 8°S - 8°N tropical zone occur on inter-annual timescales ((Figures 3.6a, 3.6b, 3.6c and 3.6d). In both basins, variability in the background rest state on both seasonal and inter-annual timescales is seen to be small for the 15°S - 15°N tropical zone (Figures 3.6e and 3.6f).

Why is seasonal variability in the background rest state of the Atlantic larger than in the Pacific (Figure 3.6)? As mentioned in Section 2, seasonal changes in the global upper ocean heat content of the equatorial (3°N - 3°S) and tropical Atlantic (8°N - 8°S) are seen to occur due to seasonal changes in warm water formation and escape (Merle, 1980a; Philander and Pacanowski, 1986b; Lee and Csanady, 1999; Bunge and Clarke, 2009). Large seasonal changes occur in the heat content of the upper tropical Atlantic with the mean depth of the thermocline reaching a maximum in October-November and minimum in June-July (Merle, 1980a). The tropical Atlantic imports heat from July to October, but for the rest of the year it exports heat resulting in a net annual export of heat (Merle, 1980a). These observed seasonal changes in the heat content of the equatorial and tropical Atlantic are depicted in Figure 3.7 which shows the observed seasonal cycle in the average temperature of the equatorial and tropical Atlantic regions based on WOA05 (Locarnini et al., 2006) temperature data.

Figure 3.8 illustrates the seasonal cycle in the average density of the upper equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) and the upper tropical Atlantic (8°S - 8°N 60°W - 15°E 0-400m) calculated based on simulated temperature and salinity fields from the SODA and ORCA ocean models as well as observed temperature and salinity fields from the WOA05. Seasonal changes in the mass of both the equatorial and tropical regions are consistent with the observed seasonal variability in the heat content of these regions described in Section 2 and depicted in Figure 3.7. Seasonal changes in density values (Figure 3.8) mirror temperature changes (Figure 3.7) peaking in July when the average temperature of the tropical and equatorial domains is at a minimum and reaching a minimum in October/November when the average temperature of the tropical and equatorial domains is at a maximum. From this comparison two inferences are made: firstly, that seasonal variability in the density field is determined primarily by seasonal variability in the temperature field, and secondly that the SODA and ORCA simulations have captured the seasonal warm water formation and escape process. However the ORCA simulation, which unlike the SODA simulation was unassisted by the assimilation of WOA data, appears to have underestimated the magnitude of seasonal density changes within the tropical and equatorial regions.

In Section 5.3.3 it is shown that seasonal changes in the BPE of the equatorial and tropical domains are controlled by seasonal changes in advection of the density field and surface buoyancy forcing with seasonal changes in mixing playing a relatively small role. Furthermore these seasonal changes in advection of

¹For brevity, when the same figure has been created based on SODA and ORCA data, and they both reveal consistent results, the figure based on ORCA data has been relegated to Appendix A.

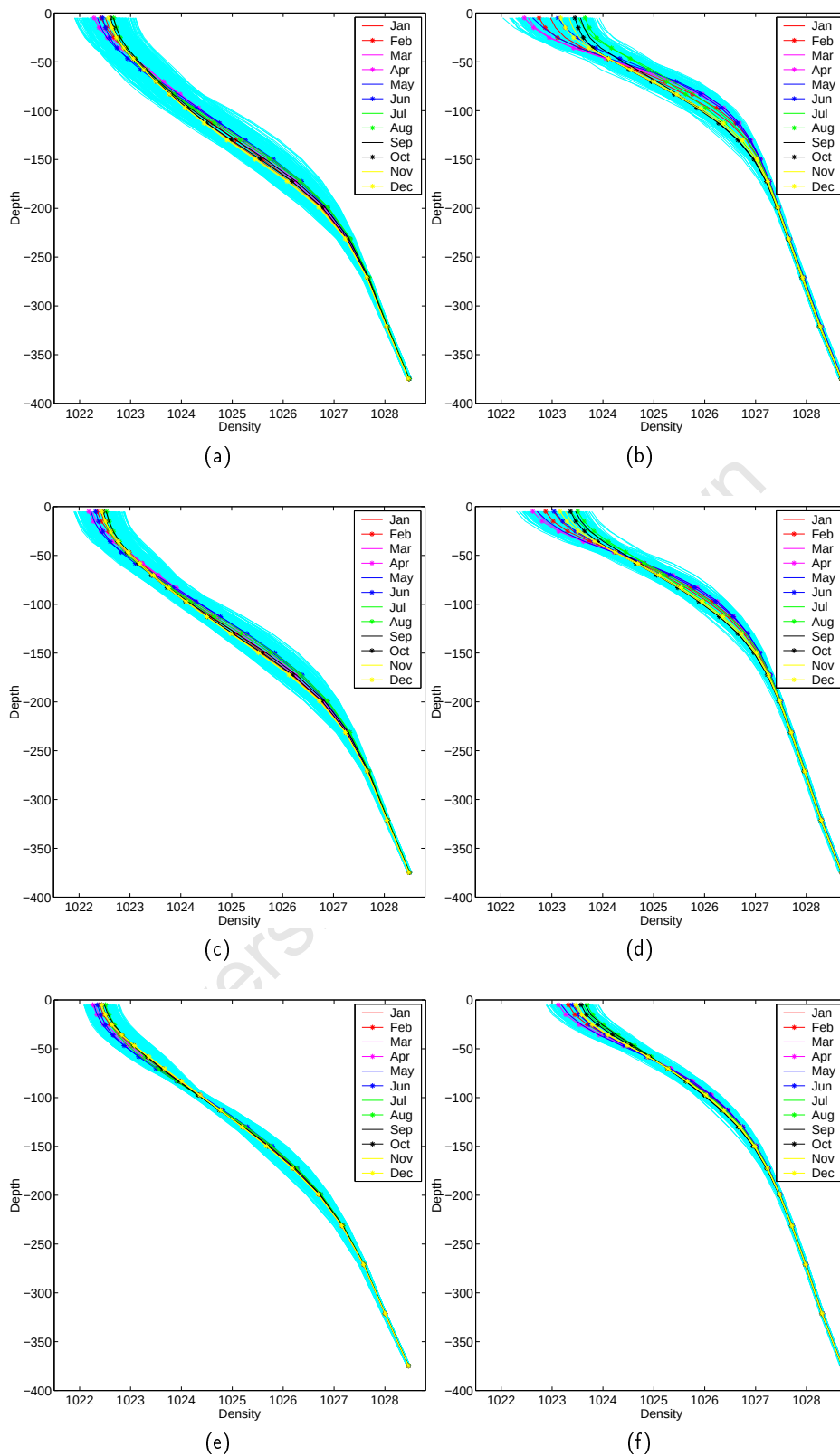


Figure 3.6: Variability in the vertical density profile estimated by horizontally averaging the monthly mean density field within the chosen domain: (a) the equatorial Pacific (5°N-5°S 130°E-85°W), (b) the equatorial Atlantic (3°S-3°N 60°W-15°E), (c) the tropical Pacific (8°N-8°S 130°E-85°W), (d) the tropical Atlantic (8°S-8°N 60°W-15°E), (e) the extended tropical Pacific (15°N-15°S 130°E-85°W), and (f) the extended tropical Atlantic (15°S-15°N 60°W-15°E). Seasonal variability in the vertical density profile is overlayed on the total spread of vertical density profiles between 1958-2004 shown in cyan. Depth (m) is represented on the y-axis and density (kgm^{-3}) on the x-axis. Monthly density fields are derived from the monthly SODA temperature and salinity fields.

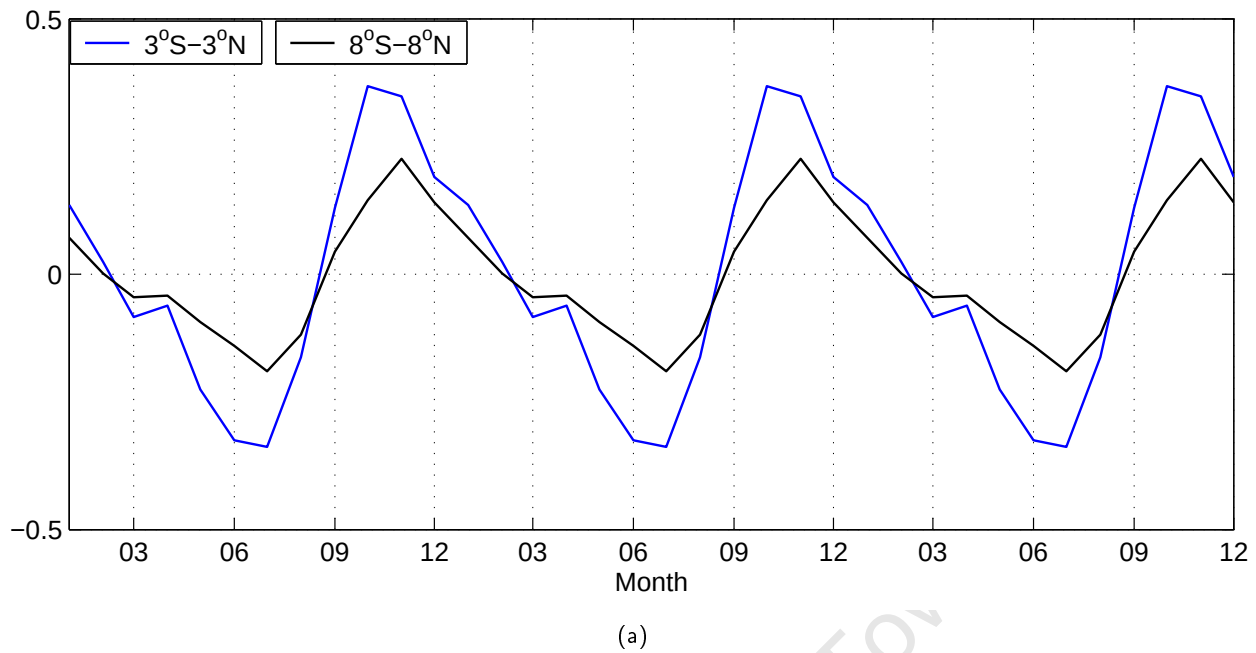


Figure 3.7: Seasonal changes in the average temperature of the upper equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) and the upper tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) based on the WOA05 (Locarnini et al., 2006) climatological temperature field.

the density field and surface buoyancy forcing appear to be related with seasonal changes in warm water formation (regarded in Section 5.3.3 in terms of surface buoyancy forcing rather than entrainment) and escape (largely controlled by off-equatorial wind stress curl) as described in Section 2.

Heat content changes in equatorial Atlantic are therefore seen to be the cause of these significant seasonal changes in the background rest state which have to be considered in order to accurately represent seasonal APE changes. Figure 3.9 shows the relationship between seasonal changes in the APE of the equatorial Atlantic computed as delineated by Equation 3.3, seasonal changes in the APE and the BPE of the equatorial Atlantic computed as delineated by Equation 3.2, as well as, seasonal changes in the mass of the equatorial Atlantic volume. The time series have been normalised to facilitate a better comparison. Assuming global conservation of mass over this oceanic volume biases the APE values obtained using method 1 (Figures 3.3 and 3.9). As expected, seasonal changes in the BPE of the equatorial Atlantic are in-phase with changes in the mass of the equatorial volume (Figures 3.8 and 3.9) and the observed seasonal changes in the upper ocean heat content of the equatorial Atlantic (Merle, 1980a; Bunge and Clarke, 2009). Climatological BPE values appear to accurately represent changes in the heat content of the upper equatorial Atlantic.

Temporal changes in BPE have to be taken into account in order to cleanly separate changes in the PE of an equatorial region due to the sloping of isopycnals (APE) from changes in the PE due seasonal or inter-annual changes in the amount of warm water within the domain under consideration. When evaluating inter-annual variability in tropical Pacific (15°N-15°S 130°E-85°W 0-400m) APE, the assumption of global conservation of mass over this oceanic volume may be safer with diabatic changes remaining relatively small (Figure 3.6e) in comparison to adiabatic changes. However, this study focuses on APE values for the equatorial Atlantic region (3°S-3°N 60°W-15°E 0-400m) where the transfer of warm surface waters from the equatorial region to off equatorial regions is significant as well as the tropical Atlantic region (8°S-8°N 60°W-15°E 0-400m) where the seasonal warm water formation and escape process results in significant seasonal changes in the region's

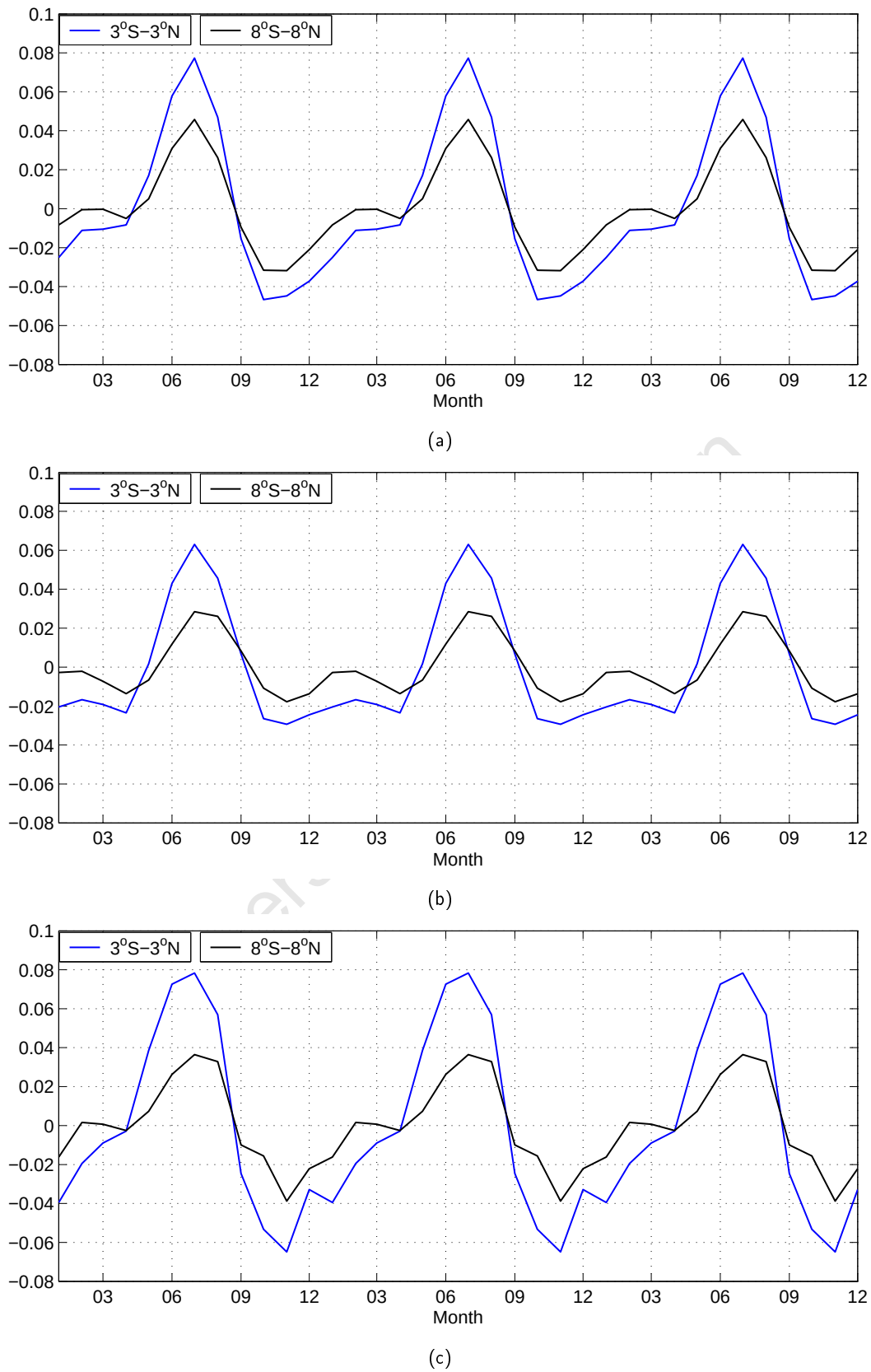


Figure 3.8: Seasonal changes in the average density of the upper equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) and the upper tropical Atlantic (8°S-8°N 60°W-15°E 0-400m). (a) Climatological density values have been calculated based on simulated temperature and salinity fields from the SODA model. (b) Climatological density values have been calculated based on simulated temperature and salinity fields from the ORCA model. (c) Climatological density values have been calculated based on WOA05 climatological temperature and salinity fields.

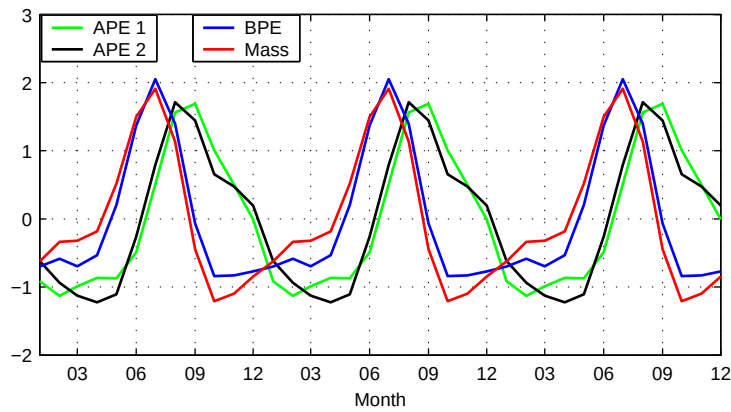


Figure 3.9: Seasonal changes in equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE obtained using Equation 3.3, APE and BPE obtained using Equation 3.2, as well as seasonal changes in the Mass of this domain. Mass, BPE and APE values have been normalised. Values shown here have been calculated using SODA data, the same figure only based on ORCA data, Figure A.2, can be found in Appendix A.

background rest state. The method outlined by Equation 3.2 has therefore been chosen to evaluate APE for this study. To facilitate the comparison between equatorial Atlantic and equatorial Pacific energetics, total, climatological and anomalous APE values have been calculated based on density fields derived from the monthly temperature and salinity fields of the SODA and ORCA simulation.

3.3 Equatorial Atlantic vs Equatorial Pacific Analysis

3.3.1 Equatorial Atlantic vs Equatorial Pacific APE - SST Relationship

Figures 3.10 and 3.11, depict the relationship between total values in equatorial APE and central-eastern basin SST indices for the Pacific and Atlantic respectively. A correlation coefficient of $r=-0.83$ ($r^*=-0.25$) is seen between the Niño3.4 index and equatorial APE based on SODA data (Figure 3.10) and one of $r=-0.88$ ($r^*=-0.23$) based on ORCA data (Figure A.3) while a correlation coefficient of $r=-0.72$ ($r^*=-0.3$) is seen between the Niño3 index and equatorial APE based on SODA data (Figure 3.10) and one of $r=-0.74$ ($r^*=-0.35$) based on ORCA data (Figure A.3). The robust relationship between central-eastern basin SST and equatorial APE observed in the Pacific (Goddard and Philander, 2000) is also seen to be present in the Atlantic basin. A correlation coefficient of $r=-0.85$ ($r^*=-0.51$) is seen between the Atl3 index and equatorial APE based on SODA data (Figure 3.11) and one of $r=-0.89$ ($r^*=-0.53$) based on ORCA data (Figure A.4). This result is not surprising considering that both equatorial basins are characterised by a mean state comprising of easterly wind stress and a zonally sloping thermocline which shoals in the east.

Equatorial SST is determined by the competing effects of surface heating, advection and mixing processes. In the western tropical Pacific and Atlantic basins, a deep thermocline means that temperature gradients near the surface are weak. Subsequently fluctuations in the magnitude of horizontal and vertical temperature advection associated with surface currents are weak in comparison to fluctuation in diabatic forcing by vertical mixing and the surface heat flux (Wang and Enfield, 2001; Yu et al., 2006). However in the central-eastern Pacific (Hayes et al., 1991; Chang and Philander, 1994; Wang and McPhaden, 1999) and Atlantic (Weingartner and Weisberg, 1991; Carton and Zhou, 1997; Peter et al., 2006), where the thermocline is

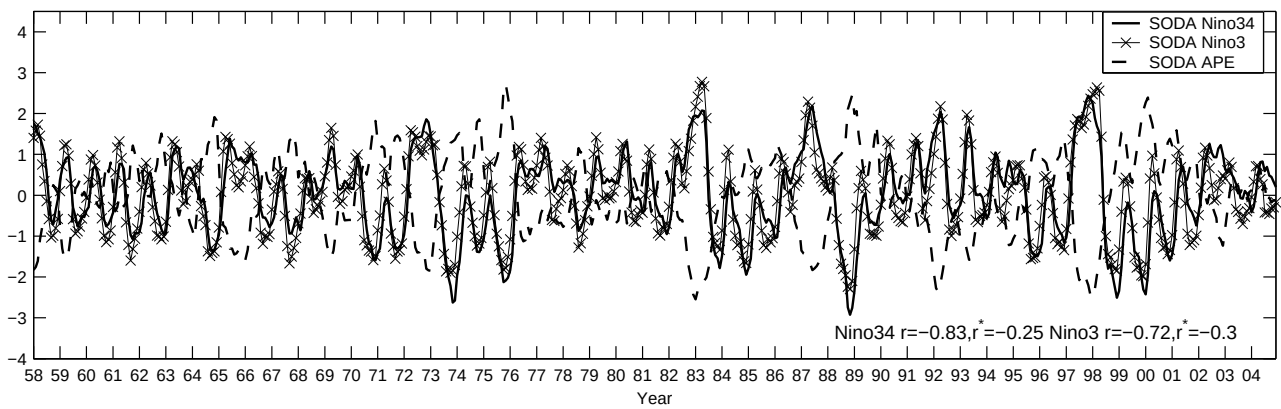


Figure 3.10: A comparison between total Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE variability. SST and APE values were calculated using SODA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.3, can be found in Appendix A.

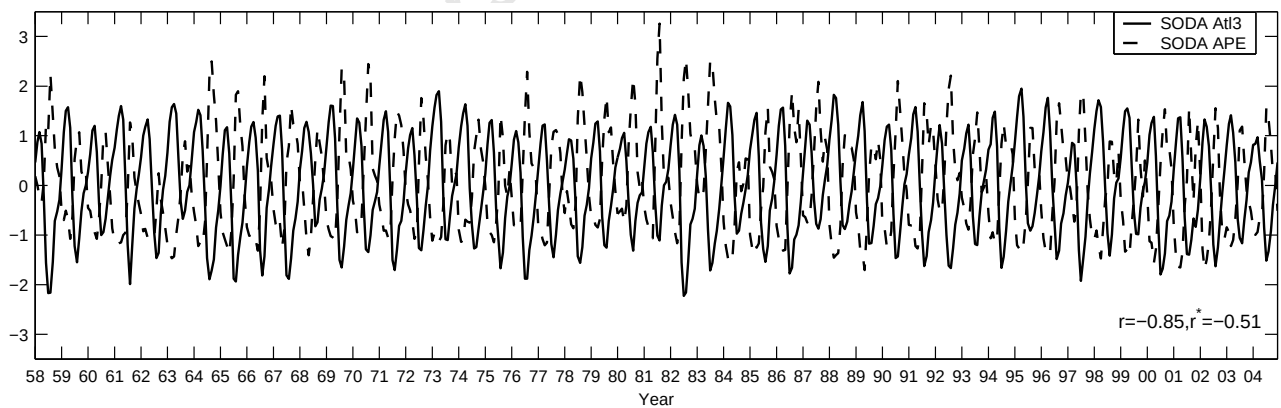


Figure 3.11: A comparison between total Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE variability. SST and APE values were calculated using SODA data. The correlation between the Atl3 SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.4, can be found in Appendix A.

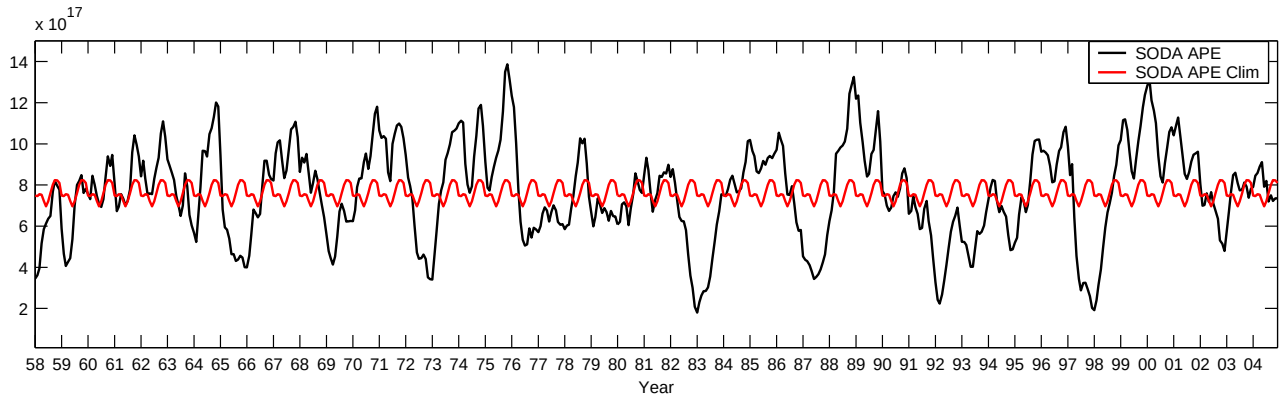


Figure 3.12: A comparison between total and climatological equatorial Pacific (5°N - 5°S 130°E - 85°W 0-400m) APE values. APE values were calculated following Equation 3.2 using SODA data and have been detrended. The corresponding figure based on ORCA data, Figure A.5, can be found in Appendix A.

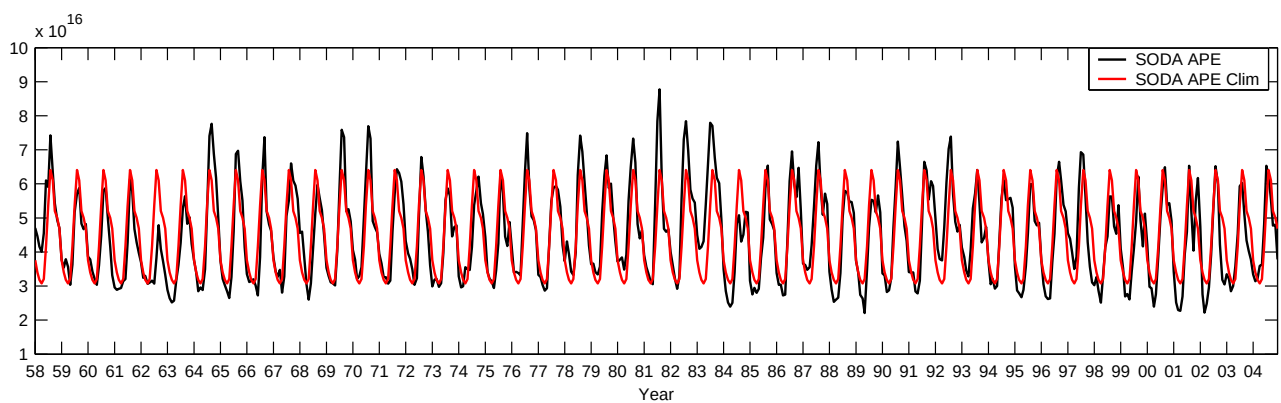


Figure 3.13: A comparison between total and climatological equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE values. APE values were calculated following Equation 3.2 using SODA data and have been detrended. The corresponding figure based on ORCA data, Figure A.6, can be found in Appendix A.

shallow, vertical velocities near the surface act on a strong vertical temperature gradient. The vertical advection of colder sub-thermocline water into the upper layer, where it can be mixed into the surface mixed layer, is therefore amongst the dominant processes regulating SST.

Changes in the magnitude of this vertical advection may result from either locally forced changes in vertical Ekman velocities or remotely forced thermocline depth changes. The presence of a common SST-APE relationship in both basins whereby SST fluctuations in the eastern-central equatorial basin are highly anti-correlated with equatorial APE, suggests that the primary dynamic ocean process responsible for the large inter-annual Niño 3.4 and Niño 3 SST fluctuations plays an equally important role in determining Atl3 SST variability. The primary dynamic ocean process responsible for the large inter-annual Niño 3.4 and Niño 3 SST fluctuations being anomalous vertical advection in the central-eastern basin resulting from thermocline depth changes associated with the large scale redistribution of warm surface water (McPhaden et al., 1998). SST changes forced by this dynamic process are anti-correlated with APE. The shoaling (deepening) of the thermocline in the east, as the slope of the thermocline increases (decreases), results in negative (positive) SST anomalies and positive (negative) APE anomalies. Central-eastern basin SST changes in the equatorial Pacific and Atlantic that correlate with APE changes, are merely surface expressions of the subsurface changes in the temperature structure associated with these APE changes (Goddard and Philander, 2000).

In the equatorial oceans, APE changes are driven primarily by the horizontal redistribution of warm surface waters in response to changes in wind driven circulation (Goddard and Philander, 2000). The biggest distinction between these two basins exists in the period of dominant APE fluctuations. Figures 3.12 and 3.13 show the magnitude of total APE changes versus climatological APE changes in the equatorial Pacific and Atlantic respectively. Unlike in the Pacific, the largest changes in APE occur annually in the Atlantic. While the east-west tilt of the thermocline hardly varies seasonally in the Pacific (McPhaden et al., 1998), large seasonal variations in the zonal slope of the thermocline are observed seasonally in the Atlantic (Merle, 1980a; Vauclair and du Penhoat, 2001). The analysis conducted within this chapter and Chapter 5 suggests that this distinction between the two basins in terms of the time scale of dominant APE fluctuations is due to the fact that while the Bjerknes feedback acts on inter-annual timescales in the Pacific, it acts on seasonal timescales in the Atlantic. This difference is thought to be due to the fact that the width of the Atlantic is less than a third of the Pacific, reducing the time scale of oceanic adjustment and allowing the oceanic response in the equatorial Atlantic to be closer to equilibrium with the seasonal forcing (Merle and Arnault, 1985; Philander and Pacanowski, 1981).

The extension of the meridional domain over which APE is evaluated to include APE changes due to the meridional redistribution of mass within the tropical domain does not have a significant effect on the signature of APE variability (Figure 3.14). The seasonal signal is still dominant in the Atlantic. Fluctuations in the APE of the equatorial domain extending from 3°S-3°N 60°W-15°E correlate best with corresponding SST changes in the Atl3 region at zero lag, while APE values for the 5°S-5°N 60°W-15°E domain, the 8°S-8°N 60°W-15°E domain and the 15°S-15°N 60°W-15°E domain correlate best with SST changes in the Atl3 region at one months lag. The strength of this one month lagged correlation for the the 5°S-5°N 60°W-15°E, 8°S-8°N 60°W-15°E and 15°S-15°N 60°W-15°E domains decreases as the meridional extent increases. This result is ascribed to the fact that the larger domains take longer to adjust to seasonal changes in the wind forcing due to the slower phase speed of off equatorial Rossby waves in comparison to equatorial Kelvin waves.

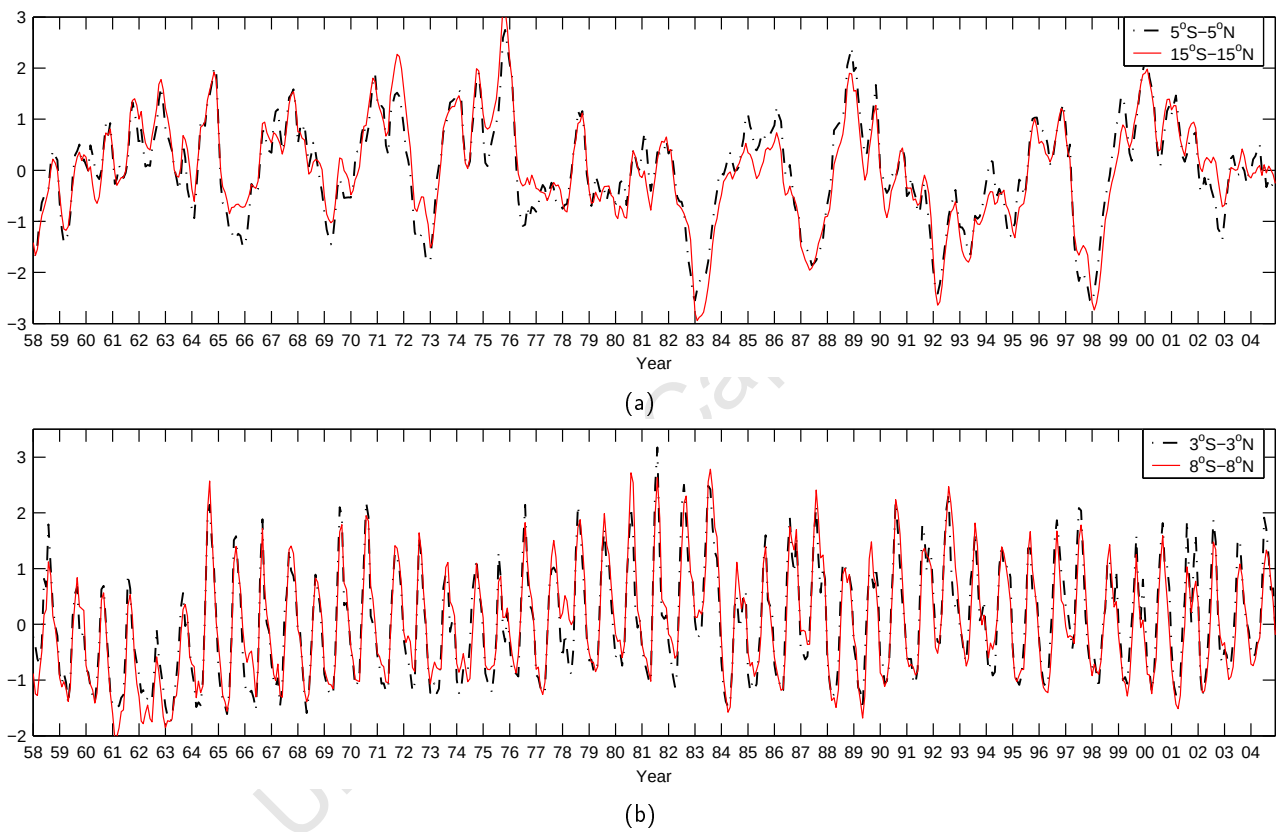


Figure 3.14: (a) Equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE versus tropical Pacific (15°N-15°S 130°E-85°W 0-400m) APE. (b) Equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE versus tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE. APE values were calculated using SODA data following Equation 3.2. The corresponding figure based on ORCA data, Figure A.7, can be found in Appendix A.

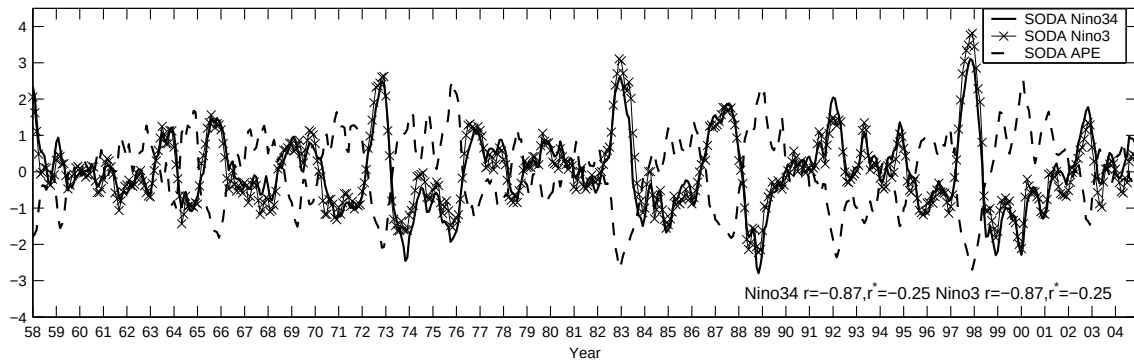


Figure 3.15: A comparison between inter-annual Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE anomalies. SST and APE anomalies were calculated using SODA data. The correlation between anomalies in each SST index and APE (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.8, can be found in Appendix A.

3.3.2 Equatorial Pacific APE - SST relationship on Seasonal and Inter-annual time scales

As mentioned in the introduction to this chapter, inter-annual SST fluctuations in the central-eastern equatorial Pacific are highly anti-correlated with the amount of gravitational APE in the equatorial Pacific (Goddard and Philander, 2000). The strength of this relationship, which is a result of the Bjerknes feedback mechanism operating at inter-annual time scales in the Pacific, is depicted in Figure 3.15. A correlation coefficient of $r=-0.87$ ($r^*=-0.25$) is seen between Niño3.4 anomalies and equatorial APE anomalies based on SODA data (Figure 3.15) and one of $r=-0.91$ ($r^*=-0.25$) based on ORCA data (Figure A.8) while a correlation coefficient of $r=-0.87$ ($r^*=-0.25$) is seen between the Niño3 anomalies and equatorial APE anomalies based on SODA data (Figure 3.15) and one of $r=-0.89$ ($r^*=-0.25$) based on ORCA data (Figure A.8).

Figure 3.16 depicts the relationship between seasonal Niño 3.4 and Niño 3 SST changes and seasonal equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE changes. The correlation between seasonal SST changes in the central-eastern Pacific and APE, is less than that observed between inter-annual SST and APE anomalies. A correlation coefficient of $r=-0.76$ ($r^*=-0.71$) is seen between the Niño3.4 index and equatorial APE based on SODA data (Figure 3.16) and one of $r=-0.85$ ($r^*=-0.71$) based on ORCA data (Figure A.9) while a correlation coefficient of $r=-0.64$ ($r^*=-0.71$) is seen between the Niño3 index and equatorial APE based on SODA data (Figure 3.16) and one of $r=-0.77$ ($r^*=-0.71$) based on ORCA data (Figure A.9).

Seasonally forced changes in the density field of the equatorial Pacific result in comparatively small APE changes in comparison to those which occur inter-annually (Figure 3.12, Fedorov, 2007). This result is consistent with observationally based findings that, although seasonal variations in the thermocline depth have a large amplitude north of the equator, the thermocline depth along the equator does not vary much seasonally in the Pacific (McPhaden et al., 1998). The largest seasonal temperature variations in the eastern equatorial Pacific are confined to the surface mixed layer (McPhaden et al., 1998). Seasonal changes in APE appear to be too weak to dictate seasonal SST variability. Instead, as a wealth of studies in the Pacific

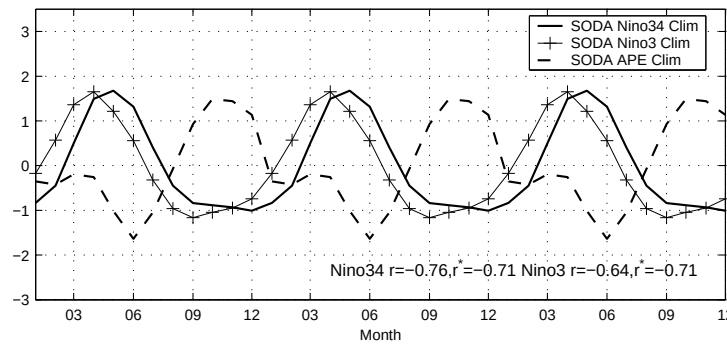


Figure 3.16: A comparison between climatological Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE. SST and APE values were calculated using SODA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.9, can be found in Appendix A.

have shown, seasonal SST changes in the Pacific's cold tongue region are highly correlated with local winds (Hayes et al., 1991; Chang, 1993, 1994; Swenson and Hansen, 1999; Wang and McPhaden, 1999). While surface heating, vertical mixing, horizontal advection and vertical advection all play a significant role, seasonal fluctuations in surface heating and subsurface cooling are seen to be the dominant processes driving seasonal SST changes in the central-eastern equatorial Pacific (Chang, 1994; McPhaden et al., 1998; Wang and McPhaden, 1999). The term subsurface cooling encompasses the combined effects of vertical advection and vertical mixing and seasonal changes in the intensity of both these processes are in-phase with the local winds (Wang and McPhaden, 1999).

SSTs in the Pacific's cold tongue region reach their minimum during September/October as strong local winds generate turbulence increasing vertical mixing and decreasing local stratification. During this period, strong local winds also enhance local Ekman upwelling velocities which bring cold water closer to the surface where it can be mixed into the surface mixed layer (Wang and McPhaden, 1999). The opposite is true during boreal spring when SSTs reach their maximum. Local stratification is enhanced as local winds are at their weakest and upwelling velocities, vertical mixing and the surface mixed layer depth are all at a minimum (Wang and McPhaden, 1999).

Seasonal variations in the intensity of the local winds and associated Ekman pumping are thought to be controlled by equatorial antisymmetric and equatorial symmetric SST-modes which govern the seasonal cycle in the eastern Pacific, where the thermocline is shallow (Chang and Philander, 1994). These unstable coupled modes are forced by the local Ekman feedback mechanism and do not involve remotely forced changes in thermocline depth. Seasonal SST changes are primarily determined by surface Ekman currents, their divergence, and changes in the thermocline depth which are in balance with local wind stress forcing (Chang and Philander, 1994).

As evidenced by Figure 1.1c, seasonal cooling in the eastern Pacific is clearly asymmetric about the equator as a zero frequency, zero zonal wave number asymmetric SST-mode with a rapid growth rate is accredited with the sudden seasonal growth in asymmetric condition about the equator (Chang and Philander, 1994). An unstable cross-equatorial Ekman feedback mechanism leads to the growth of the equatorial cold tongue-ITCZ complex from April to September. The associated meridional wind stress fluctuations along the equator are

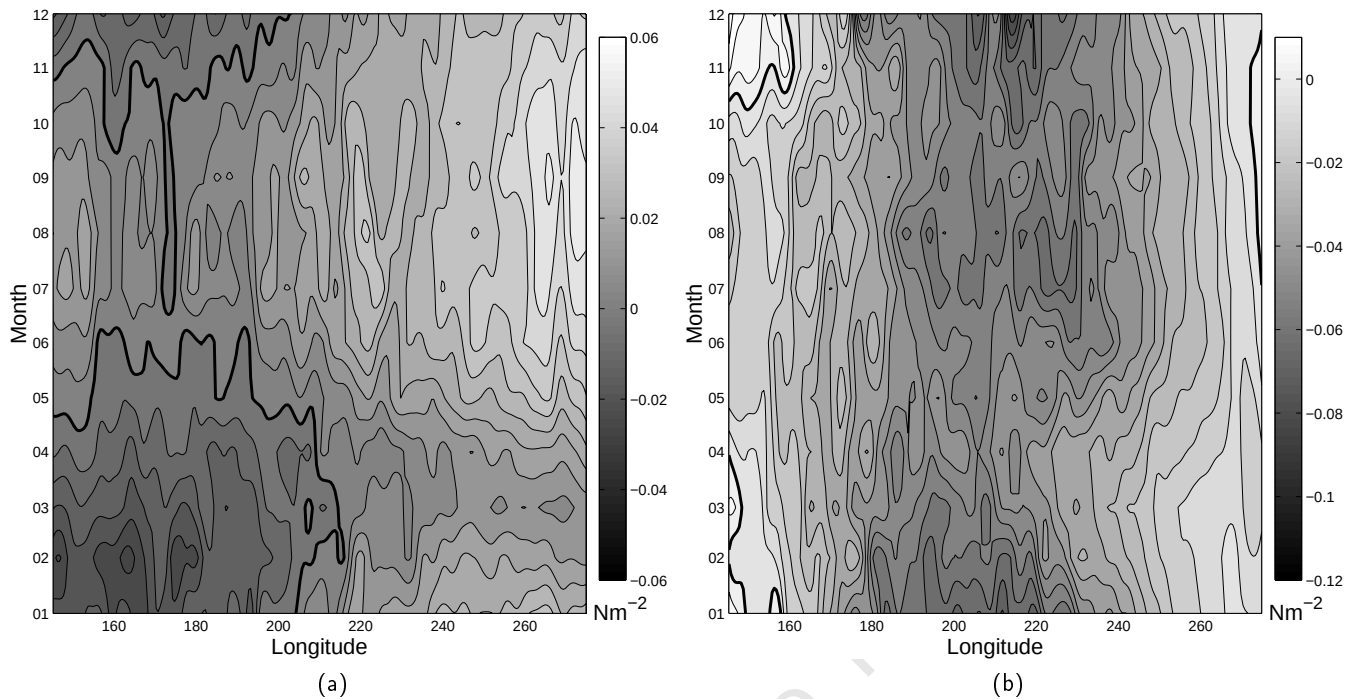


Figure 3.17: This figure depicts the seasonal cycle of (a) meridional wind stress and (b) zonal wind stress along the equator (3°S-3°N) in the Pacific. COADS data (DaSilva et al., 1994) and a contour interval of 0.005 Nm^{-2} has been used. The thick black line represents the 0 Nm^{-2} contour line.

shown in Figure 3.17a, which depicts the seasonal cycle of meridional wind stress along the equator. In the central-eastern Pacific, southerly meridional wind stress persists throughout the year as the ITCZ remains north of the equator. However, in the western basin it changes sign as the ITCZ switches hemispheres (Figure 3.17a). Meridional wind stress values in the central-eastern basin are at a minimum between March and April when equatorial SSTs are at their maximum and the ITCZ is nearest the equator (Figures 1.1c and 3.17a). As the cold tongue develops and the ITCZ migrates northward, southerly meridional wind stress on the equator increases in magnitude, peaking when asymmetrical conditions reach their maximum in September (Figures 1.1c and 3.17a). Once established, the equatorial cold tongue-ITCZ complex is robust and persists until December even as the region of maximum insolation and continental convection starts to move southwards (Figures 1.1c and 1.3). Only towards the end of the southern hemisphere summer (February-April), does insolation manage to warm south equatorial SST enough to weaken the cold tongue-ITCZ complex allowing the ITCZ to migrate southwards back towards the equator (Figure 1.1c, Chang and Philander, 1994).

Figure 3.17b shows the seasonal cycle of zonal wind stress along the equator. The strongest equatorial easterlies reside in the central basin. In April when the ITCZ is nearest the equator and SSTs are at their maximum, equatorial easterlies are at their seasonal minimum (Figures 1.1c and 3.17b). As the ITCZ migrates northwards and asymmetric condition grow, the equatorial easterlies intensify. An equatorially symmetric SST-mode is believed to be responsible for the westward propagation of these seasonally enhanced equatorial trades (Figure 3.17b), and the associated seasonal cooling (Figures 1.1b and 1.1c), along the equator. This westward propagating equatorially SST-mode results in the lagged, more equatorially symmetric, seasonal cooling in the Niño 3.4 region seen in Figures 1.3 and 1.1b (Chang and Philander,

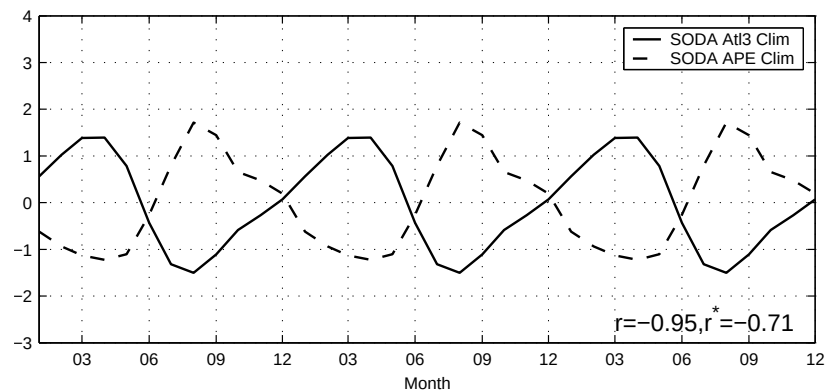


Figure 3.18: A comparison between climatological Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE. SST and APE values were calculated using SODA data. The correlation between the Atl3 SST index and equatorial Atlantic APE values (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.10, can be found in Appendix A.

1994). Thus, the seasonally excited Ekman feedback mechanism and associated asymmetric and symmetric SST-modes account for the signature of seasonal Niño 3 and Niño 3.4 SST fluctuations.

The dominant mechanisms forcing seasonal SST variability are therefore somewhat different from that associated with inter-annual variability in the equatorial Pacific. As a result, the spatial pattern associated with seasonal SST variability in the Pacific (Figure 1.2c) differs from that associated with inter-annual variability (Figure 1.2e) in a manner that is consistent with the different physical processes driving the respective variability - seasonal variability displaying a pronounced asymmetry while inter-annual variability is equatorially focused. As illustrated by Figure 1.2d and Figure 1.2f this distinction does not exist in the Atlantic.

3.3.3 Equatorial Atlantic APE - SST relationship on Seasonal and Inter-annual time scales

Unlike the Pacific, seasonal changes in equatorial Atlantic APE are much larger than inter-annual fluctuations (Figure 3.13). This result is consistent with the large variations in the zonal slope of the thermocline observed seasonally in the Atlantic (Figure 3.4, Merle, 1980a; Vauclair and du Penhoat, 2001). Furthermore, a strong relationship exists between seasonal APE and Atl3 SST changes, as depicted in Figure 3.18 which illustrates the relationship between seasonal Atl3 SST changes and seasonal equatorial Atlantic APE changes. A correlation coefficient of $r = -0.95$ ($r^* = -0.71$) is seen between the Atl3 index and equatorial APE based on SODA data (Figure 3.18) and one of $r = -0.96$ ($r^* = -0.71$) based on ORCA data (Figure A.10).

Seasonal SST fluctuations in the central-eastern equatorial Atlantic are regulated primarily by subsurface cooling and warming by atmospheric heat fluxes and oceanic eddies (Weingartner and Weisberg, 1991; Carton and Zhou, 1997; Foltz et al., 2003; Peter et al., 2006). The correlation between large seasonal fluctuations in APE and SST shown above implies that remotely forced thermocline variability plays a key role in influencing seasonal SST variability via subsurface cooling. This influence is found to be the case in

Section 5.2 of Chapter 5, where the relative contribution of each surface mixed layer process to seasonal SST variability in the central-eastern equatorial Atlantic is assessed.

As shown in Figure 3.18, seasonal cooling of SSTs in the Atl3 region is significant and occurs rapidly between April-August as APE sharply increases and the cold tongue develops. Warming on the other hand is gradual and occurs over the remainder of the year. Based on this and the fact that the largest APE fluctuations occur seasonally as opposed to inter-annually, it certainly appears as though the noticeable difference in the time-scale of the largest SST fluctuations between the Atlantic and Pacific basins (Figure 1.4) is due to the fact that the mechanism responsible for the large inter-annual SST changes in the equatorial Pacific, namely the Bjerknes mechanism, is in fact seasonally excited in the Atlantic.

The nature of seasonal zonal and meridional wind stress fluctuations in the Atlantic can be compared against those in the Pacific by contrasting Figures 2.3 and 3.17. The meridional wind stress changes associated with the growth of the equatorial cold tongue-ITCZ complex from April are similar in both basins, except that the increase in southerly flow is far larger in the western Atlantic than in the western Pacific (Figures 2.3 and 3.17). In both basins, an intensification of the equatorial easterlies, which propagates westward, is seen as the ITCZ migrates northwards and asymmetric conditions grow. However, while this intensification is largest in the central-western basin in the Atlantic, it is largest in the central-eastern basin in the Pacific.

Currently, the primary idealised coupled mode and associated feedback mechanism deemed responsible for the seasonal evolution of coupled conditions and the development of the cold tongue in the Atlantic is thought to be equivalent to that responsible for the seasonal development of the cold tongue in the Pacific. SST modes and their associated Ekman feedback mechanism are believed to be behind the growth of the cold tongue in the Atlantic (Mitchell and Wallace, 1992; Chang and Philander, 1994). However, as pointed out by Figure 1.3, the signature of seasonal SST changes for the Atl3 index is somewhat distinct from that of Niño3 and Niño3.4 indices whose seasonal behaviour has likewise been ascribed to seasonally excited SST-modes. The seasonal decline of eastern basin SSTs commences in April in each basin as asymmetrical SST-modes are excited and asymmetrical conditions grow. The evidence offered here suggests however that the greater and steeper seasonal decrease observed in the Atlantic is due to the fact that the seasonally excited growth of asymmetric conditions is accompanied by a seasonally excited Bjerknes feedback as the thermocline shoals seasonally between April and August, a process absent in the Pacific. This statement is supported further by comparing Figures 1.1c and 1.1f. In the eastern Pacific, seasonal cooling is seen to be clearly asymmetric about the equator while in the eastern Atlantic, equatorial cooling is more symmetric presumably due to the contribution of equatorially symmetric remotely forced thermocline depth changes associated with a seasonally excited Bjerknes feedback.

The fact that remotely forced thermocline depth changes play an active role in the seasonal cooling of Atl3 SST is nevertheless not enough to prove that the Bjerknes feedback is seasonally active, as this forms only one of the three main constituents of the Bjerknes feedback mechanism. Supporting the concept that the Bjerknes feedback mechanism is excited seasonally between April and August, and plays an active role in the development of the cold tongue in the Atlantic, requires evidence that the other two essential components of the feedback mechanism participate in the development of the cold tongue.

Firstly, evidence that seasonal cooling of central-eastern basin SST enhances easterly zonal winds in the western Atlantic is offered in the following section (Section 3.3.4) where the relationship between zonal wind stress and SST is explored. Secondly, that these wind changes in the west shoal the thermocline in the east,

increasing the APE of the equatorial Atlantic as the cold tongue develops. Affirmation of this process is offered in the following section (Section 3.3.4), as the relationship between zonal wind stress and APE is explored, as well as in Section 5.3 wherein the primary processes determining the seasonal evolution of APE in the equatorial Atlantic are investigated.

Another distinction observed between seasonal SST variability in the Niño 3 and Niño3.4 regions versus the Atlantic's Atl3 region is that cool conditions stabilise and persist for longer in the Pacific than the Atlantic (Figure 1.3). As mentioned in Section 3.3.2, once established the equatorial cold tongue-ITCZ complex in the Pacific sustains itself until December even as the region of maximum insolation and continental convection moves southwards and only towards the end of the southern hemisphere summer does insolation manage to warm south equatorial SST enough to weaken the cold tongue-ITCZ complex allowing the ITCZ to migrate southwards back towards the equator (Chang and Philander, 1994). In the Atlantic, cold conditions are not as stable as the rapid seasonal cooling abates in August and warming commences in September (Figure 1.3). This seasonal warming which cuts short the cool period in the Atlantic is seen to be in-phase with a deepening of the thermocline in the east as APE increases from August (Figure 3.18), a remotely forced thermocline depth influence apparently absent from the Pacific. This delayed negative feedback, similar to that operating on inter-annual time scales in the Pacific is discussed further in section 5.3 where the physical processes driving seasonal SST changes in the Atlantic are explored.

Figure 3.19 shows the correlations between inter-annual anomalies in equatorial Atlantic APE and Atl3 SST. The corresponding figure based on ORCA data, Figure A.11, can be found in Appendix A. A correlation coefficient of $r=-0.56$ ($r^*=-0.19$) is seen between the Atl3 index and equatorial APE based on SODA data (Figure 3.19) and one of $r=-0.62$ ($r^*=-0.19$) based on ORCA data (Figure A.11). The relationship is considerably weaker than the corresponding relationship in the Pacific (Figure 3.15). This result is in agreement with the weaker relationship observed between inter-annual SST anomalies and thermocline depth anomalies in the equatorial Atlantic (Zebiak, 1993; Vauclair and du Penhoat, 2001; Keenlyside and Latif, 2007). Unlike the Pacific, where inter-annual SST anomalies are predominantly forced by the dynamic Bjerknes feedback mechanism, a stronger contribution from external forcing and disparate local air-sea interactions (i.e. the WES feedback or the Ekman feedback) occurs in the Atlantic (Chang et al., 2006a). The Bjerknes feedback mechanism does not appear to be associated with all anomalous Atl3 SST events (Zebiak, 1993; Chang et al., 2006a).

There is however a distinct seasonal dependence in the correlation between inter-annual anomalies in equatorial Atlantic APE and Atl3 SST as depicted in Figures 3.20 and A.12. Figure 3.20 explores the seasonality in the relationship between inter-annual anomalies in equatorial Atlantic APE and Atl3 SST based on SODA data while Figure A.12 is based on ORCA data. Although there is some discrepancy between the two datasets, in both Figures 3.20 and A.12 anomalous Atl3 SST correlates best with anomalous APE between April and August. This finding is in agreement with seasonality in the regression between SST and thermocline depth anomalies (assessed using sea level anomalies as a proxy) obtained by Keenlyside and Latif (2007). Furthermore, this result is consistent with the fact that zonal mode events are seen to peak in boreal summer (Zebiak, 1993; Keenlyside and Latif, 2007). As elaborated in Section 5.2, seasonally boreal summer months coincides with the development of the cold tongue when subsurface cooling is the dominant process regulating Atl3 SST and the translation of subsurface anomalies into surface anomalies is more effective due to seasonally enhanced equatorial upwelling (Latif and Grötzner, 2000; Peter et al., 2006; Keenlyside and Latif, 2007). This seasonal dependence in the correlation suggests that anomalous SST events in the Atl3

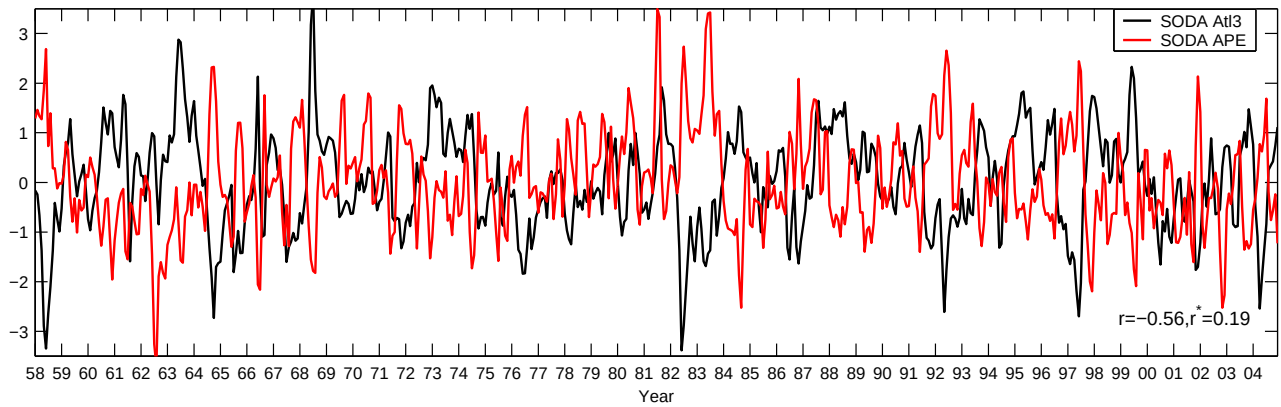


Figure 3.19: A comparison between inter-annual Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. SST and APE anomalies were calculated using SODA data. The correlation between SST and APE anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.11, can be found in Appendix A.

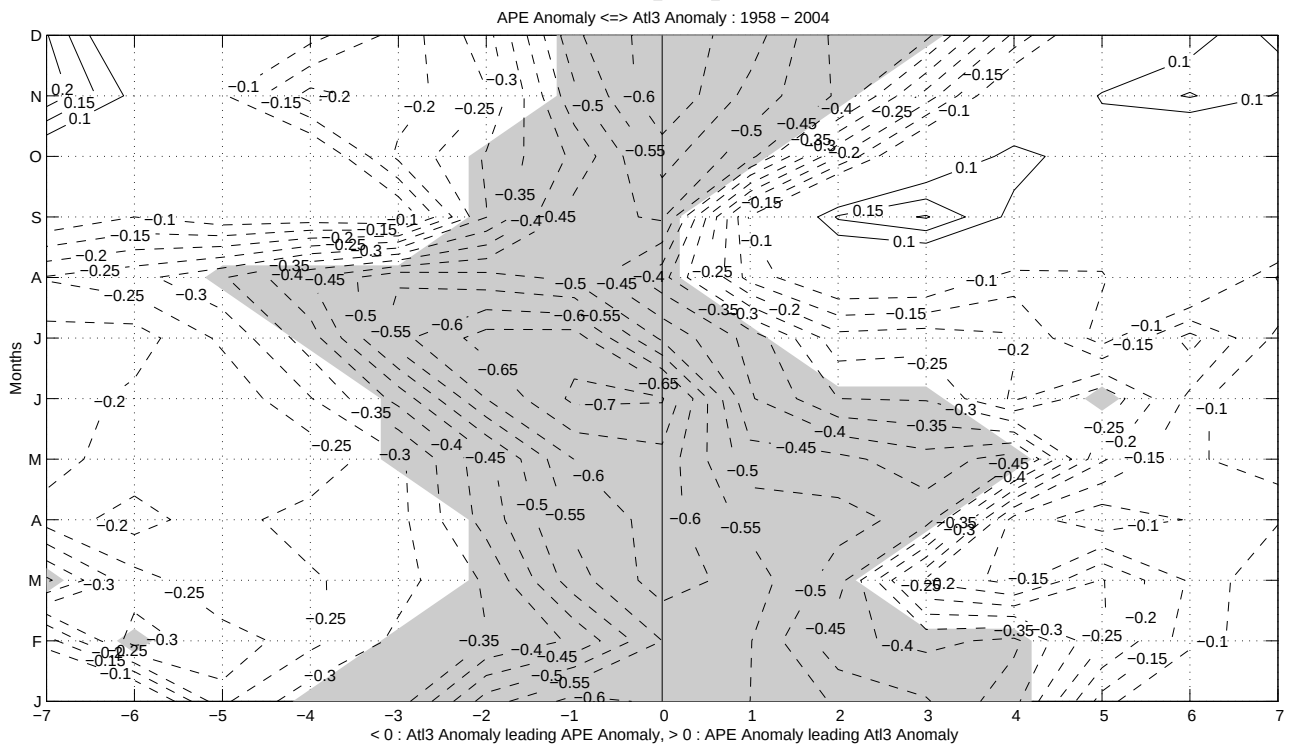


Figure 3.20: The seasonal dependence in correlations between Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. SST and APE anomalies were calculated using SODA data. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. APE values were calculated following Equation 3.2. The corresponding figure based on ORCA data, Figure A.12, can be found in Appendix A.

region forced by the Bjerknes feedback mechanism occur predominantly during boreal summer, while SST anomalies which occur outside of this season are more likely to be the result of remote atmospheric forcing or a different local air-sea interaction process.

3.3.4 Equatorial Atlantic vs Equatorial Pacific, APE - SST - Wind relationships

In this section, the relationship between variability in the Niño 3.4, Niño 3 and Atl3 SST indices, equatorial APE and zonal wind stress is explored at both the seasonal and inter-annual time scales within the respective basins. Considering that in the previous sections of this chapter the energetics results obtained from the SODA re-analysis data are seen to be consistent with those obtained based on the simulated ORCA data, the assessment of the relationship between SST indices, equatorial APE and zonal wind stress carried out in this section has been based purely on the SODA data. The SODA model was forced with the European Centre for Medium Range Weather Forecast (ECMWF) ERA-40 atmospheric re-analysis winds from 1958 to 2001 and with QuikSCAT winds from 2002 to 2004. Figure 3.21 shows the correlation between monthly zonal wind stress values used to forced the SODA ocean model and monthly Niño 3.4, Niño 3 and Atl3 time-series spanning 1958-2004. For each SST time series the climatological and inter-annual relationships with zonal wind stress across the basin are compared.

Evident when comparing Figures 3.21a and 3.21b is that the relationship between Niño 3.4 SST and zonal wind stress differs inter-annually from seasonally. While seasonal SST changes correlated best with local changes in zonal wind stress, inter-annual SST fluctuations correlate best with changes in zonal wind stress to the west of the Niño 3.4 region. The same applies to the relationship between Niño 3 SST and zonal wind stress (Figures 3.21c and 3.21d). This behaviour is completely consistent with the distinct primary mechanisms behind seasonal and inter-annual SST variability in the central-eastern Pacific. As highlighted in Section 3.3.2, seasonal SST fluctuations are governed principally by seasonal variations in the intensity of local winds, associated Ekman pumping and the Ekman feedback mechanism, while inter-annual fluctuations on the other hand are generally accompanied by anomalous wind to the west and remotely forced thermocline depth perturbations as they are grown by the Bjerknes feedback.

Turning to the Atlantic, the seasonal relationship between Atl3 SST and zonal wind stress is shown in Figure 3.21e. Unlike in the Pacific, seasonal central-eastern basin SST changes correlate best not with local zonal wind stress changes, but with changes in zonal wind stress to the west. This attribute is consistent with the fact that remotely forced thermocline variability is seen, in Section 5.2, to play a more significant role than locally forced Ekman pumping in seasonally modulating the vertical advection of cooler subsurface waters into the surface mixed layer.

The similar meridional structure seen seasonally in both basins (Figures 3.21a, 3.21c and 3.21e) is associated with the meridional movement of the ITCZ. In both the Pacific and the Atlantic basins, the development of the cold tongue is associated with a northward movement of the ITCZ and an associated change in the wind stress.

The spatial pattern in the correlation between climatological zonal wind stress and Atl3 SST seen in Figure 3.21e also suggests that, like the inter-annual scenario in the Pacific, easterlies in the western equatorial Atlantic are sensitive to seasonal Atl3 SST changes. The sensitivity of the seasonal development of zonal winds in the western-central equatorial Atlantic to zonal SST gradients which develop as a result of the

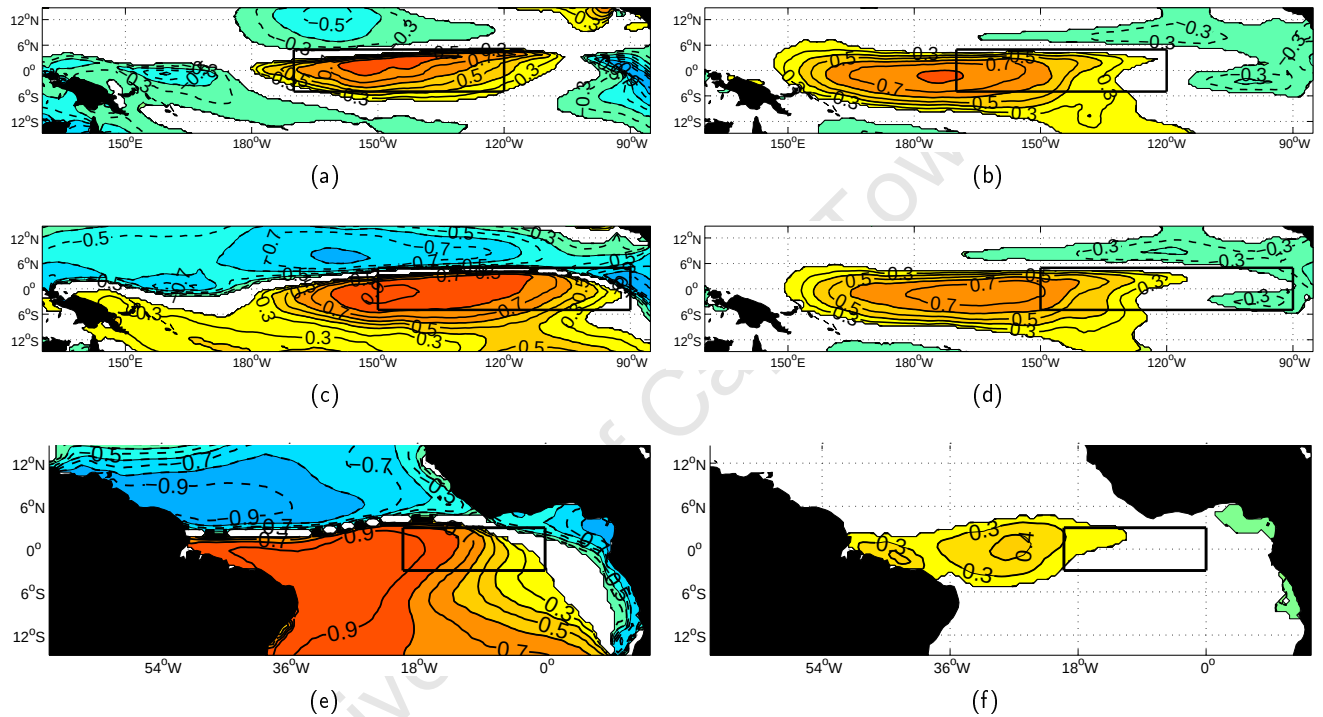


Figure 3.21: Spatial maps, based on SODA re-analysis data, of the correlation between monthly zonal wind stress values and monthly Niño 3.4, Niño 3 and Atl3 time-series spanning 1958-2004. (a) Climatological zonal wind stress and climatological Niño 3.4 SST. (b) Anomalous zonal wind stress and anomalous Niño 3.4 SST. (c) Climatological zonal wind stress and climatological Niño 3 SST. (d) Anomalous zonal wind stress and anomalous Niño 3 SST. (e) Climatological zonal wind stress and climatological Atl3 SST. (f) Anomalous zonal wind stress and anomalous Atl3 SST. Correlation values plotted are those significance at the 95% level.

seasonal development of the cold tongue has been demonstrated by Okumura and Xie (2004). Using an atmospheric general circulation model Okumura and Xie (2004) have demonstrated the sensitivity of the atmosphere to the rapid cooling of SSTs in the central-eastern equatorial Atlantic. During June-July-August, equatorial easterlies in the western-central Atlantic are greater in the control simulation forced by the full seasonal cycle in global SST than in the simulation where equatorial Atlantic SST is held constant in time from the 15 of April onwards.

Figure 3.22a shows the correlation between climatological monthly zonal wind stress values and monthly equatorial Atlantic APE values. In Figure 3.22a we see that the highest correlations between seasonal changes in equatorial Atlantic APE and zonal wind stress reside in the western-central equatorial Atlantic suggesting that seasonal wind changes in the west play a significant role in determining seasonal thermocline depth changes in the east, and hence the seasonal APE evolution. In Section 5.3, the physical processes driving the seasonal evolution of equatorial Atlantic APE are diagnosed and the role of zonal wind stress highlighted. Enhanced easterlies in the western-central equatorial Atlantic together with the sudden increase in westward zonal surface currents are seen to be responsible for the rapid increase in equatorial Atlantic APE between April and August as the cold tongue develops.

To complement Figure 1.4 which compares the nature of SST variability in the Atlantic against the Pacific and Figures 3.12 and 3.13 which compare the nature of APE variability in the Atlantic against the Pacific, Figure 3.23 compares the nature of western-central basin zonal wind stress variability in the Atlantic (3°S - 3°N 40°W - 20°W) against the Pacific (5°N - 5°S 160°E - 160°W). Consistent with the notion that the Bjerknes feedback mechanism is acting primarily at seasonal time-scales, the magnitude of seasonal fluctuation in western equatorial Atlantic zonal wind stress outweighs that of inter-annual variations. This contrast is in line with the hypothesis that seasonal cooling in the Atlantic's cool tongue region enhances zonal wind stress in the western equatorial Atlantic, facilitating the rapid increase in APE, cooling of SST and further strengthening of easterlies as a seasonally excited Bjerknes feedback results in large seasonal excursions in SST, APE and zonal wind stress.

Figure 3.21f illustrates the relationship between inter-annual anomalies in Atl3 SST and zonal wind stress and Figure 3.22b illustrates the relationship between inter-annual anomalies in equatorial Atlantic APE and zonal wind stress. In accordance with previous literature on the zonal mode based on both simulated and observed data, inter-annual SST fluctuations correlate best with anomalous zonal wind stress to the west

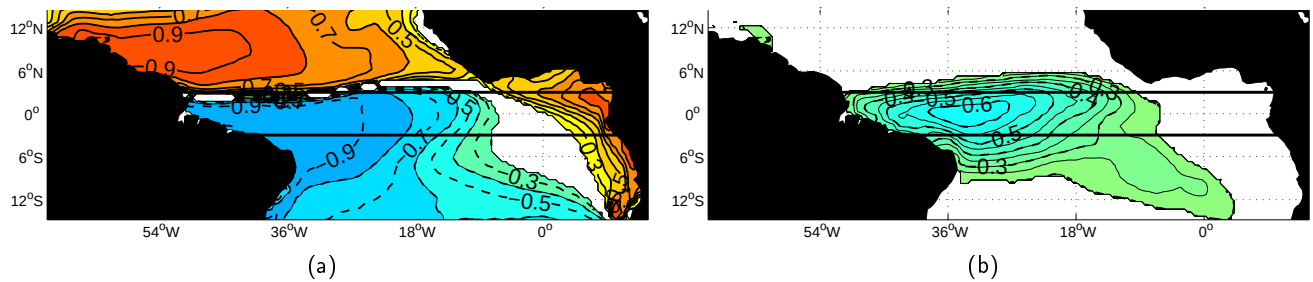


Figure 3.22: Spatial maps, based on SODA re-analysis data, of the correlation between monthly zonal wind stress values and monthly equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE spanning 1958-2004. (a) Climatological zonal wind stress and climatological APE, and (b) anomalous zonal wind stress and anomalous APE. Correlation values plotted are those significance at the 95% level. APE values were calculated following Equation 3.2.

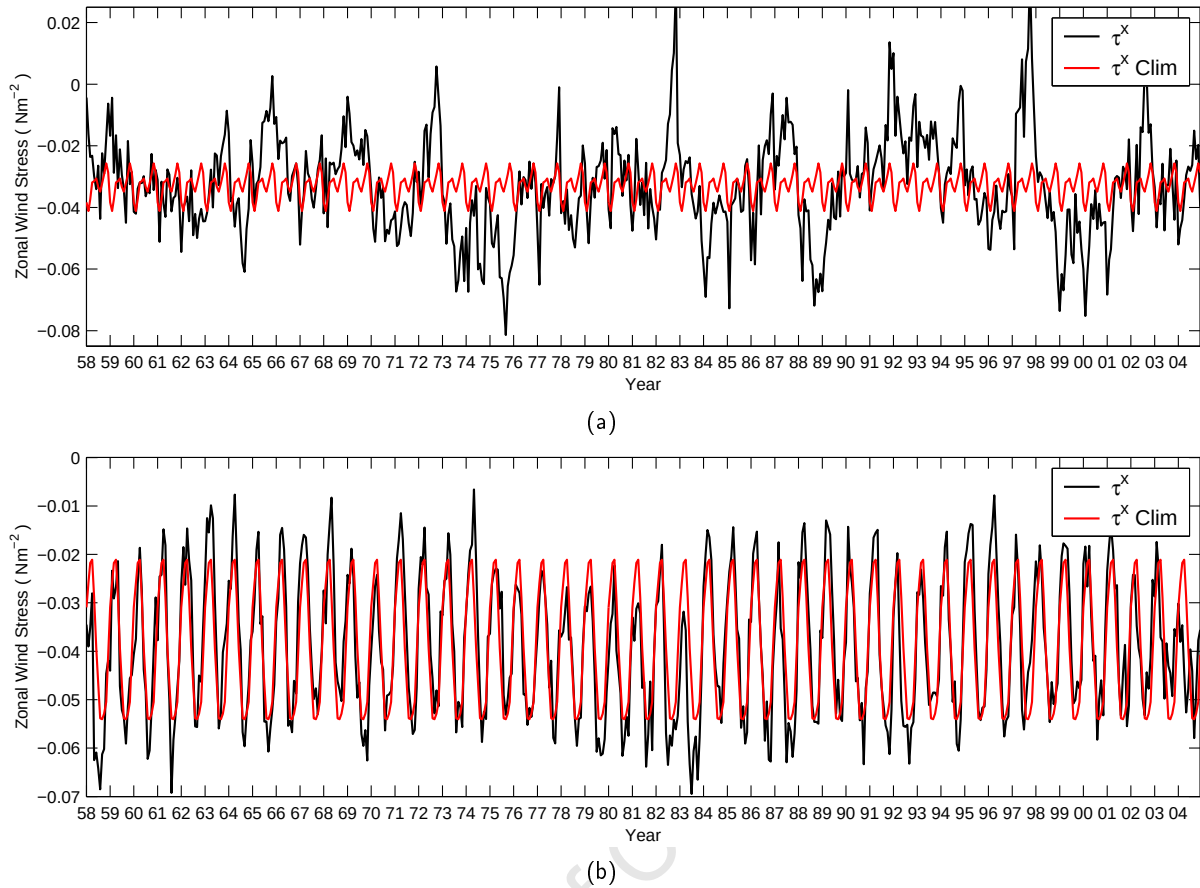


Figure 3.23: A comparison between variability in total versus climatological, (a) western-central basin zonal wind stress in the Pacific (5°N-5°S 160°E-160°W), and (b) western basin zonal wind stress in the Atlantic (3°S-3°N 40°W-20°W). Created using SODA wind stress forcing data. Wind stress values are given in Nm^{-2} and have been detrended.

of the Atl3 region (Zebiak, 1993; Ruiz-Barradas et al., 2000; Keenlyside and Latif, 2007). Furthermore the relationship between inter-annual anomalies in equatorial Atlantic APE and zonal wind stress depicted in Figure 3.22b supports the opinion that this wind-SST relationship is a result of the Bjerknes feedback mechanisms. Inter-annual Atl3 SST anomalies correlated with zonal wind stress anomalies to the west are the surface expressions of anomalous equatorial APE correlated with western-central Atlantic zonal wind stress anomalies. The wind-SST relationship depicted in Figure 3.21f is seen to be weaker than in the Pacific (3.21b and 3.21d). As mentioned previously this is believed to be the result of a larger contribution from external forcing and disparate local air-sea interactions to inter-annual variability in the Atlantic than in the Pacific (Chang et al., 2006a).

Figure 3.24 explores further the relationship between anomalous western Atlantic zonal wind stress (3°S-3°N 40°W-20°W) and Atl3 SST and Figure 3.25 the relationship between anomalous western Atlantic zonal wind stress (3°S-3°N 40°W-20°W) and equatorial Atlantic APE. A seasonal dependence is evident in the correlation between anomalous Atl3 SST and western Atlantic zonal wind stress as well as in the correlation between anomalous equatorial APE and western Atlantic zonal wind stress with correlation values peaking between April and August. Consistent with results obtained by Keenlyside and Latif (2007) based on observationally based findings, the strongest relationship between anomalous western Atlantic zonal wind stress and Atl3 SST is observed in May and the strongest relationship between anomalous western Atlantic zonal wind stress

and thermocline depth anomalies (equatorial APE) is observed in June-July.

In Figures 3.20 (Section 3.3.3), 3.24, and 3.25 which illustrate the seasonal dependence in correlations between Atl3 SST and equatorial Atlantic APE anomalies, zonal wind stress and Atl3 SST anomalies, and zonal wind stress and equatorial APE anomalies respectively, correlations are seen to peak between April and August, and lag-lead correlations of the same sign are roughly symmetric about zero lag during these months. Furthermore, Figure 3.26 shows the time lag-lead correlations between, (a) Atl3 SST and equatorial Atlantic APE, (b) Atl3 SST and western Atlantic zonal wind stress, and (c) western Atlantic zonal wind stress and equatorial Atlantic APE based on SODA data from 1958-2004. The lag-lead correlations shown in Figure 3.26 are all fairly symmetric about zero lag. These symmetric lag-lead correlations suggest that the covariability observed is the result of a positive feedback with anomalies in each variable reinforcing one another (Frankignoul and Hasselmann, 1977). On the other hand, if the lag-lead correlations were asymmetric or antisymmetric about zero lag, this would have suggested that they were the result of purely one-way forcing or a negative feedback (Frankignoul and Hasselmann, 1977). These results are consistent with the findings of Keenlyside and Latif (2007) who suggest that the Bjerknes feedback is active inter-annually in the equatorial Atlantic and that seasonal variations in the influence of SSTs on surface winds, as well as seasonal variability in subsurface-surface coupling, point to the fact that the Bjerknes feedback is predominately active between April and August. Here however, in light of results presented in this chapter suggesting that the Bjerknes is seasonally active between April and August, we suggest that anomalous SST events in the Atl3 region forced by the Bjerknes feedback mechanism occur predominantly during boreal summer as a modulation of an already seasonally active Bjerknes feedback. On the other hand, SST anomalies which occur outside of this season are more likely to be the result of remote forcing or different local air-sea interaction processes.

Also worth noting is the secondary peak seen between November and January in the seasonal dependence in correlations between anomalous Atl3 SST and equatorial Atlantic APE (Figure 3.20), anomalous zonal wind stress and Atl3 SST (Figure 3.24), and anomalous zonal wind stress and equatorial APE (Figure 3.25). This secondary peak corresponds with the secondary peak in inter-annual central equatorial Atlantic SST variability identified by Okumura and Xie (2006) as being associated with the November-December zonal mode of coupled variability.

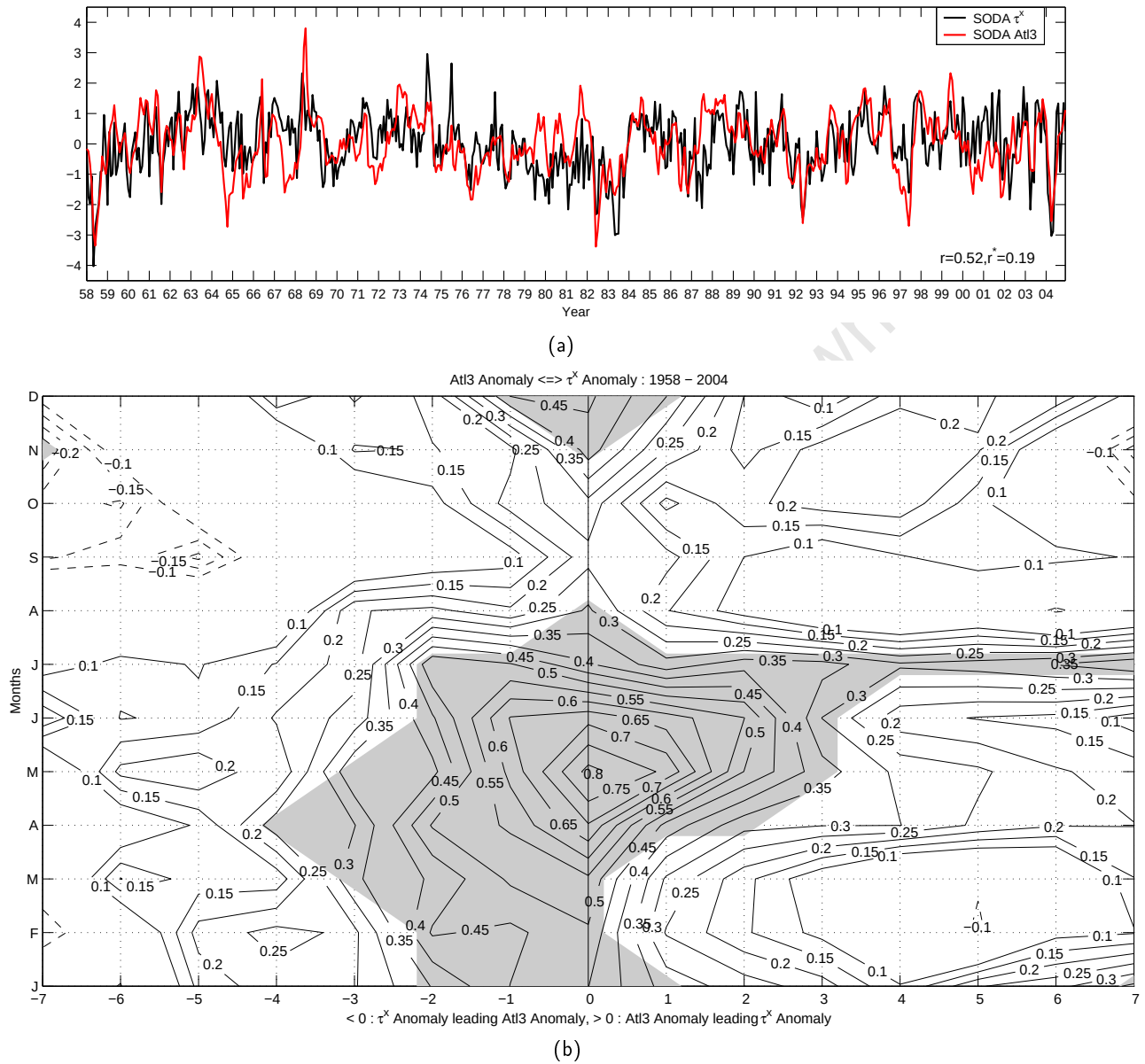


Figure 3.24: (a) A comparison between inter-annual Atl3 SST and western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies. The correlation between anomalies (r) is given together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). (b) The seasonal dependence in correlations between Atl3 SST and zonal wind stress anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST and zonal wind stress anomalies were calculated using SODA data and have been normalised and detrended.

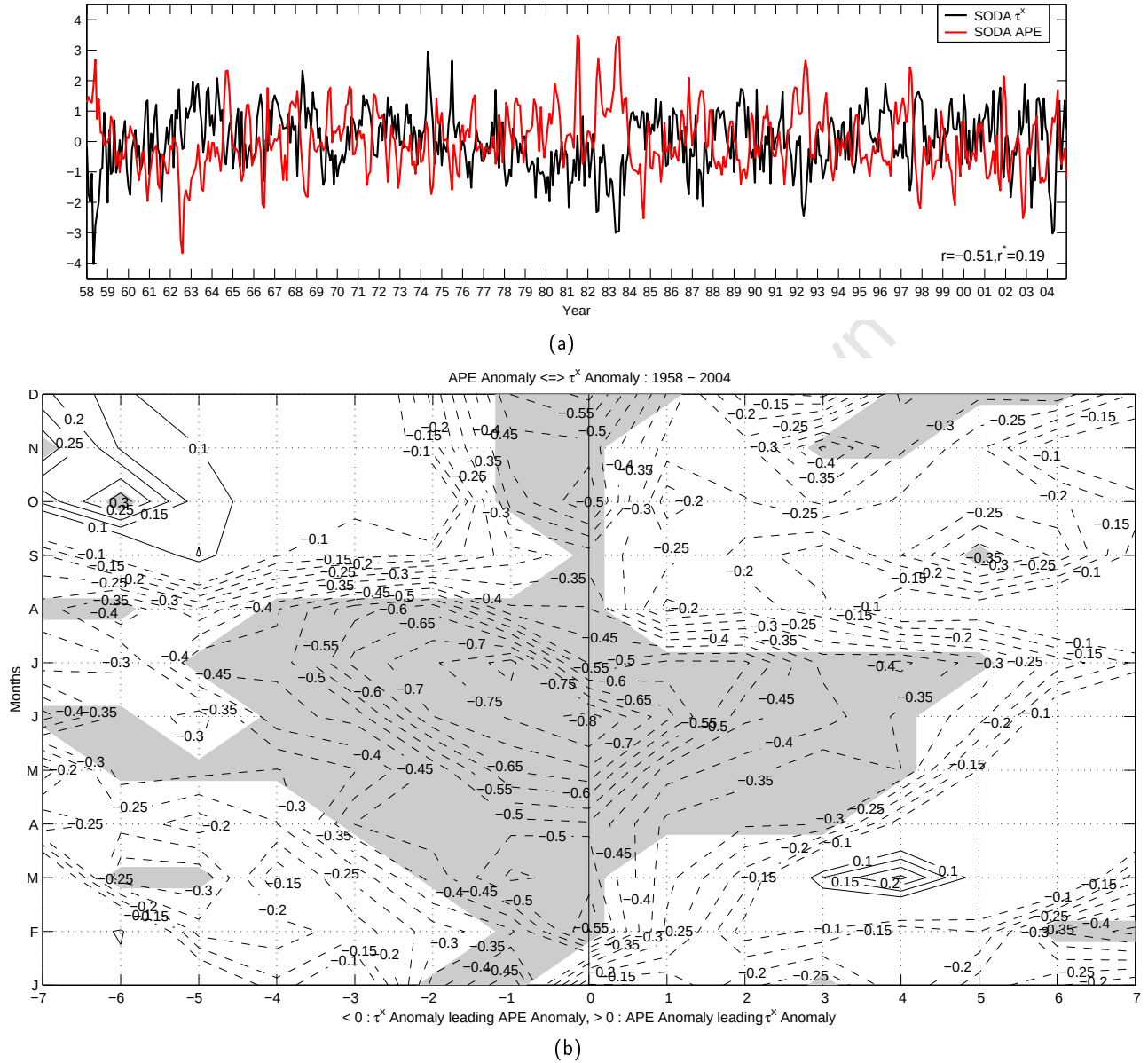
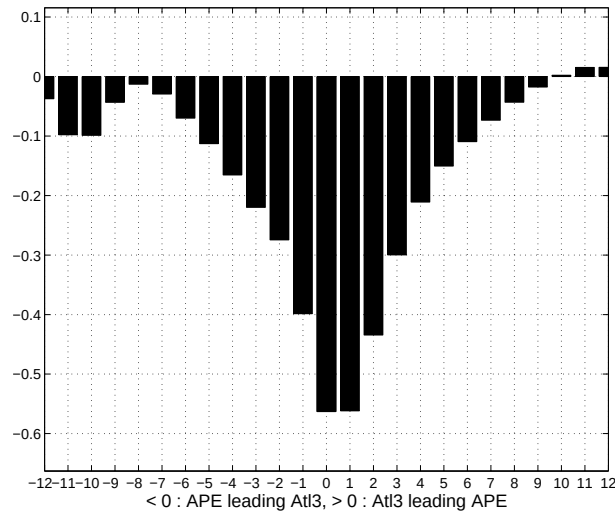
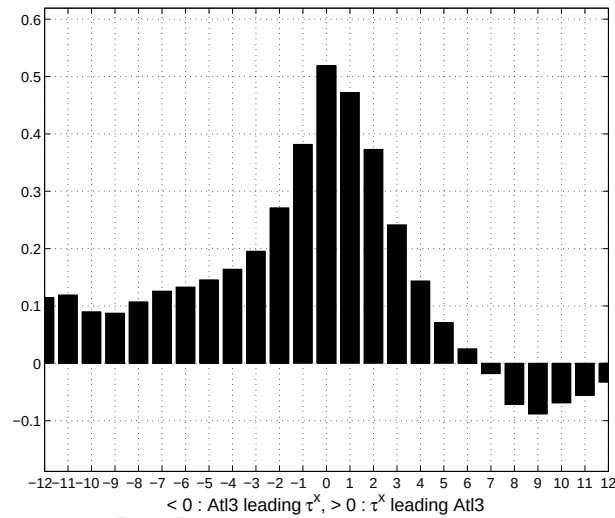


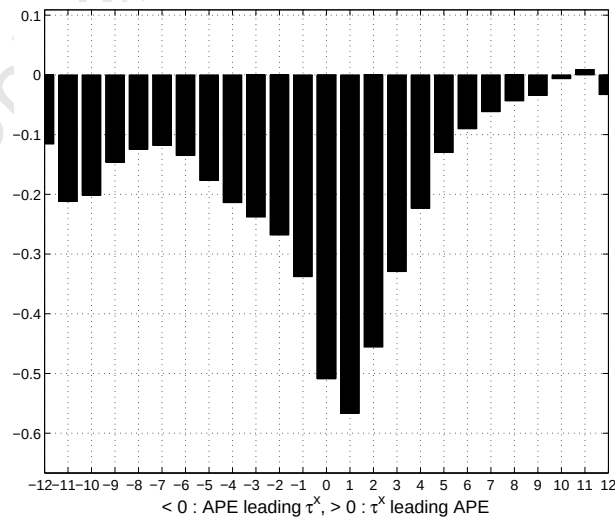
Figure 3.25: (a) A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE and western equatorial Atlantic (5°N - 5°S 40°W - 20°W) zonal wind stress anomalies. The correlation between anomalies (r) is given together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). (b) Seasonal dependence of correlation between APE and zonal wind stress anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. APE and zonal wind stress anomalies were calculated using SODA data and have been normalised and detrended. APE values were calculated following Equation 3.2.



(a)



(b)



(c)

Figure 3.26: Time lag correlations between anomalous (a) Atl3 SST and equatorial Atlantic APE (3°S - 3°N 60°W - 15°E 0-400m), (b) Atl3 SST and western equatorial Atlantic (5°N - 5°S 40°W - 20°W) zonal wind stress, and (c) equatorial Atlantic APE and western equatorial Atlantic zonal wind stress. SST, APE and zonal wind stress anomalies were derived from 1958-2004 SODA data and have been detrended. APE values were calculated following Equation 3.2.

3.4 Conclusions

This assessment wherein equatorial Atlantic variability is viewed from an energetics perspective and contrasted against the Pacific at both seasonal and inter-annual time scales, sheds light on the role of upper ocean variability within equatorial Atlantic SST variability. The presence of a strong SST-APE relationship in the Atlantic, like that observed in the Pacific, suggests that anomalous vertical advection in the central-eastern basin resulting from thermocline depth changes associated with the large scale redistribution of warm surface water, plays an equally important role in determining SST variability in the Atlantic as it does in the Pacific. However, unlike in the Pacific, the largest changes in APE occur seasonally in the Atlantic.

The strong seasonal SST-APE relationship observed in the Atlantic suggests that remotely forced thermocline depth changes play a crucial role in driving seasonal SST changes, a process which only plays a significant role inter-annually in the Pacific. In the Pacific, the primary coupled feedback mechanism associated with seasonal central-eastern basin SST variability is distinct from the that associated with inter-annual variability and this is reflected in distinct seasonal versus inter-annual relationship between SST, APE and zonal wind stress fluctuation. However, the relationship between seasonal SST, APE and zonal wind stress changes in the Atlantic supports the notion that the greater and steeper seasonal decrease in eastern-central basin SSTs observed in the equatorial Atlantic between April and August is due to the fact that the seasonally excited growth of asymmetric conditions about the equator is accompanied by a seasonally excited Bjerknes feedback as the thermocline shoals seasonally.

Furthermore, it appears that inter-annual Atl3 variability associated with the zonal mode may be best understood as the result of the modulation either in phase or amplitude of this seasonally active process. Unlike in the Pacific where the inter-annual ENSO signal dominates SST variability in the central-eastern basin, the annual cycle dominates in the Atlantic. This difference in the time-scale of the largest SST fluctuations between the two basins is attributed to the fact in the Pacific the Bjerknes feedback operates on inter-annual time-scales whereas it operates seasonally in the Atlantic. Seasonality in the occurrence of anomalous Atl3 events correlated with thermocline depth and zonal wind anomalies is in line with the hypothesis that inter-annual Atl3 variability associated in the zonal mode may be best understood as the result of the modulation, either in phase or amplitude, of a seasonally active Bjerknes feedback.

Chapter 4

Regional Ocean Modelling System - Tropical Atlantic Simulation

In order to facilitate a more in depth investigation into the energetics of the Atlantic, as well as the physical processes driving SST variability, a simulation of inter-annual variability in the tropical Atlantic from the 1980 to 2004 has been conducted using the Regional Ocean Modelling System. This simulation was conducted primarily to acquire the temperature, salinity and momentum evolution equation tendency terms required to calculate surface layer heat budgets and, evaluate the processes determining the APE evolution within a chosen domain (as outlined in Sections 5.2.2 and 5.3.2).

4.1 The Regional Ocean Modelling System

ROMS is a high resolution, split-explicit, free-surface, sigma (terrain-following) coordinate ocean model based on the primitive equations with the boussinesq and hydrostatic approximations made (Shchepetkin and McWilliams, 2005). ROMS utilises the split-explicit, hydrodynamic, computational kernel of Shchepetkin and McWilliams (2005), designed to improve computational efficiency in high resolution ocean models. This time-stepping kernel employs predictor-corrector integration algorithms together with a split-explicit time-stepping technique in which the time-step used for solving the fast barotropic mode is significantly less than that chosen for the slower baroclinic modes. The barotropic momentum equation and continuity equation are integrated several times within each baroclinic time-step to provide the baroclinic momentum and tracer equations with surface elevation and barotropic momentum fields. A weighted temporal averaging technique for barotropic fields provides tracer equations with consistent, conserved barotropic fields at the baroclinic time step. Temporal discretisation is based on a third order leap-frog predictor and Adams-Moulton corrector scheme. Enhanced solution stability achieved by this split-explicit hydrodynamic kernel allows for the use of increased time steps without compromising the accuracy of the solution (Shchepetkin and McWilliams, 2005). The Arakawa-C grid is employed for horizontal discretisation (Arakawa and Lamb, 1977) and sigma coordinates for vertical discretisation. To reduce the pressure gradient error associated with the discretisation of the horizontal pressure gradient in sigma coordinates, the pressure-gradient algorithm of Shchepetkin and McWilliams (2003) is employed.

4.2 Tropical Atlantic Configuration

The regional simulation analysed in the following chapters, hereafter referred to as ROMS-TAtl, has a Mercator grid extending zonally from 60°W-15°E and meridionally from 10°S-14°N with a horizontal resolution of $\sim 1/6$ th of a degree (longitudinal increment of $1/6$ th of a degree). The vertical grid consists of 50 sigma levels with a bottom stretching parameter of zero ($\theta_b = 0$) and surface stretching parameter of 5.5 ($\theta_s = 5.5$) and a minimum depth of 10m ($h_c = 10\text{m}$) (Song and Haidvogel, 1994). The model's topography was derived from the 1' General Bathymetric Chart of the Oceans (GEBCO) (www.gebco.net) using bilinear interpolation to convert to the model grid. Once on the model grid further smoothing of the GEBCO topography was undertaken to prevent the generation of pressure gradient errors. The topography was smoothed by reiteratively applying a selective Shapiro smoother to ensure a slope parameter ($r \sim \frac{\Delta h}{h}$) of less than 0.25 (Penven et al., 2008). A nonlinear equation of state was employed by the model to calculate density (Jackett and McDougall, 1995), while the reference density was defined as $\rho_o = 1025\text{kgm}^{-3}$.

The lateral momentum advection-diffusion scheme employed was the third-order upstream biased scheme (Shchepetkin and McWilliams, 1998) with lateral viscosity set to zero. For the tracer lateral advection-diffusion scheme, the RSUP3 scheme of Marchesiello et al. (2009) was chosen rather than the third-order upstream biased scheme as the simulated equatorial Atlantic thermocline was seen to be less diffuse when using the RSUP3 scheme. The RSUP3 scheme of Marchesiello et al. (2009) is designed to preserve water mass characteristics and prevent spurious diapycnal mixing associated with using a third-order upwind advection scheme in a sigma coordinate model while still preserving its low dispersion and diffusion capabilities. In the RSUP3 scheme, diffusion is split from advection and is represented by a rotated biharmonic diffusion scheme with flow-dependent hyperdiffusivity satisfying the Peclet constraint (Marchesiello et al., 2009). Subgrid-scale vertical mixing processes are parametrised by the non-local, K-Profile Planetary (KPP) boundary layer scheme of Large et al. (1994).

The adaptive algorithm of Marchesiello et al. (2001), where inward and outward information fluxes at the open boundaries are treated separately, was used to deal with the open boundaries. Inward fluxes are nudged towards external data with prescribed surface elevation, barotropic horizontal velocity, baroclinic horizontal velocity, temperature and salinity provided by the SODA re-analysis product (Carton et al., 2000; Carton and Giese, 2008). Outward fluxes are treated with a two-dimensional radiation algorithm (Marchesiello et al., 2001). The width of the sponge layer at the open boundaries was set to 150km and free slip boundary conditions were used at the closed boundaries (land). Surface vertical boundary conditions were provided by surface fluxes derived, based on the bulk formula of Kondo (1975), from atmospheric parameters supplied by the National Center of Environmental Prediction (NCEP)-Department of Energy, atmospheric model inter-comparison project re-analysis product (Kanamitsu et al., 2002).

The simulation was spun up for 10 years using lateral and surface boundary conditions for 1980, the first year of the simulation period. Thereafter the model was forced with inter-annually varying surface and lateral boundary conditions from 1980-2004. A time step of 960 seconds was used for the baroclinic mode, with 45 barotropic time-steps integrated for every baroclinic time-step (barotropic time-step ~ 21 seconds). Prognostic variables from the ROMS-TAtl simulation include surface elevation, barotropic and baroclinic horizontal velocity components, temperature and salinity, which have all been saved as two day averages and analysed at this temporal resolution. Furthermore, the temperature, salinity and momentum evolution equation tendency terms, required in Chapters 5 and 6 to calculate surface layer heat budgets and evaluate

the processes determining the APE evolution within a chosen domain, have been saved as two day averages and analysed at this temporal resolution.

4.3 Validation

To establish confidence in the ROMS-TAtl solution, simulated output is compared with available observations. Figure 4.1 compares the standard deviation of monthly, total SST variations, seasonal SST fluctuations and inter-annual SST anomalies based on simulated ROMS-TAtl SST versus observed Hadley OI SST (Rayner et al., 2003) data spanning 1980-2004. Comparing Figures 4.1a and 4.1b, 4.1c and 4.1d, and 4.1e and 4.1f, it is evident that the ROMS-TAtl simulation captures the observed spatial signature in SST variability at both seasonal and inter-annual time scales.

Further comparisons are made between simulated and observed temperature values in Figures 4.2, 4.3 and 4.4. Figure 4.2a, compares observed climatological values for the Atl3 SST index (3°S - 3°N 20°W - 0°E) derived from monthly Hadley OI SST (1980-2004) with those derived from ROMS-TAtl SST (1980-2004). While simulated seasonal variations in Atl3 SST display the same phase and amplitude as observed values, they exhibit a cold bias which at its maximum reaches the order of 0.5°C . In Figure 4.2b, observed and simulated inter-annual Atl3 SST anomalies are shown. The ROMS-TAtl simulation has managed to capture a significant percentage ($r=0.77$ $r^*=0.22$) of the inter-annual variability observed in Atl3 SST.

Figures 4.3 and 4.4 assess the accuracy with which the ROMS-TAtl simulation represents observed seasonal changes in the vertical temperature structure of the equatorial Atlantic. Figure 4.3 compares vertical cross-sections of temperature along the equator (1°N - 1°S) derived from climatological WOA05 data (Locarnini et al., 2006) with those derived from the ROMS-TAtl simulation. Figure 4.3a shows the month of April when the east-west slope of the thermocline is at its annual minimum and Figure 4.3b the month of August when the east-west slope is at its annual maximum. Figure 4.4 compares monthly climatological temperature profiles derived from PIRATA mooring data (Servain et al., 1998) at three sites along the equator $0^{\circ}\text{N}35^{\circ}\text{W}$, $0^{\circ}\text{N}10^{\circ}\text{W}$ and $0^{\circ}\text{N}0^{\circ}\text{E}$, with simulated values. While the ROMS-TAtl simulation reproduces the seasonal variations observed in the vertical temperature structure well, the thermocline is seen to be slightly more diffuse than in reality.

Figure 4.5 provides an indication of how well the ROMS-TAtl simulation has managed to reproduce seasonal SSH fluctuations north of the equator (3°N - 5°N), on the equator (1°S - 1°N), and south of the equator (3°S - 5°S) by comparing simulated climatological SSH anomalies with those derived from the 1/3° Aviso product (www.aviso.oceanobs.com). The observed and simulated climatological SSH anomalies shown in Figure 4.5 were derived from data spanning January 1993-December 2004 as this is the period when the two datasets overlapped. Simulated seasonal SSH variability corresponds well with observed variability at these three latitudes, with the ROMS-TAtl simulation capturing the dominant annual SSH signal in the western basin and the semi-annual signal in the east.

Figure 4.6 compares simulated seasonal equatorial Atlantic zonal surface currents averaged between 4°N - 4°S with those derived from the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). In Figure 4.6 the observationally based surface current climatology product shown is based on surface drifter observations spanning 28 October

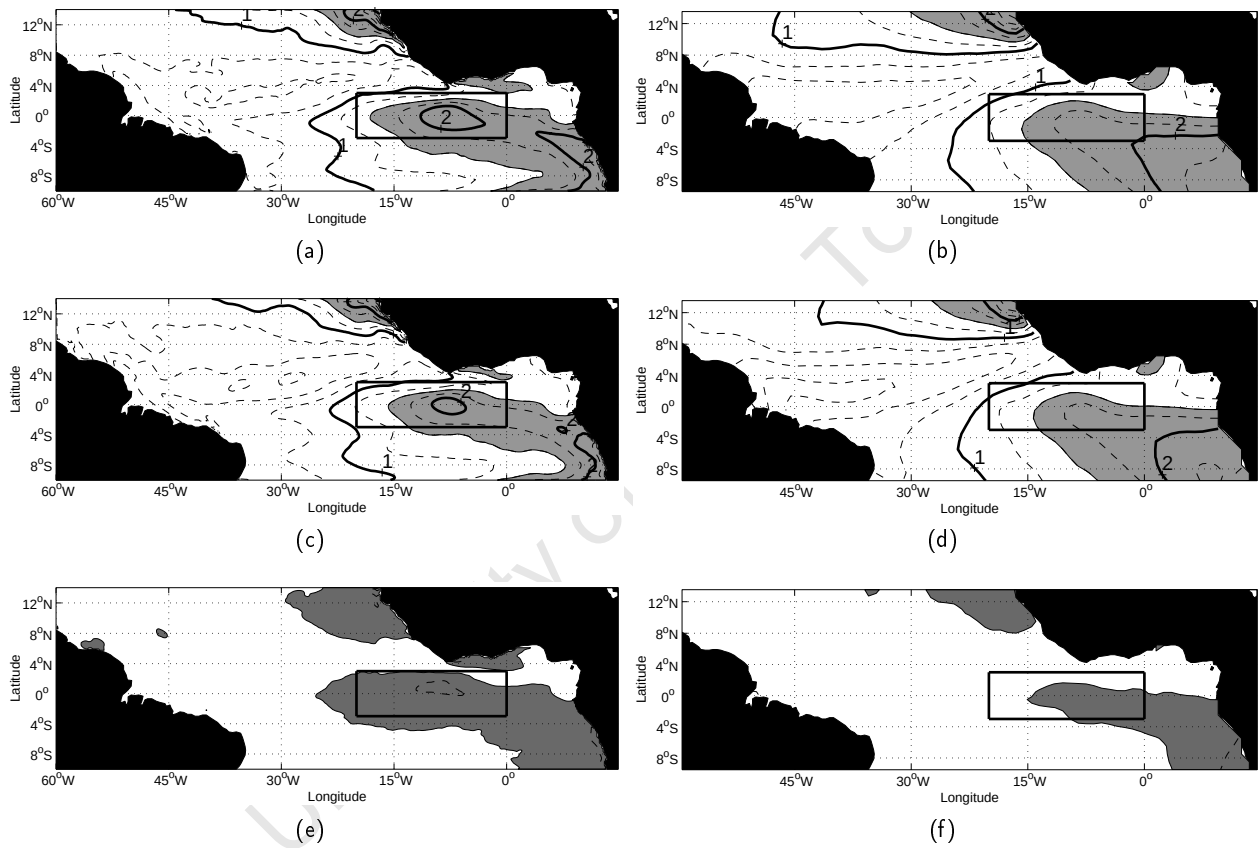


Figure 4.1: The standard deviation of monthly, (a) total TAtl-ROMS SST variations, (b) total Hadley OI SST variations, (c) seasonal TAtl-ROMS SST variations, (d) seasonal Hadley OI SST variations, (e) inter-annual TAtl-ROMS SST anomalies, and (f) inter-annual Hadley OI SST anomalies from 1980-2004. A contour interval of 0.25°C has been used, regions of >1.5°C are shaded in figures a-d and >0.5°C in figures e and f. The Atl3 (3°S-3°N 20°W-0°E) region has been demarcated.

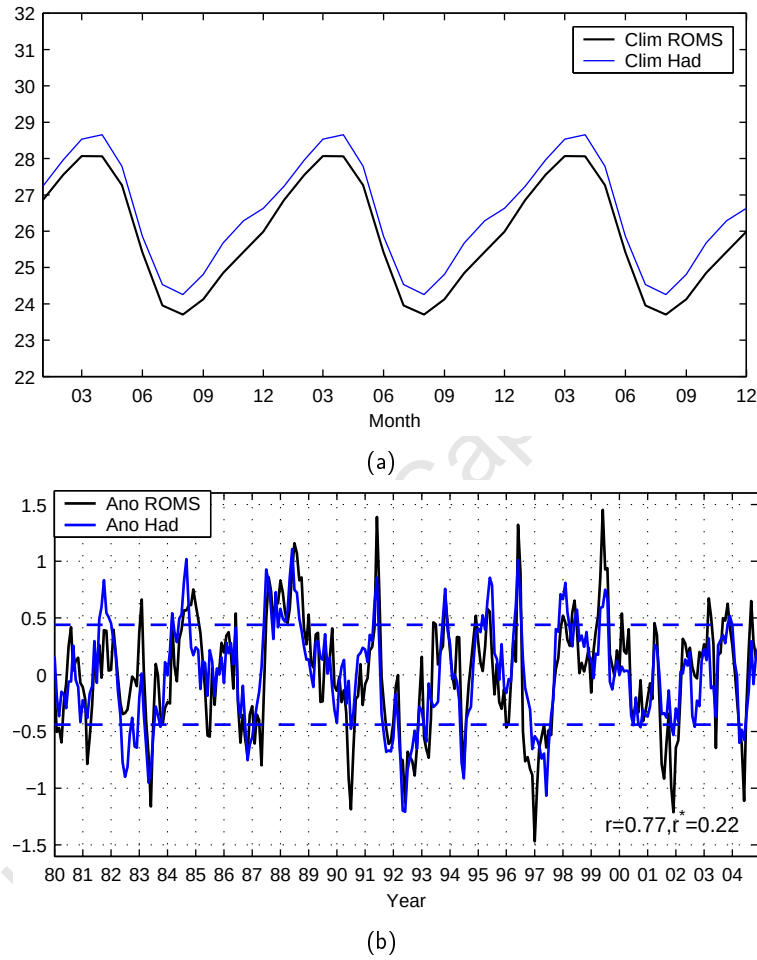


Figure 4.2: A comparison between the Atl3 (3°S-3°N 20°W-0°E) index created using Hadley OI SST data versus ROMS-TAtl SST data, (a) climatological Atl3 SST and (b) inter-annual Atl3 SST anomalies. The correlation between inter-annual anomalies in each SST index (r) is given on figure b, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST values are given in °C and have been detrended in figure b.

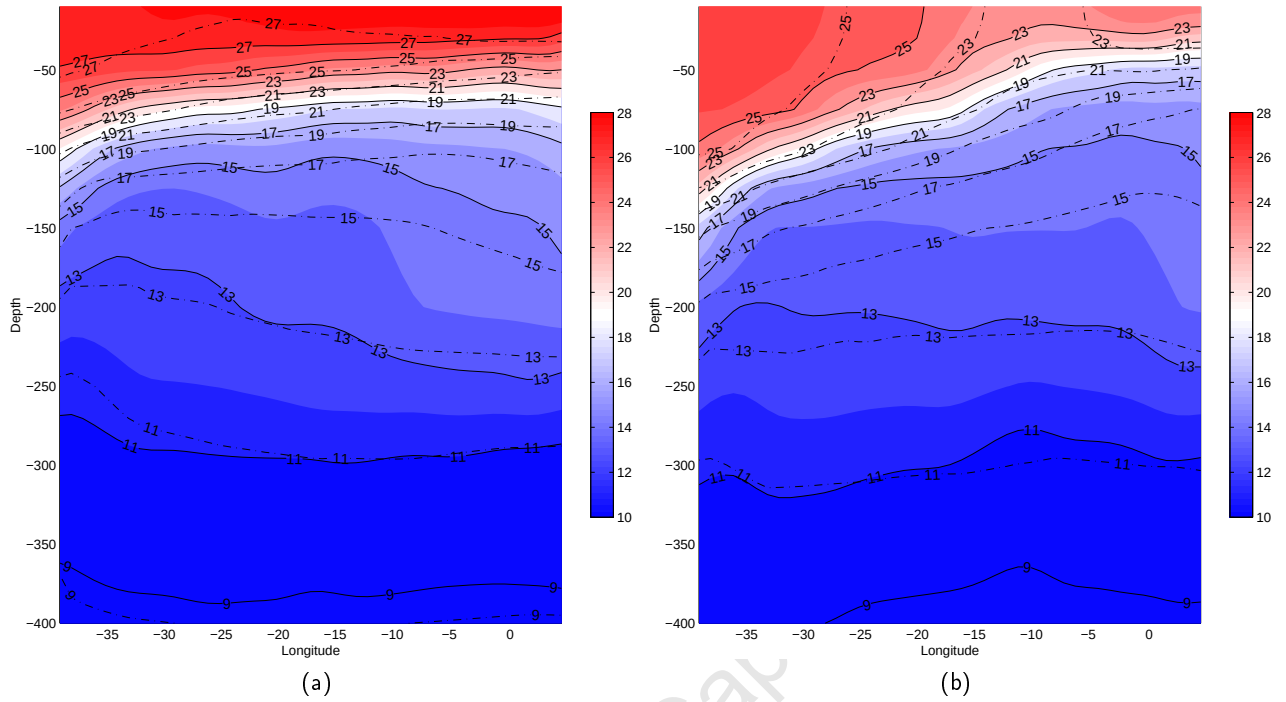


Figure 4.3: Climatological cross-sections in (a) April and (b) August of temperature ($^{\circ}\text{C}$) along the equator (1°N - 1°S) based on WOA05 data (Locarnini et al., 2006) (solid black and colour contours) and overlaid with data from the ROMS-TAtl simulation (dashed black contours).

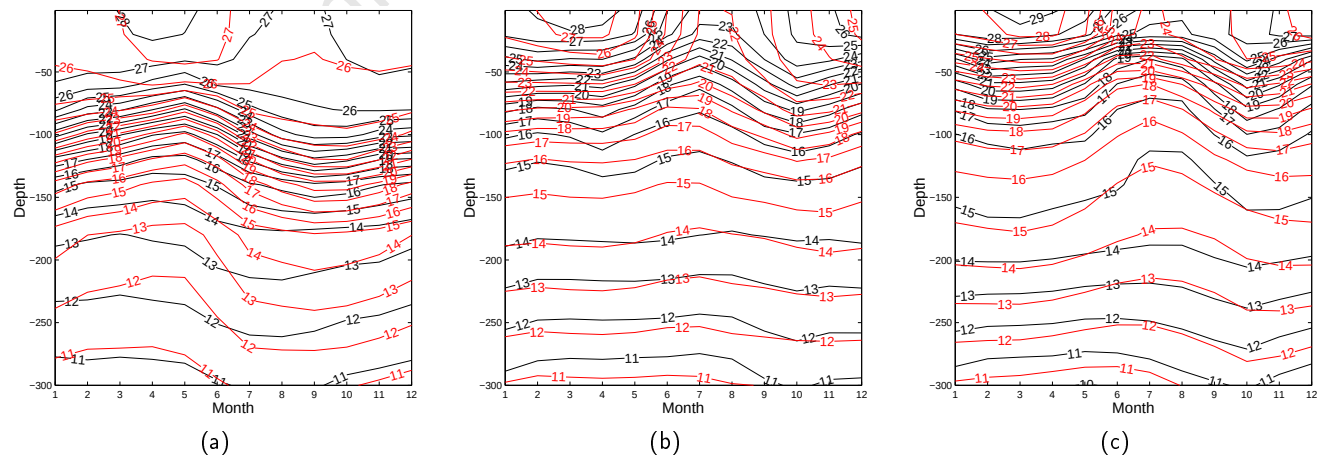


Figure 4.4: A monthly climatology of temperature profiles for three sites on the equator (a) $0^{\circ}\text{N}, 35^{\circ}\text{W}$, (b) $0^{\circ}\text{N}, 10^{\circ}\text{W}$ and (c) $0^{\circ}\text{N}, 0^{\circ}\text{E}$. (Black contours) PIRATA mooring data (Servain et al., 1998) (Red contours) ROMS-TAtl simulation.

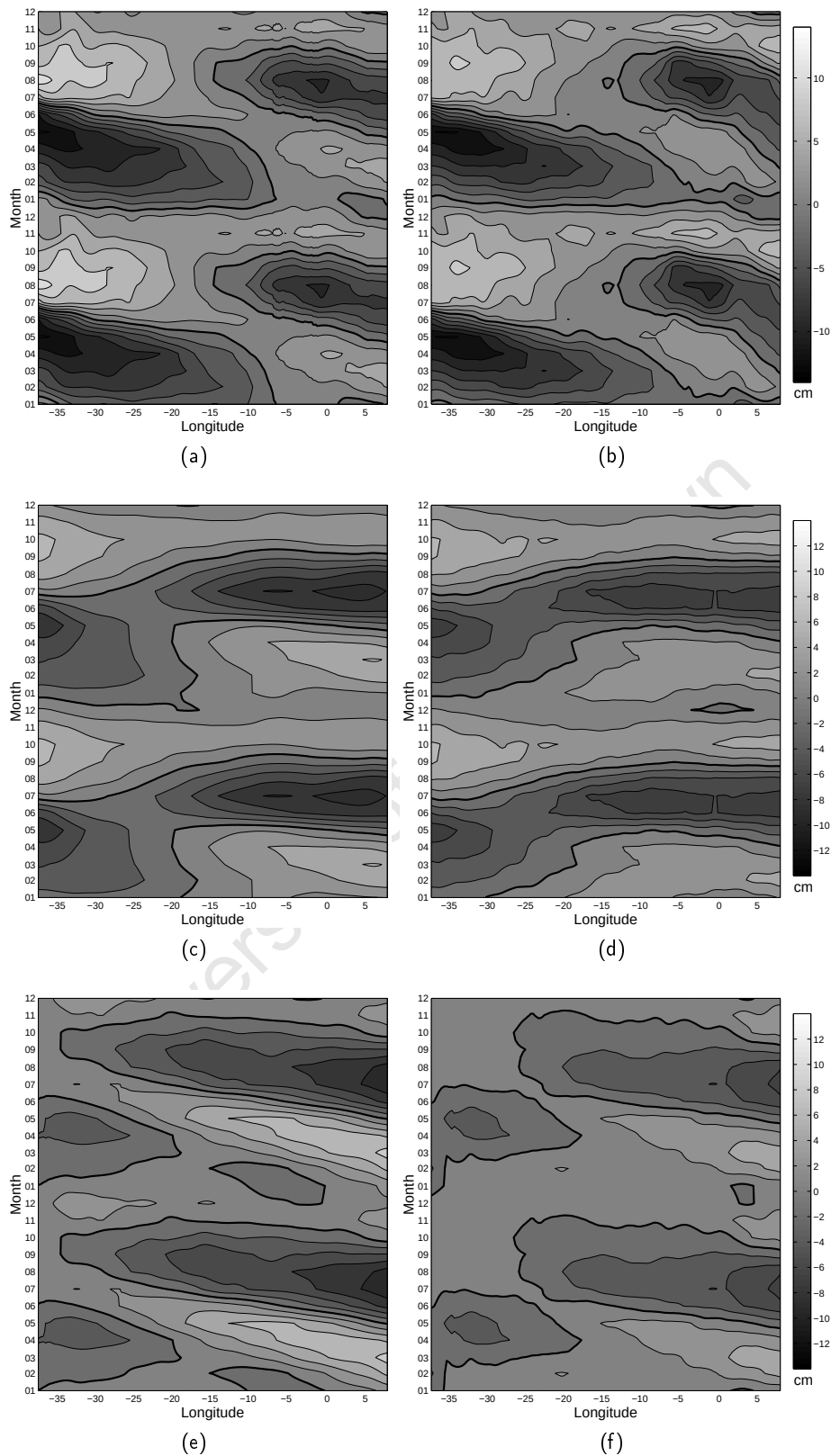


Figure 4.5: A comparison between (a) ROMS-TAtl, and (b) observed climatological SSH anomalies north of the equator (3°N-5°N); (c) ROMS-TAtl, and (d) observed climatological SSH anomalies along the equator (1°S-1°N); (e) ROMS-TAtl and (f) observed climatological SSH anomalies south of the equator (3°S-5°S). Observed climatological SSH anomalies were derived from 1/3° Aviso data (www.aviso.oceanobs.com) spanning January 1993-December 2004 and therefore the ROMS-TAtl SSH climatology shown here was derived from simulated SSH over the corresponding period.

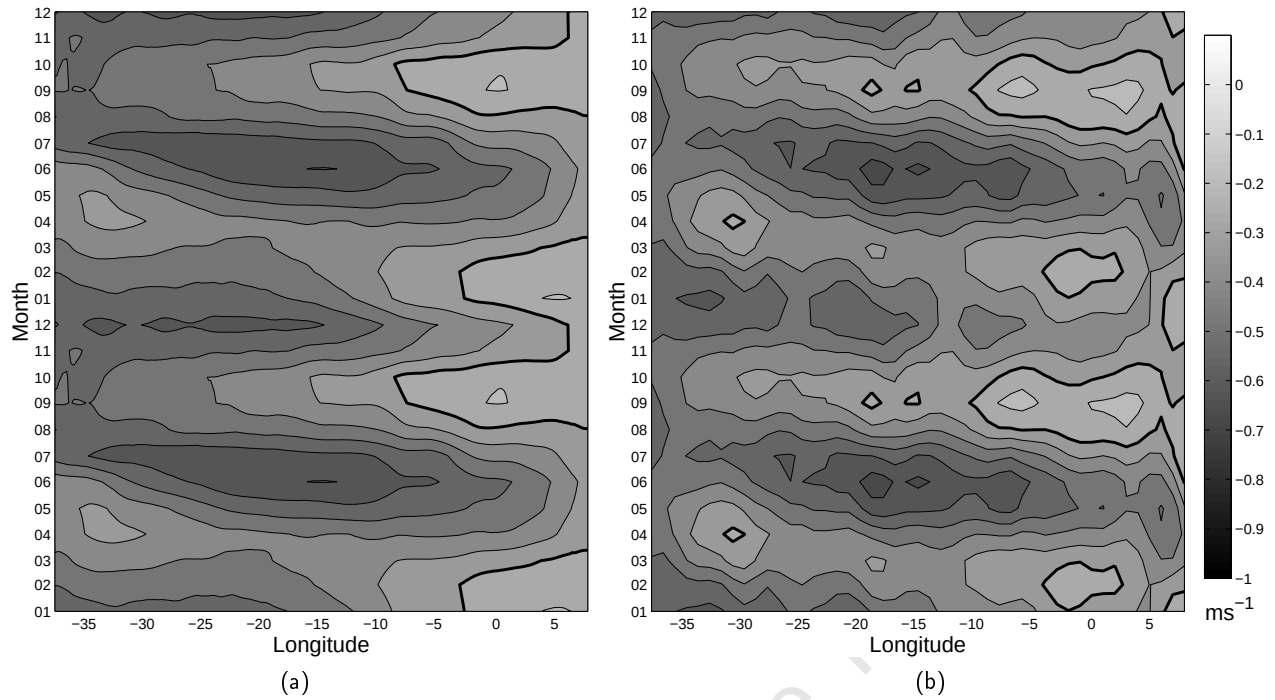


Figure 4.6: A comparison between (a) ROMS-TAtl, (b) observed, climatological, equatorial Atlantic, zonal surface currents averaged between 4°N-4°S. Observed zonal surface currents plotted in b where derived from the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html). This surface current climatology product is based on surface drifter observations spanning 28 October 1990 - 28 February 2004 and therefore the ROMS-TAtl zonal surface current climatology shown here was derived from simulated currents over the corresponding period.

1990 - 28 February 2004 and therefore the ROMS-TAtl zonal surface current climatology shown was derived from simulated currents over the corresponding period. The simulated seasonal equatorial surface current variability compares well with observed variability although, west of the Greenwich meridian, the simulated surface current variability appears to slightly underestimate westward surface current deceleration in September-October and overestimate westward surface current acceleration in November-December.

Figure 4.7 acts to establish how well the ROMS-TAtl simulation has managed to reproduce the mean structure of the EUC at three points along the equator. Figure 4.7 can be contrasted against a comparable figure show in Brandt et al. (2010) (Figure 2) based on observations. The ROMS-TAtl simulation captures the EUC with mean velocities of $>40\text{ms}^{-1}$ seen at the core of the EUC. Centred at approximately 100m in the west, the core of the EUC rises towards the east.

The results obtained in Chapters 5 and 6 further validate the ROMS-TAtl simulation which is found to be consistent with the results obtained in Chapter 3 that are based on SODA and ORCA data.

4.4 APE Evaluation

In Chapters 5 and 6, equatorial and tropical Atlantic APE is evaluated based on ROMS-TAtl output following Equation 3.2. Unlike in Chapter 3 where a non-linear equation of state was used when deriving density values from SODA and ORCA temperature and salinity data, a linear equation of state (Equation 5.9 in Section

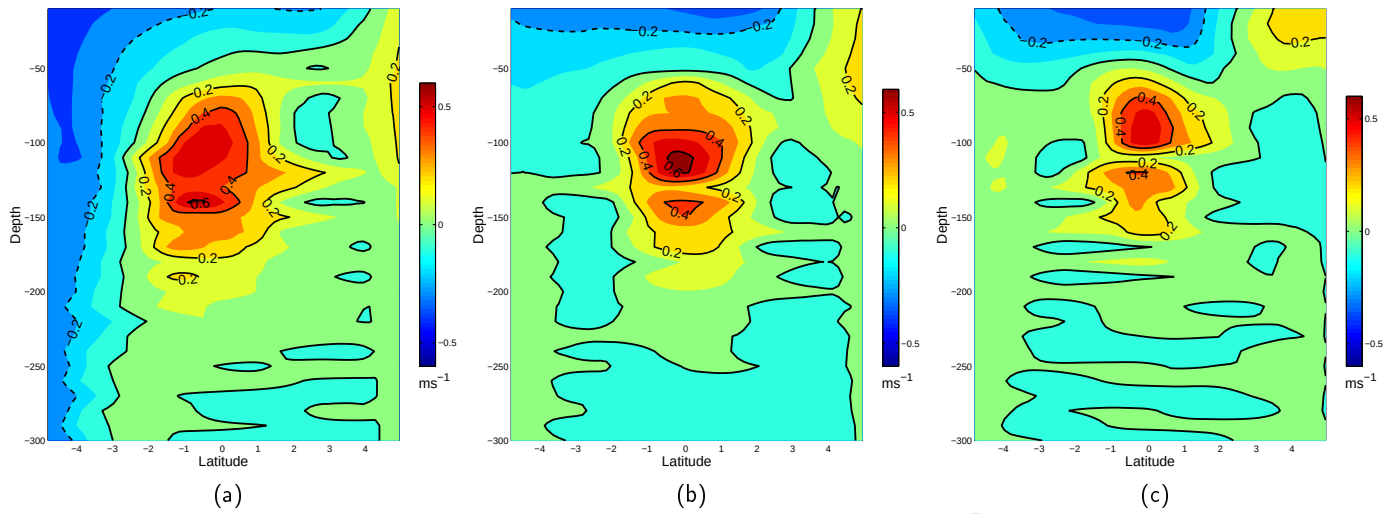


Figure 4.7: Meridional sections of the mean simulated zonal velocity field (given in ms^{-1}) at (a) 35°W , (b) 23°W and (c) 10°W .

5.3.2) has been used when deriving density values from ROMS-TAtl temperature and salinity data. As outlined in Section 5.3.2, the use of a linear rather than a non-linear equation of state facilitates the evaluation of an APE evolution equation by ensuring that density is conserved.

Figure 4.8 compares APE values evaluated following Equation 3.2 based on a density field that has been derived from simulated ROMS-TAtl temperature and salinity output following a non-linear equation of state (Jackett and McDougall, 1995) versus a linear equation of state (Equation 5.9 in Section 5.3.2). In Figure 4.8 the use of a linear equation of state is seen as being a valid approximation, having a minimal effect on the evaluation of APE fluctuations. The APE values seen in Figure 4.8 have had their mean subtracted from them.

4.5 Summary

In this chapter the ROMS-TAtl configuration, upon which the analysis carried out in Chapters 5 and 6 is based, has been introduced and validated. The validation undertaken within this chapter provides confidence in the ROMS-TAtl solution as seasonal and inter-annual variability in the tropical Atlantic has been reasonably reproduced.

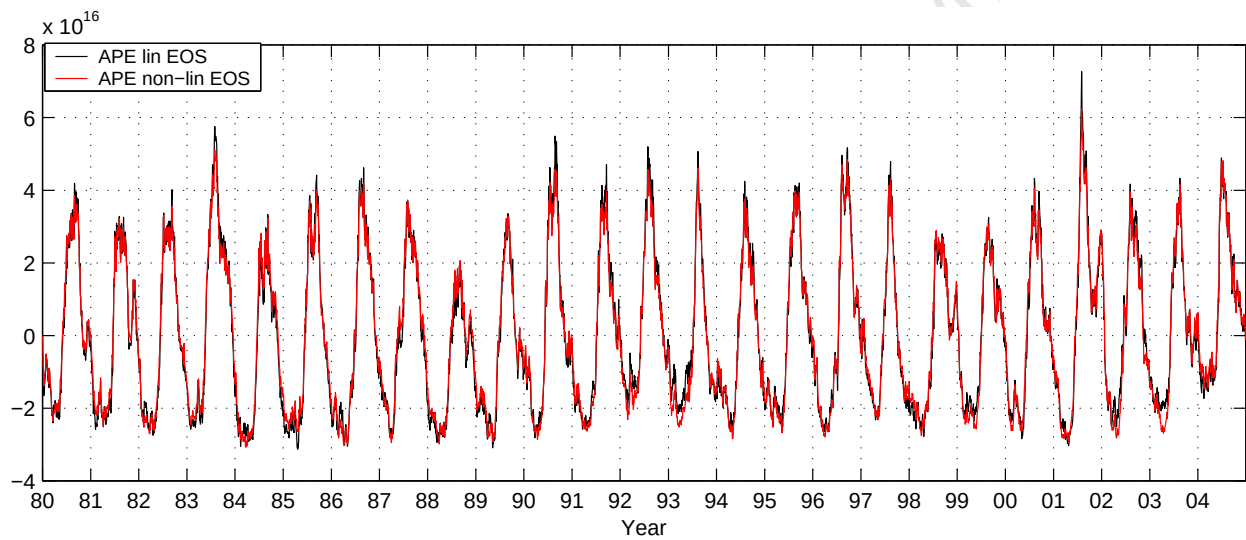


Figure 4.8: A comparison between equatorial Atlantic (3°S - 3°N 60°W - 15°E) APE values calculated based on density values derived from ROMS-TAtl temperature and salinity output using a non-linear equation of state, and APE values calculated based on density values derived using a linear equation of state. APE values were evaluated following Equation 3.2 and have had the mean subtracted.

Chapter 5

Tropical Atlantic Seasonal Cycle

5.1 Introduction

This chapter deals purely with gaining a better understanding of the physical processes associated with the seasonal cycle in the tropical Atlantic. It is aimed in particular at addressing the following questions raised in Chapter 1:

- What distinct processes occurring in the Atlantic may account for the observed differences in seasonal central-eastern basin SST signature between the Pacific and the Atlantic?
- Why do inter-annual SST fluctuations dominate over seasonal fluctuations in the central-eastern Pacific while the opposite is true in the Atlantic? Is it because, unlike in the Pacific, the Bjerknes mechanism acts primarily on seasonal time scales in the Atlantic?
- What is the role of ocean memory within central-equatorial Atlantic SST variability?

The analysis conducted in this chapter has been divided into two parts. The results obtained in Chapter 3 revealed a strong correlation between seasonal SST and APE changes within the equatorial Atlantic. This finding suggests that remotely forced thermocline depth changes play a crucial role in driving seasonal SST changes. The validity of this statement is researched in the first part of this chapter (Section 5.2) which investigates the relative contribution of each surface layer process to seasonal SST variability in the central-eastern equatorial Atlantic. Surface layer heat budgets within the Atl3 region have been conducted based on output from the ROMS-TAtl simulation as outlined in Section 5.2.2.

The results of the surface layer heat budgets conducted are discussed in Section 5.2.3 and highlight the key role played by subsurface cooling in seasonally regulating Atl3 SST. Furthermore, the role of seasonal variations in the strength of locally forced equatorial upwelling versus remotely forced thermocline depth changes in determining the degree of subsurface cooling is explored. Remotely forced thermocline depth changes due to the horizontal distribution of warm surface waters are seen to be paramount in the annual development of the equatorial Atlantic cold tongue.

Having shown in Section 5.2 that remotely forced thermocline depth variations play a significant role in driving seasonal SST variability, the following question is then raised - What are the primary processes

determining the seasonal evolution of APE in the equatorial Atlantic? The second part of this chapter, falling under Section 5.3, takes a deeper look into the mechanisms responsible for seasonal thermocline depth changes. This is achieved by investigating the processes dictating seasonal equatorial (3°S - 3°N 60°W - 15°E 0-400m) and tropical (8°S - 8°N 60°W - 15°E 0-400m) basin APE changes.

Section 5.3.2 describes the methodology employed in not only in this chapter but also in Chapter 6 when evaluating the processes determining APE evolution within an open boundary equatorial domain. An APE evolution equation is derived such that terms are decomposed in a manner relevant to diagnosing equatorial ocean dynamics in an open boundary domain. The approach used to estimate the contribution of each term based on output variables from the TAtl-ROMS simulation is detailed in this section.

The results of the seasonal APE evolution analysis are discussed next. Similar to results obtained at inter-annual time scales in the Pacific (Goddard and Philander, 2000), seasonal APE changes in the Atlantic are seen to be primarily forced by seasonal changes in buoyancy power. These seasonal changes in buoyancy power are in turn seen to be driven by seasonal changes in the work done by the wind on the ocean over the tropical Atlantic.

Consistent with the suggestion raised in Chapter 3 that the Bjerknes feedback is seasonally excited in the Atlantic, the sudden and rapid increase in APE between April - August is seen to be driven by increased wind work associated with the seasonal increase in zonal wind stress. Furthermore the decrease in APE observed in September-October is found to result from the influence of seasonal surface current fluctuations on the ability of seasonally enhanced surface winds to do work on the ocean. Inter-annually in the Pacific, thermocline perturbations associated with off-equatorial Rossby waves arriving in the west during the transition from one phase of ENSO to the next, affect surface currents and hence the ability of the wind to do work on the ocean, consistent with the delayed oscillator theory. It is proposed here that this ocean memory mechanism is seasonally active and acts as the delayed negative feedback process responsible for the decay of the seasonally excited Bjerknes feedback in the Atlantic.

5.2 Surface Layer Heat Budgets

5.2.1 Introduction

Previous studies have shown that seasonal SST fluctuations in the central-eastern equatorial Atlantic are regulated primarily by warming, due to the surface heat flux and oceanic eddies, and subsurface cooling (Weingartner and Weisberg, 1991; Carton and Zhou, 1997; Foltz et al., 2003; Peter et al., 2006). In this section, surface layer heat budgets within the Atl3 region (the calculation of which is based on results from the ROMS-TAtl simulation as outlined in Section 5.2.2) have been conducted. The goal is to facilitate a comparison with not only previous results but also the energetics of the equatorial Atlantic in an attempt to establish the role of remotely forced ocean dynamics within seasonal central-eastern equatorial Atlantic SST variability.

Remotely forced SST changes in the central-eastern equatorial basin, originating from thermocline depth changes associated with the large scale redistribution of warm surface water / APE changes, are merely surface expressions of subsurface changes in the temperature structure. These subsurface changes are driven by the zonal, meridional and vertical advection of temperature as warm surface waters are redistributed.

Adiabatic advection terms will therefore be active in driving SST changes that are remotely forced. In the central-eastern equatorial Atlantic where the thermocline is shallow, vertical velocities near the surface act on a strong vertical temperature gradient and as a result subsurface variability largely manifests as fluctuations in vertical advection. SST changes in this region that are remotely forced would therefore predominantly be established through changes in vertical advection. As a result, this section investigates the role of vertical advection in driving seasonal SST variability in the Atl3 region.

Two different surface layer definitions are used to diagnose the relative contribution of each surface layer process to seasonal SST variability in the central-eastern equatorial Atlantic. Firstly the surface mixed layer, the depth of which, $h(t,x,y)$, is defined according to the mixed layer depth criterion of Lorbacher et al. (2006). This mixed layer depth criterion is based on the shallowest extreme curvature of near surface layer temperature profiles and deals better with a dataset that has varying vertical resolutions (as is the case with the ROMS-TAtl vertical sigma co-ordinates) and a large variety of observed stratification profiles than defining the mixed layer depth as the depth at which temperature has increased by a certain amount from its surface value. Secondly, a deeper subsurface layer of fixed depth, $h=70\text{m}$, which corresponds to the mean depth of the thermocline over the Atl3 region. The reasoning behind conducting a heat budget for the subsurface layer as well as the surface mixed layer is to resolve the vertical advection term. Vertical advection is a small term in the surface mixed layer budget as turbulent processes dominate near the surface and simulated vertical velocities at the base of the surface mixed layer are weak. For the shallow surface mixed layer heat budget, subsurface cooling takes the form of vertical mixing. The temperature of water at the base of the surface mixed layer that is mixed into the surface mixed layer is however largely determined by vertical advection acting below the depth of the surface mixed layer, which the second surface layer heat budget of the deeper subsurface layer resolves. Furthermore, subsurface temperature variations driven by fluctuations in vertical advection at the base of the subsurface layer will only reflect in the SST field if they are translated into the surface mixed layer by vertical mixing at the base of the surface mixed layer. It is therefore important to resolve both fluctuations in the strength of subsurface vertical advection as well as fluctuations in the strength of vertical mixing at the base of the surface mixed layer, to establish the influence of remotely forced thermocline depth variability on SST.

Changes in the magnitude of vertical advection may result from either locally forced changes in vertical Ekman velocities or remotely forced thermocline depth changes. On having resolved the contribution of vertical advection to seasonal SST changes within the Atl3 region, the impact of locally forced changes in vertical Ekman velocities relative to remotely forced thermocline depth changes (quantified using equatorial APE values) on the intensity of vertical advection is investigated. To facilitate this investigation, vertical velocities induced by local wind driven Ekman divergence have been estimated following the methodology outlined in Section 5.2.2 and equatorial APE values have been evaluated using the ROMS-TAtl output.

Following Section 5.2.2, in which the methods employed in this section are outlined, the results are discussed in Section 5.2.3 and conclusions summarised in Section 5.2.4.

5.2.2 Methodology - Surface Layer Heat Budget Calculation and Estimating Ekman Vertical Velocities

5.2.2.1 Surface Layer Heat Budget Calculation

The relative contribution of each process in forcing the temporal evolution of the mean temperature ($\bar{T}$) within a surface layer of depth $h(t,x,y)$, $\frac{\partial \bar{T}}{\partial t}$, can be quantified by deriving an evolution equation (Equation 5.1) as in Vialard and Delecluse (1998a) and outlined in Appendix B:

$$\begin{aligned} \frac{\partial \bar{T}}{\partial t} = & - \underbrace{\frac{1}{h} \int_{-h}^0 u \frac{\partial T}{\partial x} dz}_{A_x} - \underbrace{\frac{1}{h} \int_{-h}^0 v \frac{\partial T}{\partial y} dz}_{A_y} - \underbrace{\frac{1}{h} \int_{-h}^0 w \frac{\partial T}{\partial z} dz}_{A_z} \\ & + \underbrace{\frac{1}{h} \int_{-h}^0 K_{th} \nabla^2 T dz}_{D_h} - \underbrace{\frac{1}{h} \left\{ \frac{\partial h}{\partial t} (\bar{T} - T_{z=-h}) - (K_{tv} \frac{\partial T}{\partial z})_{z=-h} \right\}}_B + \underbrace{\frac{q_* + q_s(1 - f_{z=-h})}{h \rho_o C_p}}_F \end{aligned} \quad (5.1)$$

Where $T_{z=-h}$ is the temperature at the base of the chosen surface layer, K_{th} represents the horizontal temperature diffusion coefficient, K_{tv} the vertical temperature diffusion coefficient, q_s the surface solar radiative flux, $f(z)$ the fraction of solar radiation reaching depth z and q_* is the non penetrative part of the surface heat flux consisting of the net long-wave radiative flux, the latent heat flux and the sensible heat flux components. Each term represents a physical processes which may contribute to the variability of the mean temperature ($\bar{T}$) within the chosen surface layer: zonal advection A_x , meridional advection A_y , vertical advection A_z , horizontal diffusion D_h , entrainment and vertical mixing at the base of the layer B , and surface heat flux forcing F .

In this study, surface layer heat budgets for the Atl3 region have been calculated following Equation 5.1 by vertically integrating two day averages of each term within the temperature evolution equation (Equation B.4) over the chosen surface layer depth. Equation 5.1 has been calculated for each grid point in the horizontal plane and an area weighted average derived for the Atl3 region.

5.2.2.2 Estimating Ekman Vertical Velocities

To establish an estimate of the vertical velocities induced by local wind driven Ekman divergence, a method which has been employed in previous studies (Zebiak and Cane, 1987; Sterl and Hazeleger, 2003; Okumura and Xie, 2004; Barreiro et al., 2005) is used to estimate Ekman layer transports near the equator. Two day averages of the ROMS-Tatl wind stress forcing, have been used to calculate vertical velocities at the base of the Ekman layer (here after referred to as the Ekman vertical velocity W_E) at each ROMS-Tatl grid point following Equations 5.2, 5.3 and 5.4.

$$W_E = \frac{\partial U_E}{\partial x} + \frac{\partial V_E}{\partial y} \quad (5.2)$$

$$U_E = \frac{f \tau_y + r \tau_x}{\rho(f^2 + r^2)} \quad (5.3)$$

$$V_E = \frac{-f\tau_x + r\tau_y}{\rho(f^2 + r^2)} \quad (5.4)$$

where, W_E is the Ekman vertical velocity, U_E is the zonal Ekman transport, V_E is the meridional Ekman transport, ρ is the density of sea water, τ_x the zonal wind stress, τ_y the meridional wind stress and f is the Coriolis parameter. The coefficient r is the decay constant given to the linear friction term representing the loss of energy due to shear stresses within the surface currents. The linear friction term is introduced since $f \rightarrow 0$ at the equator. The derivation of Equations 5.2, 5.3 and 5.4 is given in Appendix C. Following Zebiak and Cane (1987) a value of $0.5d^{-1}$ is given to r as in Sterl and Hazeleger (2003); Okumura and Xie (2004); Barreiro et al. (2005). Ekman vertical velocities calculated for each grid point in the horizontal plane are then area weighted averaged over the Atl3 region, to provide a measure of the intensity of locally forced Ekman pumping for the Atl3 region.

5.2.3 Results

Seasonal variations in the temperature of the surface mixed layer and SST for the Atl3 region are compared in Figure 5.1a. Seasonal variations in the mean temperature of the surface mixed layer correspond closely to seasonal SST changes (Figure 5.1a). As shown in Figure 5.1b, the dominant terms forcing these seasonal changes in the mean temperature of the surface mixed layer are the surface heat flux and vertical mixing. Note that a 14 day running mean has been applied to each time series in Figure 5.1 to smooth out the high frequency variability. The horizontal advection terms play a secondary but still significant role, while contributions from the vertical advection, horizontal mixing and entrainment are minimal. Simulated vertical velocities at the base of the surface mixed layer are weak as turbulent processes dominate, hence vertical advection is a small term and subsurface cooling takes the form of vertical mixing at this depth. As discussed in the appendix of Menkes et al. (2006), the negligible effect of entrainment is due to the distinction between the mixed layer and the mixing layer depths (Brainerd and Gregg, 1995). In the model, the depth down to which vertical mixing is strong is the mixing layer depth which can vary substantially within a few time steps of the simulation. With the exception of barrier layer regions (Vialard and Delecluse, 1998b), the simulated mixed layer is normally shallower than the mixing layer (Menkes et al., 2006) which means that vertical mixing at the base of the mixed layer is strong. If one were to perform a surface layer heat budget down to the depth of the models mixing layer, the vertical mixing term would be negligible and entrainment strong. As the focus here is on the relatively slow variability associated with the seasonal cycle and not the higher frequency variability associated with diurnal changes in the mixing depth, the use of surface mixed layer depth and not the mixing layer depth is more appropriate. According to Menkes et al. (2006), whether one chooses to analyse the mixed layer or the mixing layer the sum of entrainment and vertical mixing remains identical. Figure 5.2 shows how the average depth of the surface mixed layer, over the Atl3 region, changes seasonally. The surface mixed layer is at its shallowest in March/April and deepest in November.

The temperature of the subsurface water that is mixed into the surface mixed layer is however largely determined by vertical advection, as the second surface layer heat budget serves to illustrate (Figure 5.1c). Seasonal fluctuations in the mean temperature of this layer mirror seasonal SST changes (Figure 5.1a) particularly between April and October, when large changes in vertical advection occur, and to a lesser extent from November to March when surface heating quickly warms the surface mixed layer. The dominant term forcing these seasonal changes in the mean temperature of the subsurface layer is vertical advection (Figure

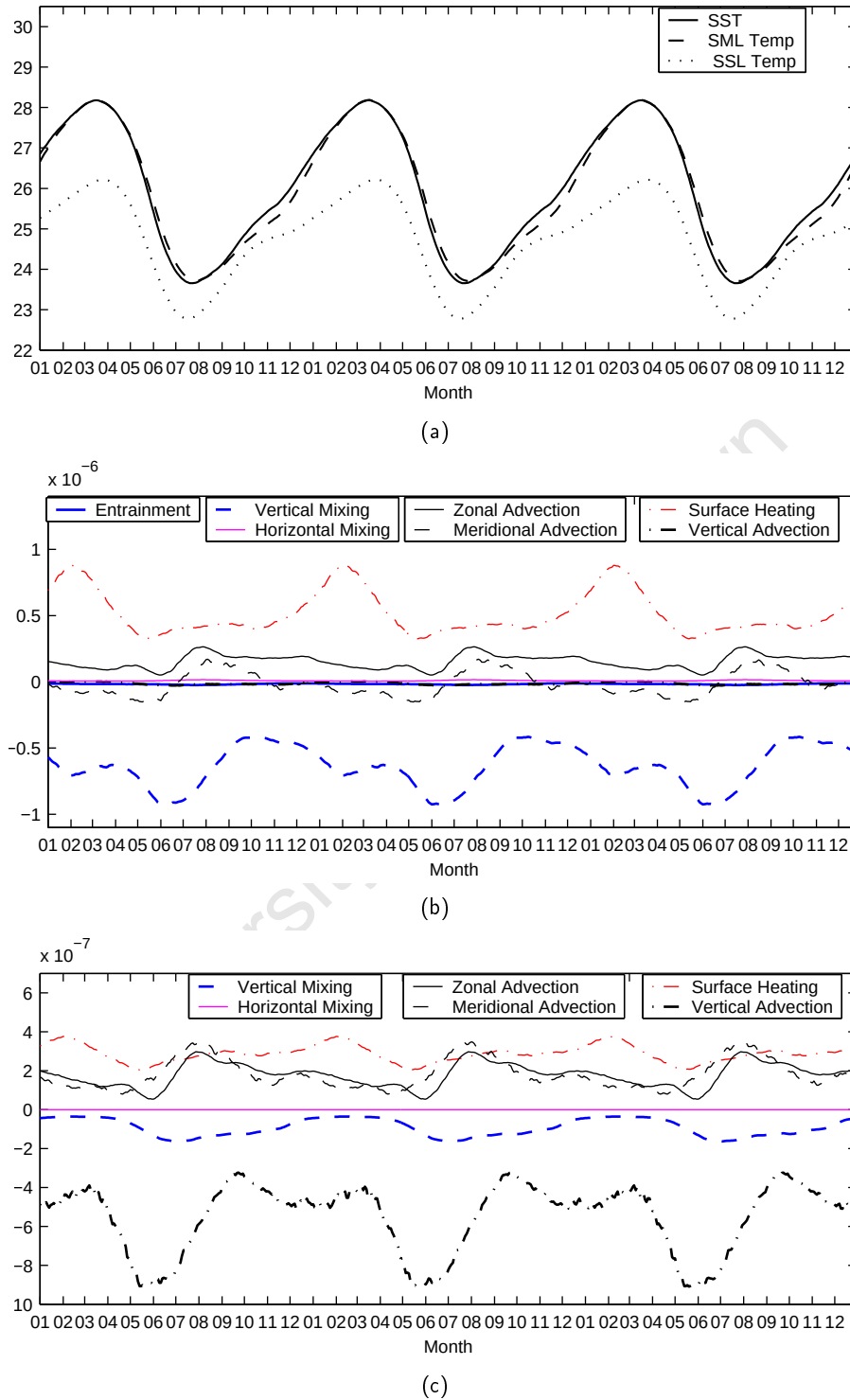


Figure 5.1: (a) Seasonal variations in SST, the average temperature of the Surface Mixed Layer (SML) (varying depth), and the average temperature of the SubSurface Layer (SSL) (fixed depth = 70m) for the Atl3 region. Temperature values are given in °C. (b-c) Climatology of terms (°C s⁻¹) contributing to seasonal variations in the mean temperature of (b) the surface mixed layer, and (c) the subsurface layer. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

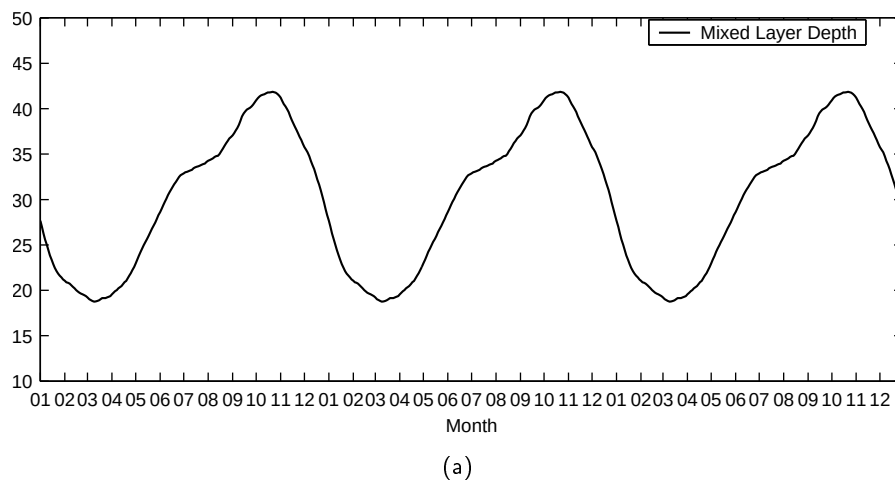


Figure 5.2: Seasonal variations in the average depth of the surface mixed layer for the Atl3 region given in meters. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

5.1c), followed by the horizontal advection terms and surface heating. Contributions from the mixing terms are minimal.

From January to March/April, the surface mixed layer warms across the entire width of the equatorial Atlantic, as the ITCZ moves towards its southernmost position, resulting in weak winds and surface currents. The surface heat flux is the dominant term responsible for the increased mixed layer heat storage during this period, as subsurface cooling and tropical instability waves are weak (Peter et al., 2006). The results obtained for the ROMS-TAtl Atl3 surface mixed layer heat budget are consistent with this seasonality (Figure 5.1b). It is during these months that the translation of subsurface temperature fluctuations/thermocline depth variations into surface mixed layer temperature and hence SST fluctuations is suppressed by weak wind stress induced vertical mixing. SST reaches an annual maximum in March/April, when APE along the equator is at its minimum, together with zonal wind stress, westerly surface currents and the zonal slope of the equatorial thermocline (Philander and Pacanowski, 1986a).

In May, the ITCZ starts to migrate northwards and the south east trades intensify near the equator. This seasonal intensification of the trades coincides with the sharp annual decline of central-eastern basin SST and the appearance of the equatorial cold tongue as the intensity of subsurface cooling increases (Weingartner and Weisberg, 1991; Carton and Zhou, 1997; Foltz et al., 2003; Peter et al., 2006). For both the surface mixed layer and subsurface layer seasonal heat budgets (Figures 5.1b and 5.1c), a sharp increase in the intensity of subsurface cooling is seen to be responsible for the rapid temperature change with subsurface cooling taking the form of enhanced vertical mixing for the surface mixed layer and vertical advection for the subsurface layer.

Similar to Chapter 3 that used monthly SODA and ORCA data, Figure 5.3 shows the relationship between seasonal equatorial Atlantic APE and Atl3 SST fluctuations calculated using ROMS-TAtl data. The in-phase relationship between seasonal APE and Atl3 SST fluctuations suggests that the large annual increase in the intensity of cooling due to vertical advection is not purely a locally forced phenomena, but largely due to the horizontal redistribution of warm surface waters in response to large scale changes in wind forcing. The seasonal shoaling of the thermocline in the east as APE increases is clearly visible in observed climatological

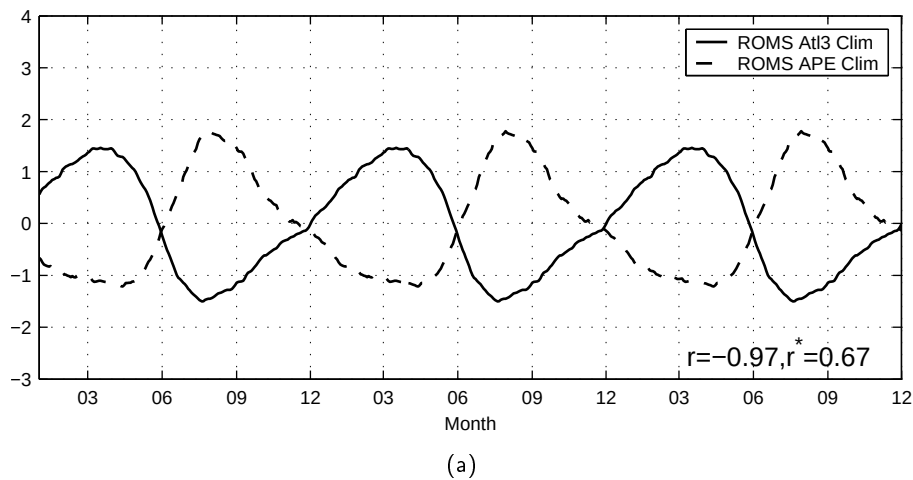


Figure 5.3: A comparison between climatological AtI3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE. SST and APE values were calculated using ROMS-TAtI data and have been normalised. The correlation between the AtI3 SST and equatorial Atlantic APE (r) is given together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). APE values were calculated following Equation 3.2.

temperature fields along the equator, as depicted in Figure 2.7. One can compare Figure 2.7, with a similar figure created for the Pacific, Figure 9 of McPhaden et al. (1998), in which the thermocline is seen to hardly vary seasonally.

An attempt is made here to diagnose the relative contribution of local Ekman pumping versus remotely forced changes in the vertical temperature structure in driving the seasonal increase in vertical advection associated with the development of the cold tongue. Based on the ROMS-TAtI wind stress forcing, seasonal variations in zonal and meridional wind stress as well as local Ekman vertical velocities for the AtI3 region are depicted in Figure 5.4. Seasonal variations in local Ekman pumping for the AtI3 region exhibit a primary dependence on seasonal variability in the zonal wind stress component. Figure 5.5 compares seasonal variations in the magnitude of the vertical advection term for the AtI3 subsurface layer with seasonal variations in Ekman pumping velocities and the rate of change of equatorial Atlantic APE (Note that vertical advection values have been inverted in Figure 5.5 to facilitate the comparison).

Clearly illustrated by Figure 5.5 is the fact that increased cooling due to vertical advection coincides with the forcing of equatorial Atlantic APE changes rather than locally enhanced vertical velocities due to Ekman pumping. This points to the importance of remotely forced changes in the horizontal distribution of warm surface waters, in the annual development of the equatorial Atlantic cold tongue. As a surface expression of this large seasonal change in the subsurface temperature structure of the eastern equatorial, AtI3 SST reaches an annual minimum during July-August as APE values in the eastern equatorial Atlantic peak (Figure 5.3). The very slight lag seen in Figure 5.5 between the peak in cooling due to vertical advection and the peak in the rate of change of equatorial Atlantic APE is thought to be due to the fact that the rate of change of equatorial Atlantic APE evaluates changes in thermocline depth across the entire width of the equatorial basin. Therefore, while remotely forced shoaling of the thermocline in the central-eastern Atlantic is responsible for the peak in cooling due to vertical advection at the base of the subsurface layer between May and June, $d\text{APE}/dt$ only reaches its maximum between June and July as the slower off-equatorial adjustment delays the deepening of the thermocline in the west.

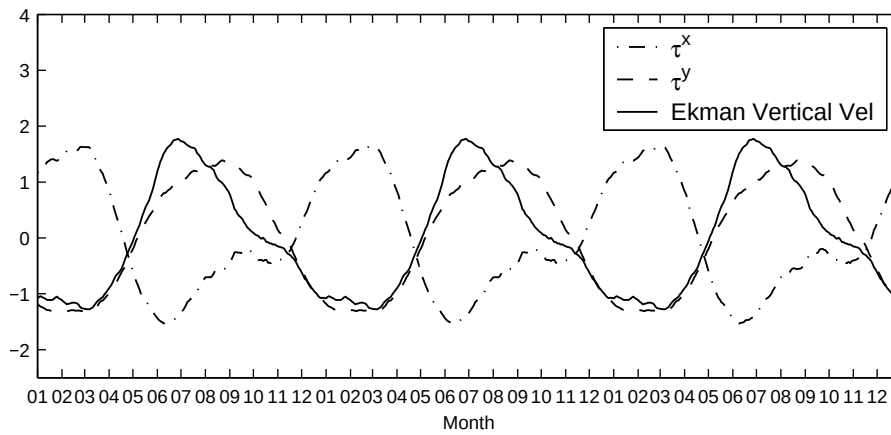
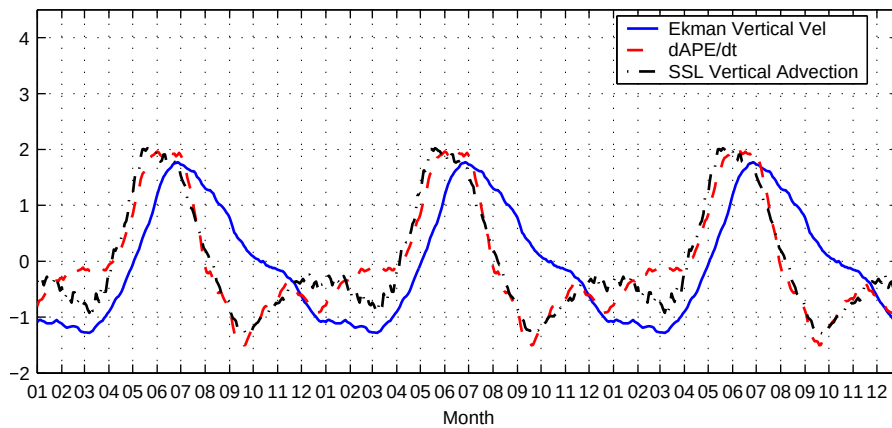


Figure 5.4: Seasonal variations in zonal and meridional ROM-TAtl wind stress forcing as well as local Ekman vertical velocities for the Atl3 region. Wind stress and Ekman vertical velocities values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.



(a)

Figure 5.5: Seasonal variations in the intensity of the vertical advection term for the Atl3 SubSurface Layer (SSL) heat budget compared with seasonal variation in Ekman vertical velocities for the Atl3 region and the rate of change of equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE. Note that, vertical advection values have been inverted to facilitate the comparison. Ekman vertical velocities, APE rate of change and vertical advection values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

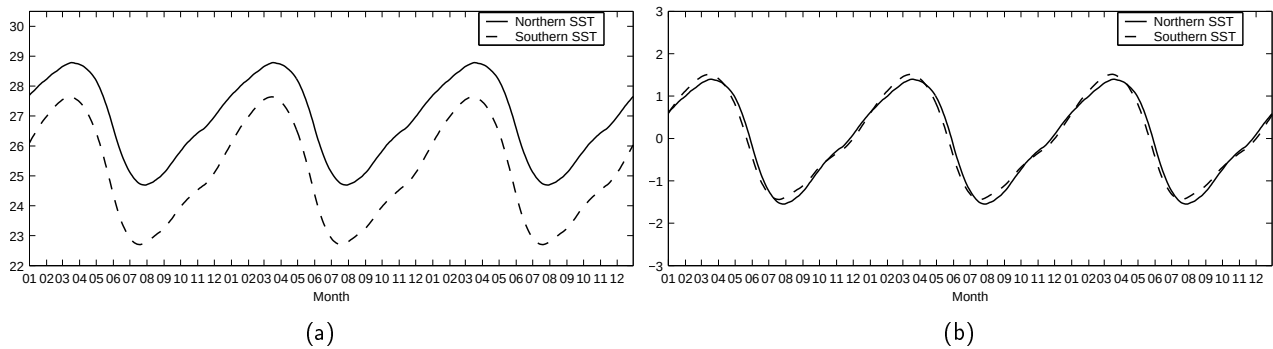


Figure 5.6: (a) Seasonal variations in SST ($^{\circ}\text{C}$) for the southern (0°N - 3°S 60°W - 15°E) and northern (3°N - 0°S 60°W - 15°E) Atl3 regions. (b) Same as (a) only SST values have been normalised for a closer comparison. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

In August the shoaling of the thermocline in the east slows and it begins to deepen during September and October (Figure 4.4). APE values decrease during this period (Figure 5.3) together with the cooling due to vertical advection as both local upwelling velocities and the vertical temperature gradient at the base of the mixed layer weaken (Figure 5.5) (Peter et al., 2006). SST in the central-eastern basin increases during September and October as decreased subsurface cooling together with increased solar heating raise the heat storage of the surface mixed layer (Figure 5.1) (Peter et al., 2006).

In the central-eastern Atlantic, the zonal component of the trades intensifies slightly in November/December abating the decrease in vertical Ekman velocities for the Atl3 region (Figure 5.4). The thermocline is also seen to shoal slightly (Figure 4.4). This shoaling of the thermocline is reflected in the APE signal (Figure 5.3) as the decrease in equatorial APE slows during these months. In the Atl3 SST index, only the rate of increase in Atl3 SST is seen to decrease during this period (Figure 5.1), as this somewhat weaker second cooling event during the months of November-December is confined to a small near-equatorial region within the central-eastern equatorial Atlantic (Okumura and Xie, 2006).

5.2.3.1 North-South Asymmetry

The Atl3 region represents a large domain and its symmetrical latitudinal extent each side of the equator isolates east-west effects, masking the asymmetries that result from the significant meridional component in the local wind stress of the central-eastern Atlantic. These equatorial asymmetries can be shown by dividing the Atl3 region into a southern Atl3 region (0°N - 3°S 60°W - 15°E) and northern Atl3 region (3°N - 0°S 60°W - 15°E). Figure 5.6 compares the seasonal cycle in SST for the southern and northern Atl3 regions. Atl3 SST north of the equator is significantly warmer (Figure 5.6a) and seasonal changes south of the equator are seen to lead slightly those north of the equator (Figure 5.6b). To explore the forcing of this asymmetry, Figures 5.1b, 5.1c and 5.4 have been repeated in Figure 5.7 for the southern and northern Atl3 regions.

Figures 5.7a and 5.7b show seasonal variations in zonal and meridional wind stress, as well as Ekman vertical velocities for the southern and northern Atl3 regions respectively. The southerly cross-equatorial component in winds over the Atl3 region drives positive Ekman vertical velocities south of the equator and negative velocities north of the equator. Although not evident in Figures 5.7a and 5.7b, as Ekman vertical velocities values have been normalised for comparison with wind stress values, it is worth noting that Ekman vertical

velocities for the southern Atl3 region are positive, while velocities for the north Atl3 region are negative, throughout the year. This distinction is thought to be the primary reason why the magnitude of cooling, due to vertical mixing at the base of the surface mixed layer, is significantly less in the northern Atl3 region than in the southern Atl3 region (Figures 5.7c and 5.7d).

The prominent annual cycle in local Ekman pumping for the southern Atl3 region exhibits a dependence on both the zonal and meridional wind stress components. These changes in local Ekman pumping are weakly aligned with seasonal variations in the intensity of subsurface cooling for the southern Atl3 region. Changes in both the magnitude of vertical mixing for the southern Atl3 surface mixed layer (Figure 5.7c) and the magnitude of vertical advection for the southern Atl3 subsurface layer (Figure 5.7e) are found to be poorly related to changes in local Ekman vertical velocities. This relationship suggests that remotely forced changes in the vertical temperature structure plays a larger role in determining seasonal changes in subsurface cooling for the southern Atl3 region. Figure 5.8a, compares seasonal variations in the magnitude of the vertical advection term for the southern Atl3 subsurface layer with seasonal variations in local Ekman pumping velocities for the southern Atl3 region and the rate of change of equatorial Atlantic APE. Here increased cooling, due to vertical advection, coincides better with the forcing of equatorial Atlantic APE changes than locally enhanced vertical velocities due to Ekman pumping. This relationship suggests that seasonal cooling of the southern Atl3 subsurface layer due to enhanced vertical advection corresponds better with remotely forced changes in the vertical temperature structure rather than locally enhanced vertical velocities due to Ekman pumping. As seen in Figure 5.5 for the entire Atl3 domain, the slight lag between the peak in cooling due to vertical advection and the peak in the rate of change of equatorial Atlantic APE seen in Figure 5.8a is thought to be due to the fact that the rate of change of equatorial Atlantic APE evaluates changes in thermocline depth across the entire equatorial basin. Therefore, remotely forced shoaling of the thermocline in the central-eastern Atlantic is still thought to be responsible for the peak in cooling due to vertical advection at the base of the subsurface layer between May and June. However, $dAPE/dt$ only reaches its maximum between June and July as the slower off-equatorial adjustment delays the deepening of the thermocline in the west.

Despite very similar seasonal cycles in the rate of change of surface mixed layer temperature for the northern and southern Atl3 regions, the dominant terms forcing these temperature rates are somewhat different (Figures 5.7c and 5.7d). In the south, the signature of seasonal fluctuation is attributable mainly to fluctuations in vertical mixing and surface heating (Figure 5.7c). In the northern Atl3 region, the horizontal advection terms are comparable in magnitude to the vertical mixing and surface heating terms, becoming an important element in driving seasonal surface mixed layer temperature changes (Figure 5.7d).

The seasonal signal in local Ekman pumping fluctuations for the northern Atl3 region is quite different from that of southern Atl3 region. Increased southerly wind stress from May to November only translates into enhanced negative Ekman vertical velocities during September and October when the easterly wind stress component decreases. A stronger relationship appears to exist between changes in local Ekman pumping and seasonal variations in the intensity of subsurface cooling in the northern Atl3 region (Figure 5.8b). It is suggested that this is because the influence of remotely forced thermocline depth variations is suppressed by the fact that Ekman vertical velocities are negative north of the equator.

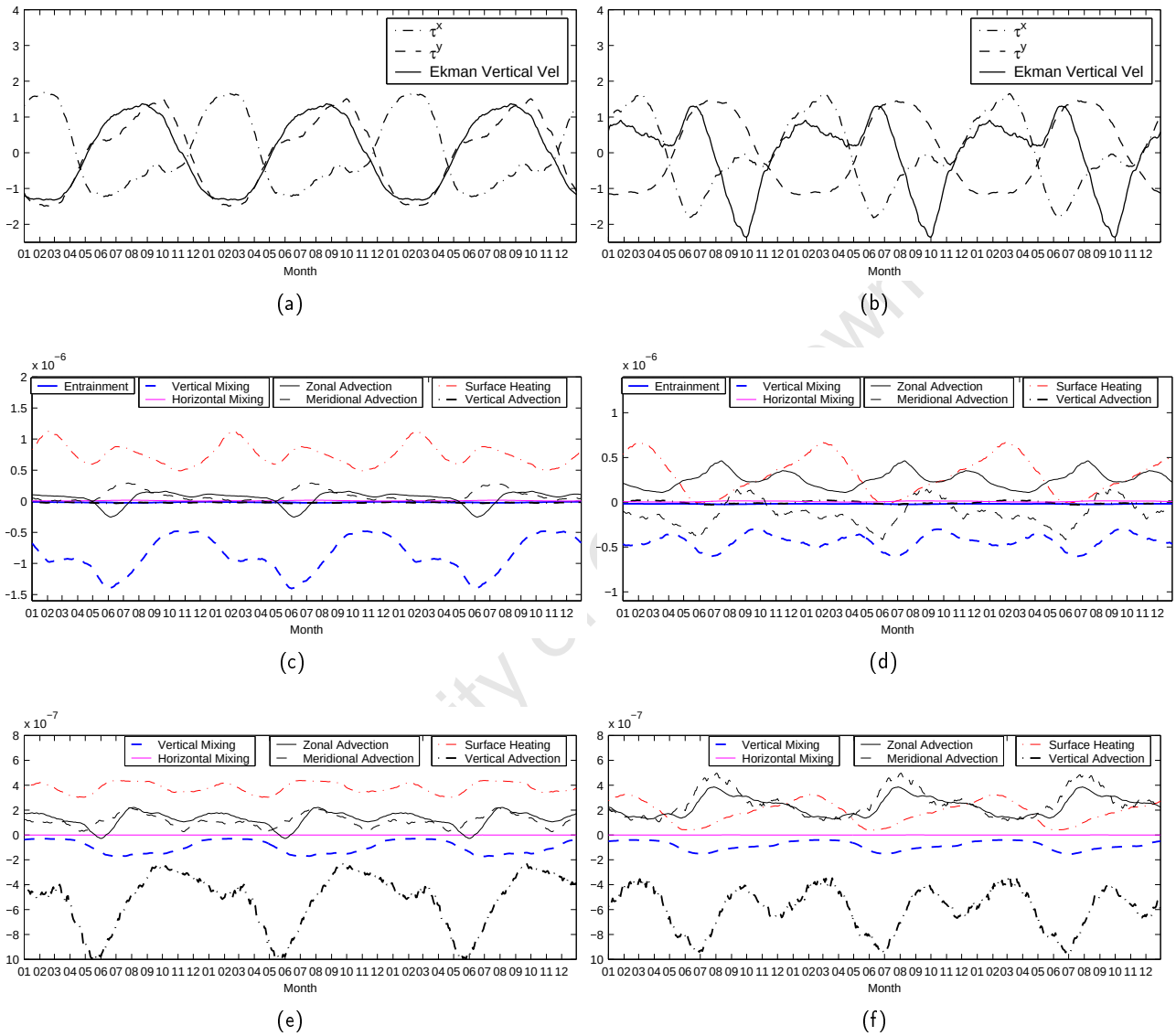


Figure 5.7: (a-b) Seasonal variations in zonal and meridional wind stress as well as local Ekman vertical velocities for (a) the southern Atl3 region, and (b) the northern Atl3 region. Wind stress and Ekman vertical velocities values have been normalised. (c-f) Climatology of terms ($^{\circ}\text{C s}^{-1}$) contributing to seasonal variations in the mean temperature of, (c) the southern Atl3 region surface mixed layer, (d) the northern Atl3 region surface mixed layer, (e) the southern Atl3 region subsurface layer, and (f) the northern Atl3 region subsurface layer. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

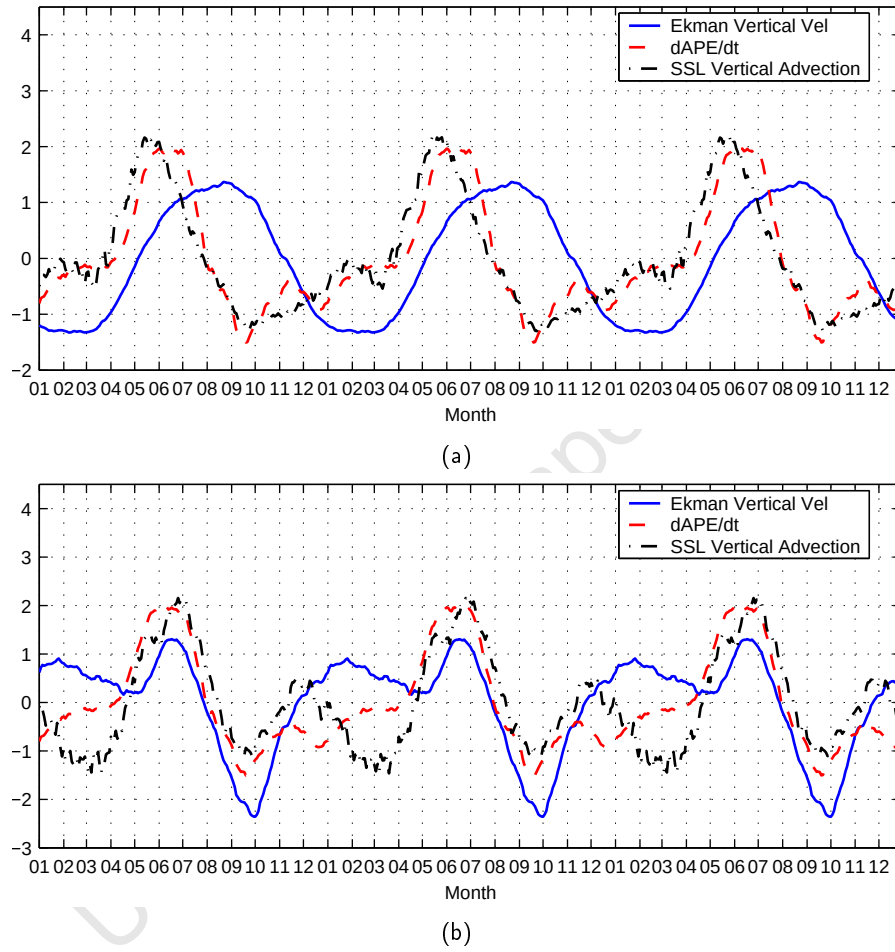


Figure 5.8: Seasonal variations in the intensity of the vertical advection term for (a) the southern Atl3 region and (b) the northern Atl3 region, subsurface layer heat budget, compared with seasonal variation in Ekman vertical velocities for (a) the southern Atl3 region and (b) the northern Atl3 region, and the rate of change of equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE. Note that, vertical advection values have been inverted to facilitate the comparison. Ekman vertical velocities, APE rate of change and vertical advection values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

5.2.4 Conclusions

Like in the Pacific, the dynamic process to which the development of the equatorial cold tongue is attributed is anomalous vertical advection due to anomalous Ekman pumping and its associated Ekman feedback mechanism (Mitchell and Wallace, 1992; Chang and Philander, 1994). However the in-phase relationship between seasonal APE and Atl3 SST fluctuations (Figure 5.3) suggests that the large annual increase in the intensity of cooling due to vertical advection is not purely a locally forced phenomena, but is largely due to the horizontal redistribution of warm surface waters in response to large scale changes in wind forcing.

The results revealed in this section show that seasonal changes in the vertical advection of cooler waters into the subsurface layer are predominantly related to remotely forced thermocline depth changes associated with the large scale redistribution of warm surface waters. These seasonal changes in the vertical advection of cooler subsurface waters into the subsurface layer are then translated into seasonal SST changes as they modulate the intensity of vertical mixing at the base of the surface mixed layer.

While the seasonality in local Ekman pumping over the Atl3 region as a whole is seen to play a secondary role to remote forcing, spatial variability in the magnitude/sign of local Ekman induced velocities is seen to be responsible for the North-South asymmetry in cooling associated with the development of the cold tongue, as a strong southerly cross-equatorial component in winds over the Atl3 region drives positive upwelling Ekman vertical velocities south of the equator and negative downwelling velocities north of the equator.

5.3 Seasonal Changes in Equatorial and Tropical Atlantic APE

5.3.1 Introduction

As established in the previous section, remotely forced thermocline depth variability plays a key role in the seasonal cycle of Atl3 SST. This section investigates the mechanism responsible for these seasonal thermocline depth changes by investigating the processes dictating seasonal, equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) and tropical Atlantic (8°S - 8°N 60°W - 15°E 0-400m) APE changes.

In the Pacific, energetics analysis of inter-annual variability has elucidated the mechanisms through which ocean-atmosphere interactions give rise to ENSO (Goddard and Philander, 2000; Fedorov et al., 2003; Fedorov, 2007). By examining the processes by which the tropical Pacific ocean gains and redistributes energy, Goddard and Philander (2000) were able to test the delayed oscillator theory (Battisti, 1988; Suarez and Schopf, 1988; Battisti and Hirst, 1989). Here, an investigation into the mechanisms driving seasonal APE changes provides us with insight into the role of the ocean within equatorial Atlantic ocean-atmosphere interactions occurring at seasonal time scales.

The primary physical processes determining the seasonal rate of change of APE can be quantitatively established by evaluating the various terms in the APE evolution equation. The following section details the methodology employed in this section as well as in Chapter 6 to evaluate the processes determining the APE evolution within a chosen domain. Thereafter, the results of this APE evolution analysis are discussed in Section 5.3.3 and conclusions summarised in Section 5.3.4.

5.3.2 Methodology - Evaluating the APE Evolution Equation

This section provides a detailed outline of the methodology employed in not only in this chapter but also in Chapter 6 when evaluating the processes determining the APE evolution within a chosen domain. While the approach detailed below draws on a number of previously detailed techniques, several aspects are novel as adaptations have been made to deal with evaluating the APE evolution of an open boundary equatorial domain, the BPE of which changes with time.

Previous studies which have dealt with diagnosing the processes determining the APE evolution of an open boundary equatorial domain have assumed global conservation of mass and employ a local definition of APE defined relative to a constant background rest state (Equation 3.3). However in this study, for reasons outlined in Section 3.2, global conservation of mass is not assumed and APE has been defined relative to a background rest state (BPE) which changes in time (Equation 3.2). An evolution equation for this global APE definition has been previously derived and employed (Winters et al., 1995; Huang, 1998). However these studies have focused on somewhat different physical processes and have dealt with closed domains. Here, an APE evolution equation is derived such that terms are decomposed in a manner relevant to the focus here on equatorial ocean dynamics in an open boundary equatorial domain.

The APE evolution analysis is conducted based on output from the TAtl-ROMS simulation. Based on output variables from this simulation, the strategy used in estimating each term in the various energy evolution equations, such that a satisfactory balance between the terms is achieved, is also described.

For the chosen definition of APE, given by Equation 3.2, the evolution equation for APE is derived by taking its time derivative:

$$\begin{aligned}
\frac{dAPE}{dt} &= \frac{d(\iiint g\rho z dV - \iiint g\rho z_* dV)}{dt} \\
&= \frac{dPE}{dt} - \frac{dBPE}{dt}
\end{aligned} \tag{5.5}$$

The evolution equation of APE is therefore derived by subtracting the evolution equation for BPE from the evolution equation for PE (Winters et al., 1995; Huang, 1998).

5.3.2.1 Potential Energy Evolution Equation

The PE of a fluid volume is defined as:

$$PE = \iiint g\rho z dV \tag{5.6}$$

where ρ is the density of the fluid and z is the vertical co-ordinate. An evolution equation for PE is derived by taking the time derivative of Equation 5.6:

$$\begin{aligned}
\frac{dPE}{dt} &= \frac{d(\iiint g\rho z dV)}{dt} \\
&= \iiint \frac{\partial(g\rho z)}{\partial t} dV + \iiint g\rho z \frac{dV}{dt}
\end{aligned} \tag{5.7}$$

$\iiint g\rho z \frac{dV}{dt} = 0$ assuming the integration volume is fixed in time. Therefore:

$$\begin{aligned}
\frac{dPE}{dt} &= \iiint \frac{\partial(g\rho z)}{\partial t} dV \\
&= \iiint \left(g\rho \frac{\partial z}{\partial t} + g \frac{\partial \rho}{\partial t} z \right) dV \\
&= \iiint g \frac{\partial \rho}{\partial t} z dV
\end{aligned} \tag{5.8}$$

as $\frac{\partial z}{\partial t} = 0$.

To ensure that the density field is conserved, a linear equation of state:

$$\rho = \rho_c - \alpha T + \beta S \tag{5.9}$$

is used to approximate density values from simulated temperature and salinity fields which observe the tracer conservation equation, where ρ_c is a constant α and β are the temperature and salinity expansion coefficients respectively ($\rho_c = 1000$, $\alpha = 2.60 \text{ kgm}^{-3}$ and $\beta = 7.65 \text{ kgm}^{-3}$). Therefore in order to facilitate the APE evolution equation analysis in this chapter and Chapter 6, a linear equation of state has been used when evaluating APE based on ROMS-TAtl temperature and salinity fields. This use of a linear equation of state differs from Chapter 3 where a non-linear equation of state was used to evaluate APE based on

SODA and ORCA temperature and salinity fields. However, this linear approximation is shown to be valid in Chapter 4 as seasonal APE fluctuations calculated based on density values derived using it are consistent with those calculated based on density values derived using a non-linear equation of state.

The density evolution equation can then be expressed in terms of the temperature and salinity evolution equations:

$$\begin{aligned}\frac{\partial \rho}{\partial t} &= \frac{\partial (\rho_c - \alpha T + \beta S)}{\partial t} \\ &= -\alpha \frac{\partial T}{\partial t} + \beta \frac{\partial S}{\partial t}\end{aligned}\quad (5.10)$$

where:

$$\frac{\partial T}{\partial t} = -\underbrace{\vec{u} \cdot \nabla T}_{A_t} + \underbrace{\nabla \cdot \gamma_t}_{D_t} + \underbrace{Q_t}_{Q_t} \quad (5.11)$$

is the temperature evolution equation. $\gamma_t = K_{th} \frac{\partial T}{\partial x} + K_{th} \frac{\partial T}{\partial y} + K_{tv} \frac{\partial T}{\partial z}$ where K_{th} represents the horizontal temperature diffusion coefficient and K_{tv} the vertical temperature diffusion coefficient. Q_t is an additional source term due the penetration of the solar radiative flux $Q_t = \frac{q_s}{\rho_0 C_p} \frac{df}{dz}$ (q_s the surface solar radiative flux and $f(z)$ the fraction of solar radiation reaching depth z).

$$\frac{\partial S}{\partial t} = -\underbrace{\vec{u} \cdot \nabla S}_{A_s} + \underbrace{\nabla \cdot \gamma_s}_{D_s} \quad (5.12)$$

is the salinity evolution equation. $\gamma_s = K_{sh} \frac{\partial S}{\partial x} + K_{sh} \frac{\partial S}{\partial y} + K_{sv} \frac{\partial S}{\partial z}$ where K_{sh} represents the horizontal salinity diffusion coefficient and K_{sv} the vertical salinity diffusion coefficient.

Substituting Equations 5.11 and 5.12 into Equation 5.10 yields:

$$\begin{aligned}\frac{d\rho}{dt} &= -\alpha (-\vec{u} \cdot \nabla T + \nabla \cdot \gamma_t + Q_t) + \beta (-\vec{u} \cdot \nabla S + \nabla \cdot \gamma_s) \\ &= -\vec{u} \cdot \nabla (-\alpha T + \beta S) - \alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t\end{aligned}\quad (5.13)$$

Equation 5.13 is then placed into 5.8 :

$$\begin{aligned}\frac{dPE}{dt} &= \iiint g (-\vec{u} \cdot \nabla (-\alpha T + \beta S) - \alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) z dV \\ &= \underbrace{-\iiint g \vec{u} \cdot \nabla (-\alpha T + \beta S) z dV}_{A_{PE}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) z dV}_{D_{PE}}\end{aligned}\quad (5.14)$$

In this form, the temporal evolution of PE within a fluid volume is driven by two terms:

A_{PE} - PE changes due to the advection of the density field.

D_{PE} - PE changes due to density diffusion as well as the penetrative solar radiation flux.

Each term in the simulated temperature and salinity evolution equations of the TAtl-ROMS simulation (Equations 5.11 and 5.12), has been saved as a two day average. Using this output, the PE balance is calculated in the form given by Equation 5.14. In Equation 5.14, term:

A_{PE} - is obtained from two day averages of term A_t in Equation 5.11 (the temperature advection term) and two day averages of term A_s in Equation 5.12 (the salt advection term) multiplied by two day averages of the depth of each sigma level.

D_{PE} - is obtained from two day averages of terms D_t and Q_t in Equation 5.11 (the temperature diffusion term and the penetrative solar radiation flux term) and two day averages of term D_s in Equation 5.12 (the salt diffusion term) multiplied by two day averages of the depth of each sigma level.

The balance of terms achieved for Equation 5.20 is depicted in Figure 5.9.

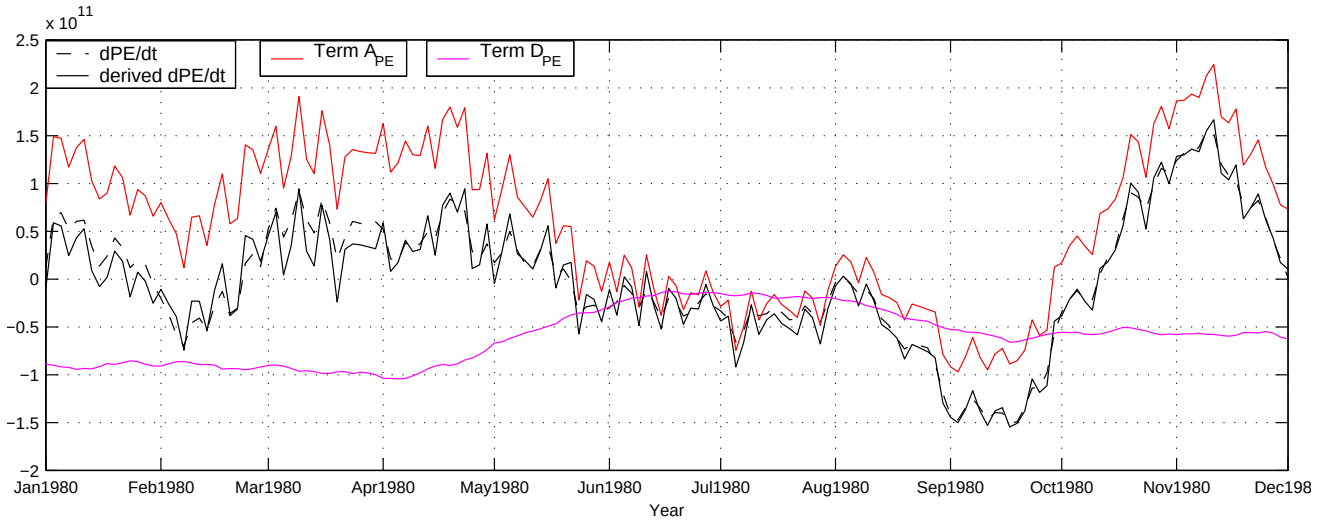


Figure 5.9: The balance of terms for the PE evolution equation in the form given by Equation 5.14, assessed over the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m). dPE/dt represents the actual rate of change of PE where as *derived* dPE/dt represents the rate of change of the PE estimated by balancing the terms within Equation 5.14. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

5.3.2.2 Background Potential Energy Evolution Equation

The BPE of a fluid volume is defined as:

$$BPE = \iiint g\rho z_* dV \quad (5.15)$$

where $z_*(\vec{x}, t)$ is the vertical position of the fluid parcel at position $(\vec{x}, t)$ with density $\rho(\vec{x}, t)$ in the reference state of minimum PE. The evolution equation for BPE is derived by taking the time derivative of Equation 5.15:

$$\begin{aligned}\frac{dBPE}{dt} &= \frac{d(\iiint g\rho z_* dV)}{dt} \\ &= \iiint \frac{\partial(g\rho z_*)}{\partial t} dV + \iiint g\rho z_* \frac{dV}{dt}\end{aligned}\quad (5.16)$$

$\iiint g\rho z_* \frac{dV}{dt} = 0$ assuming the integration volume is fixed in time. Therefore:

$$\begin{aligned}\frac{dBPE}{dt} &= \iiint \frac{\partial(g\rho z_*)}{\partial t} dV \\ &= \iiint \left(g\rho \frac{\partial(z_*)}{\partial t} + g \frac{\partial\rho}{\partial t} z_*\right) dV \\ &= \iiint g\rho \frac{\partial(z_*)}{\partial t} dV + \iiint g \frac{\partial\rho}{\partial t} z_* dV\end{aligned}\quad (5.17)$$

As shown in the appendix of Huang (1998) $\iiint g\rho \frac{\partial(z_*)}{\partial t} dV$ is equal to zero. Therefore:

$$\frac{dBPE}{dt} = \iiint g \frac{\partial\rho}{\partial t} z_* dV \quad (5.18)$$

Substituting Equation 5.13 into 5.18:

$$\begin{aligned}\frac{dBPE}{dt} &= \iiint g (-\vec{u} \cdot \nabla (-\alpha T + \beta S) - \alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t + \beta Q_s) z_* dV \\ &= \underbrace{- \iiint g \vec{u} \cdot \nabla (-\alpha T + \beta S) z_* dV}_{A_{BPE}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) z_* dV}_{D_{BPE}}\end{aligned}\quad (5.19)$$

In this form, the temporal evolution of BPE within a fluid volume is driven by two terms:

A_{BPE} - BPE changes due to the advection of the density field.

D_{BPE} - BPE changes due to density diffusion as well as the penetrative solar radiation flux.

Using the temperature and salinity evolution equation terms from the TATl-ROMS simulation, the BPE balance is calculated in the form given by Equation 5.19. In Equation 5.19, term:

A_{BPE} - is obtained from two day averages of term A_t in Equation 5.11 (the temperature advection term) and two day averages of term A_s in Equation 5.12 (the salt advection term) multiplied by z_* values.

D_{BPE} - is obtained from two day averages of terms D_t and Q_t in Equation 5.11 (the temperature diffusion term) and two day averages of term D_s in Equation 5.12 (the salt diffusion term) multiplied by z_* values.

The balance of terms achieved for Equation 5.19 is depicted in Figure 5.10.

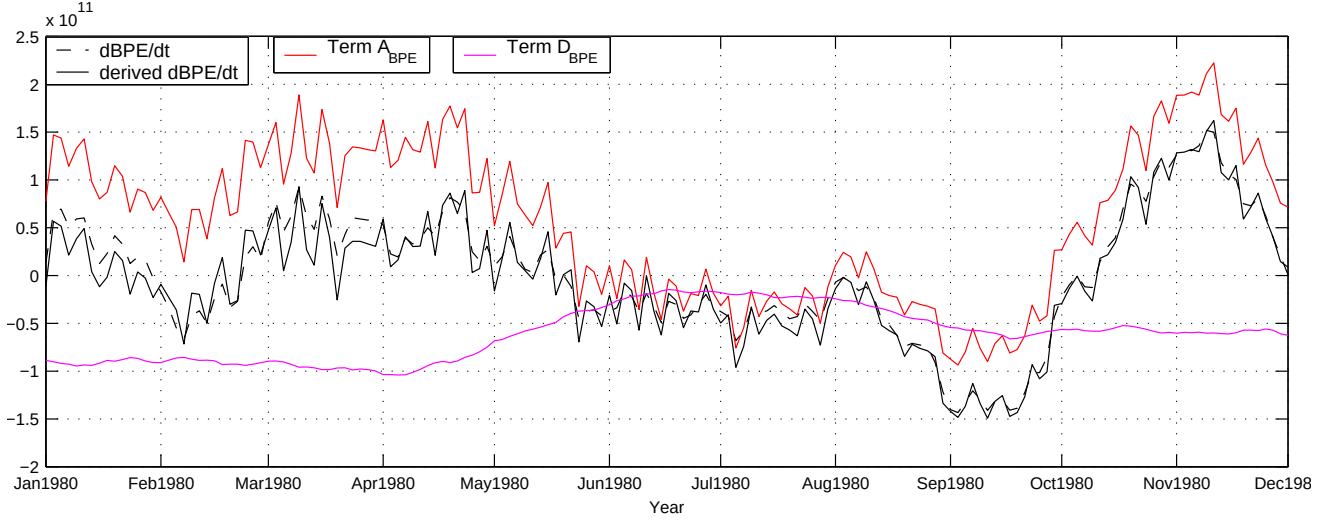


Figure 5.10: The balance of terms for the BPE evolution equation in the form given by Equation 5.19, assessed over the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m). $dBPE/dt$ represents the actual rate of change of BPE where as *derived* $dBPE/dt$ represents the rate of change of the BPE estimated by balancing the terms within Equation 5.19. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

5.3.2.3 Available Potential Energy Evolution Equation

Substituting Equations 5.14 and 5.19 into 5.5:

$$\begin{aligned}
 \frac{dAPE}{dt} &= \left(\underbrace{- \iiint g \vec{u} \cdot \nabla (-\alpha T + \beta S) z dV}_{A_{PE}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) z dV}_{D_{PE}} \right) \\
 &\quad - \left(\underbrace{- \iiint g \vec{u} \cdot \nabla (-\alpha T + \beta S) z_* dV}_{A_{BPE}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) z_* dV}_{D_{BPE}} \right) \\
 &= \underbrace{- \iiint g \vec{u} \cdot \nabla (-\alpha T + \beta S) (z - z_*) dV}_{A_{APE}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) (z - z_*) dV}_{D_{APE}} \quad (5.20)
 \end{aligned}$$

In this form, the temporal evolution of APE within a fluid volume is driven by two terms:

A_{APE} - APE changes due to the advection of the density field.

D_{APE} - APE changes due to density diffusion as well as the penetrative solar radiation flux.

In Equation 5.20, term:

A_{APE} - is obtained by subtracting term A_{BPE} of Equation 5.19 from term A_{PE} of Equation 5.14.

D_{APE} - is obtained by subtracting term D_{BPE} of Equation 5.19 from term D_{PE} of Equation 5.14.

The balance of terms achieved for Equation 5.20 is depicted in Figure 5.11.

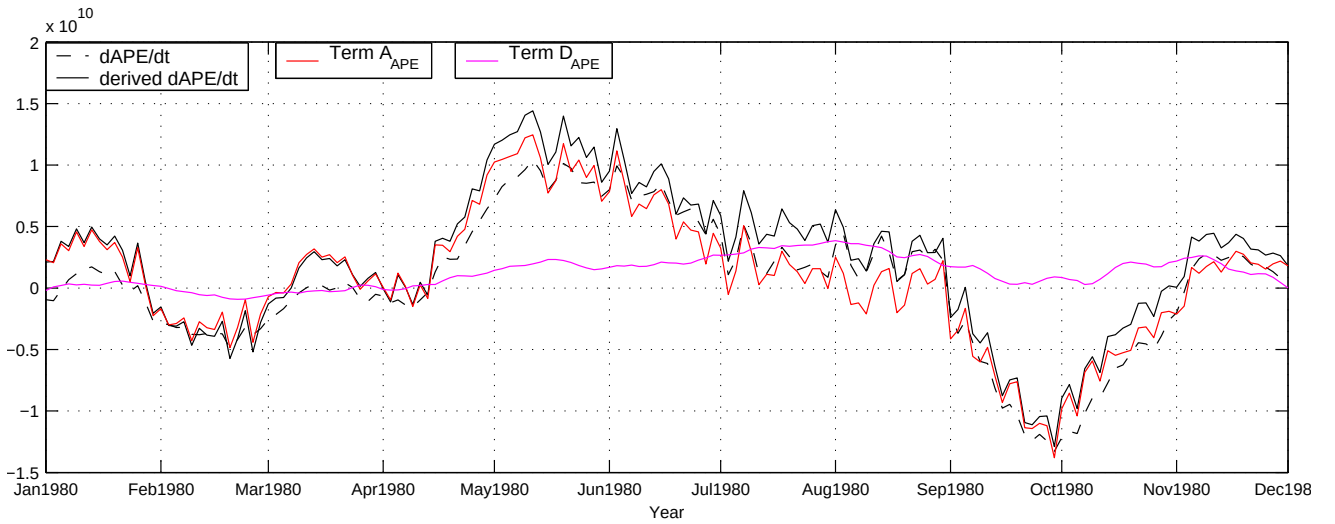


Figure 5.11: The balance of terms for the APE evolution equation in the form given by Equation 5.20, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m). $dAPE/dt$ represents the actual rate of change of APE where as *derived* $dAPE/dt$ represents the rate of change of the APE estimated by balancing the terms within Equation 5.20. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

However the APE evolution equation in the form given by Equation 5.20 yields little information as to the physical processes responsible for the density advection and diffusion dictating APE changes. Terms A_{APE} and D_{APE} in Equation 5.20 can be deconstructed further.

As in Winters et al. (1995) and Huang (1998), the term D_{APE} can be decomposed into:

Φ_{me} changes in PE due to mixing within the domain as internal energy is converted into PE.

Φ_{mr} the rate of change of BPE due to changes in the density field within the chosen volume driven by diapycnal mixing.

Φ_s changes in APE due to the surface buoyancy flux.

However, considering that when dealing with an open-boundary equatorial domain D_{APE} is seen to be a small term in comparison to A_{APE} (Figure 5.11), D_{APE} is left in this form and referred to as Φ_{apd} which represents the combined effects of the above mentioned processes on the evolution of APE.

Term A_{APE} can be decomposed into:

Φ_{apk} changes in APE due to buoyancy power, an exchange term with the KE evolution equation.

Φ_{apa} changes in APE due to the advection of APE through the surfaces of the domain.

as follows:

$$A_{APE} = \underbrace{\iiint g\vec{u} \cdot \nabla(-\alpha T + \beta S)z dV}_{A_{PE}} - \underbrace{\iiint g\vec{u} \cdot \nabla(-\alpha T + \beta S)z_* dV}_{A_{BPE}} \quad (5.21)$$

Considering $\nabla\rho_c = 0$ and Equation 5.9, $\nabla(-\alpha T + \beta S) = \nabla(\rho_c - \alpha T + \beta S) = \nabla\rho$. Equation 5.21 may therefore be written as:

$$A_{APE} = \underbrace{\iiint g\vec{u} \cdot \nabla\rho z dV}_{A_{PE}} - \underbrace{\iiint g\vec{u} \cdot \nabla\rho z_* dV}_{A_{BPE}} \quad (5.22)$$

As $z_*(\vec{x}, t)$ is by definition constant on an isopycnal surface, it can be regarded as a function of density $z_*(\rho)$. Therefore, considering that spatial variations in z_* are determined by ρ and using the Leibniz Rule, $\nabla\rho z_* = \nabla \int^\rho z_*(\rho') d\rho'$ (Winters et al., 1995), together with the continuity equation $\nabla \cdot \vec{u} = 0$ and the divergence theorem, term A_{BPE} can be rewritten as:

$$A_{BPE} = \oint g \left(\int^\rho z_*(\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS \quad (5.23)$$

To extract from A_{PE} the exchange term with the Kinetic Energy (KE) evolution equation $\Phi_{apk} = \iiint g\tilde{\rho}w dV$ (the buoyancy term), density is decomposed into, $\rho^*(z, t)$ (the vertical hydrostatically balanced density profile associated with the minimum PE rest state/BPE), and $\tilde{\rho}(\vec{x}, t)$ (perturbations from this reference state), hence $\rho = \rho^*(z, t) + \tilde{\rho}(\vec{x}, t)$. $\rho^*(z, t) = \rho(z^*, t)$ is the density profile of the background rest state in which all density surfaces are level. Substituting $\rho = \rho^*(z, t) + \tilde{\rho}(\vec{x}, t)$ into A_{PE} :

$$\begin{aligned} A_{PE} &= \iiint g\vec{u} \cdot \nabla(\rho^* + \tilde{\rho})z dV \\ &= \iiint g\vec{u} \cdot \nabla\rho^*z dV + \iiint g\vec{u} \cdot \nabla\tilde{\rho}z dV \end{aligned} \quad (5.24)$$

As mentioned above, $z_*(\vec{x}, t)$ is by definition a function of density $z_*(\rho)$, and when we consider z_* as a function of the density field associated with the background rest state, ρ^* , in which all density surfaces are level, $z_*(\rho^*) = z$. Substituting this into the first term of Equation 5.24, $(\iiint g\vec{u} \cdot \nabla\rho^*z dV = \iiint g\vec{u} \cdot \nabla\rho^*z_*(\rho^*)dV)$ and using the Leibniz Rule $(\nabla\rho^*z_*(\rho^*) = \nabla \int^{\rho^*} z_*(\rho') d\rho')$:

$$\iiint g\vec{u} \cdot \nabla\rho^*z dV = \iiint g\vec{u} \cdot \left(\nabla \int^{\rho^*} z_*(\rho') d\rho' \right) dV \quad (5.25)$$

Then invoking the continuity equation $\nabla \cdot \vec{u} = 0$ as well as the divergence theorem:

$$\begin{aligned} \iiint g \vec{u} \cdot \left(\nabla \int_{z_*}^{\rho^*} (\rho') d\rho' \right) dV &= \iiint \nabla \cdot \left(g \vec{u} \int_{z_*}^{\rho^*} (\rho') d\rho' \right) dV \\ &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS \end{aligned} \quad (5.26)$$

Substituting this modified term into Equation 5.24:

$$\begin{aligned} A_{PE} &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \iiint g \vec{u} \cdot \nabla \tilde{\rho} z dV \\ &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \iiint \nabla \cdot (g \tilde{\rho} z \vec{u}) dV - \iiint g \tilde{\rho} \nabla z \cdot \vec{u} dV \\ &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \iiint \nabla \cdot (g(\rho - \rho^*) z \vec{u}) dV - \iiint g \tilde{\rho} w dV \\ &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \iiint \nabla \cdot (g(\rho z - \rho^* z) \vec{u}) dV - \iiint g \tilde{\rho} w dV \\ &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \oint g \left(\int_{\rho^*}^{\rho} z d\rho' \right) \vec{u} \cdot \hat{n} dS - \iiint g \tilde{\rho} w dV \end{aligned} \quad (5.27)$$

Leaving $A_{PE} = \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \oint g \left(\int_{\rho^*}^{\rho} z d\rho' \right) \vec{u} \cdot \hat{n} dS - \iiint g \tilde{\rho} w dV$ and $A_{BPE} = \oint g \left(\int_{z_*}^{\rho} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS$ which substituted into Equation 5.22 reveal:

$$\begin{aligned} A_{APE} &= \oint g \left(\int_{z_*}^{\rho^*} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \oint g \left(\int_{\rho^*}^{\rho} z d\rho' \right) \vec{u} \cdot \hat{n} dS - \iiint g \tilde{\rho} w dV - \oint g \left(\int_{z_*}^{\rho} (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS \\ &= \oint g \left(- \int_{\rho^*}^{\rho} z_* (\rho') d\rho' \right) \vec{u} \cdot \hat{n} dS + \oint g \left(\int_{\rho^*}^{\rho} z d\rho' \right) \vec{u} \cdot \hat{n} dS - \iiint g \tilde{\rho} w dV \\ &= \oint g \left(- \int_{\rho^*}^{\rho} (z_* (\rho') - z) d\rho' \right) \vec{u} \cdot \hat{n} dS - \iiint g \tilde{\rho} w dV \end{aligned} \quad (5.28)$$

In this form A_{APE} is then substituted back into Equation 5.20 bringing us to the final form of the APE evolution equation:

$$\begin{aligned} \frac{dA_{PE}}{dt} &= \underbrace{\iiint g \tilde{\rho} w dV}_{\Phi_{apk}} - \underbrace{\oint \left(-g \int_{\rho^*}^{\rho} (z_* (\rho') - z) d\rho' \right) \vec{u} \cdot \hat{n} dS}_{\Phi_{apa}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) (z - z_*) dV}_{\Phi_{apd}} \\ &= \underbrace{\iiint g \tilde{\rho} w dV}_{\Phi_{apk}} - \underbrace{\oint A_{PE_{local}} \vec{u} \cdot \hat{n} dS}_{\Phi_{apa}} + \underbrace{\iiint g (-\alpha \nabla \cdot \gamma_t + \beta \nabla \cdot \gamma_s - \alpha Q_t) (z - z_*) dV}_{\Phi_{apd}} \end{aligned} \quad (5.29)$$

Where, $-g \int_{\rho^*}^{\rho} (z_* (\rho') - z) d\rho' = A_{PE_{local}}$, is a local definition for APE (Holliday and McIntyre, 1981;

Molemaker and McWilliams, 2010). In this form, the temporal evolution of APE within a fluid volume is driven by three terms:

- Φ_{apk} -the buoyancy power term which is a reversible exchange term with the KE evolution equation (Equation D.9).
- Φ_{apa} -the advection of local APE across the surfaces of the volume.
- Φ_{apd} -APE changes due to the combined effects of, changes in PE due to mixing within the domain as internal energy is converted into PE, the rate of change of BPE due to changes in the density field within the chosen volume driven by diapycnal mixing, and changes in APE due to the surface buoyancy flux.

Having obtained an adequate balance between terms for Equation 5.20 (Figure 5.11), these terms are then modified to obtain the APE evolution terms as they appear in the final form of the APE evolution equation given by Equation 5.29:

- Φ_{apk} - the buoyancy power term is calculated by based on two day averages of vertical velocity, w , multiplied by perturbation density values, $\tilde{\rho}$, which have been calculated based two day averages of temperature and salinity according to the linear equation of state given by Equation 5.9.
- Φ_{apa} -the advection of local APE across the surfaces of the volume is calculated as the difference between term A_{APE} in Equation 5.20 and the buoyancy power term, Φ_{apk} .
- Φ_{apd} - is the equivalent of term D_{APE} in Equation 5.20.

The balance of terms achieved for the APE evolution equation in its final form (Equation 5.29) is depicted in Figure 5.12.

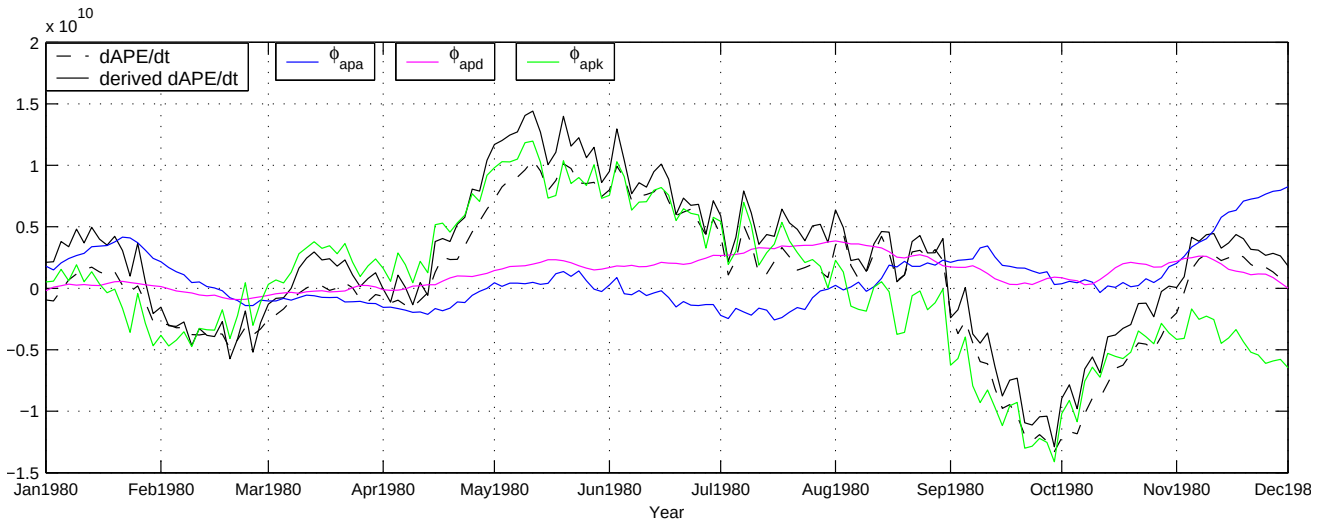


Figure 5.12: Balance of terms for the APE evolution equation in the final form given by Equation 5.29, assessed over the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0 - 400m). $dAPE/dt$ represents the actual rate of change of APE where as $derived\ dAPE/dt$ represents the rate of change of the APE estimated by balancing the terms within Equation 5.29. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

5.3.2.4 Kinetic Energy Evolution Equation

To identify the primary physical processes contributing to the buoyancy power term, Φ_{apk} , which is the reversible exchange term between APE and KE, the balance of terms in the KE evolution equation for the equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) and tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) is evaluated. The derivation of the KE evolution equation is standard and is outlined in Appendix D together with the strategy used in estimating each term in the KE evolution equation such that a satisfactory balance between the terms is achieved. The KE evolution equation in its final form:

$$\begin{aligned} \frac{dKE}{dt} = & - \underbrace{\oint \left(\frac{\rho_o}{2} \vec{v}^2 \vec{u} \right) \cdot \hat{n} dS}_{\Phi_{ka}} - \underbrace{\oint (\tilde{p} \vec{u}) \cdot \hat{n} dS}_{\Phi_{pw}} - \underbrace{\iiint \tilde{\rho} g w dV}_{\Phi_{apk}} + \underbrace{\iint_{z=0} \vec{v} \cdot \tau_s dS}_{\Phi_{ww}} \\ & - \underbrace{\iiint \rho_o \left(K_{mh} \left(\frac{\partial \vec{v}}{\partial x} \cdot \frac{\partial \vec{v}}{\partial x} \right) + K_{mh} \left(\frac{\partial \vec{v}}{\partial y} \cdot \frac{\partial \vec{v}}{\partial y} \right) + K_{mv} \left(\frac{\partial \vec{v}}{\partial z} \cdot \frac{\partial \vec{v}}{\partial z} \right) \right) dV}_{\Phi_{ss}} \end{aligned} \quad (5.30)$$

In this form, the temporal evolution of KE within a fluid volume is driven by five terms:

- Φ_{ka} -the advection of KE across the surfaces of the volume.
- Φ_{pw} -pressure work across the surfaces of the volume (Winters et al., 1995).
- Φ_{apk} -the buoyancy power term which is a reversible exchange term with the APE evolution equation (Equation 5.29).
- Φ_{ww} -work done by the wind stress acting on surface currents (Goddard and Philander, 2000).
- Φ_{ss} -work done by shear stresses of horizontal flows within the domain (Winters et al., 1995).

5.3.2.5 Evaluating Climatological APE

An approximation has been made in this section and Chapter 6 when evaluating climatological APE values. Density and the vertical position of the fluid parcel within the background state can be decomposed into their climatological component and inter-annual perturbations from this climatological component, $z_* = z_{*cl} + z'_*$ and $\rho = \rho_{cl} + \rho'$ such that:

$$\begin{aligned} APE &= \iiint g \rho (z - z_*) dV \\ &= \iiint g (\rho_{cl} + \rho') (z - (z_{*cl} + z'_*)) dV \\ &= \iiint g \rho_{cl} (z - (z_{*cl} + z'_*)) dV + \iiint g \rho' (z - (z_{*cl} + z'_*)) dV \\ &= \iiint g \rho_{cl} (z - z_{*cl}) dV - \iiint g \rho_{cl} z'_* dV + \iiint g \rho' (z - z_{*cl}) dV - \iiint g \rho' z'_* dV \end{aligned} \quad (5.31)$$

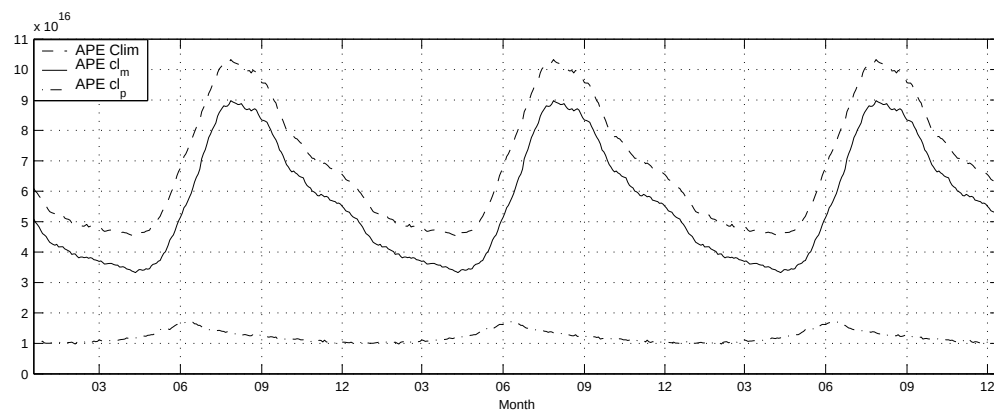
The climatological average of APE is then:

$$\begin{aligned}
APE_{cl} &= \left\{ \iiint g\rho(z - z_*)dV \right\}_{cl} \\
&= \left\{ \iiint g\rho_{cl}(z - z_{*cl})dV \right\}_{cl} - \left\{ \iiint g\rho_{cl}z'_*dV \right\}_{cl} \\
&\quad + \left\{ \iiint g\rho'(z - z_{*cl})dV \right\}_{cl} - \left\{ \iiint g\rho'z'_*dV \right\}_{cl} \\
&= \underbrace{\left\{ \iiint g\rho_{cl}(z - z_{*cl})dV \right\}_{cl}}_{cl_m} - \underbrace{\left\{ \iiint g\rho'z'_*dV \right\}_{cl}}_{cl_p}
\end{aligned} \tag{5.32}$$

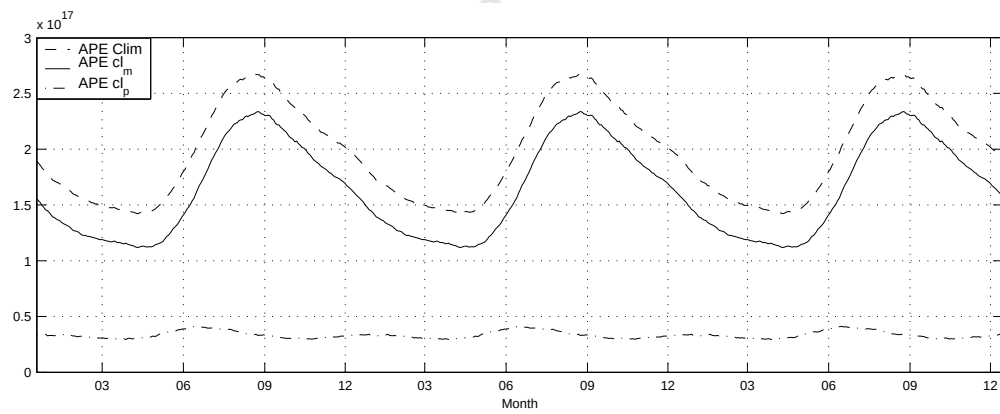
The climatological average of APE values over a given inter-annual time series, such as the climatological APE values given in Chapter 3 and Section 5.2, consist of two components - term cl_m of Equation 5.32 representing the product of climatological values in ρ and z_* and term cl_p of Equation 5.32 representing the climatological average of the product of inter-annual anomalies in ρ and z_* .

In Figure 5.13, the relative contribution of terms cl_m and cl_p of Equation 5.32 to climatological APE values is shown. The contribution of cl_p , the climatological average of the product of inter-annual anomalies in ρ and z_* to climatological APE changes is seen to be small and so the approximation of climatological APE changes by term cl_m , representing the product of climatological values in ρ and z_* is seen to be reasonable. This approximation simplifies the analysis undertaken in the following section and in Chapter 6, as the climatological APE evolution equation depends only on the climatological ρ , z_* , w , A_t , D_t , Q_t , A_s and D_s values derived from the ROMS-TAtl simulation. Likewise, the climatological KE evolution equation (outlined in Appendix D) only depends on climatological $\vec{u}$, τ , A_m , P_m and D_m values.

$$APE_{cl} \approx \iiint g\rho_{cl}(z - z_{*cl})dV \tag{5.33}$$



(a)



(b)

Figure 5.13: A comparison between climatological APE values and the two components which make up climatological APE values, cl_m and cl_p for (a) the equatorial Atlantic and (b) the tropical Atlantic.

5.3.3 Results

For the tropical Pacific (15N-15S), Goddard and Philander (2000) have shown the dominant term driving large inter-annual APE changes to be the buoyancy power term. Representing the vertical motion of the mass field, buoyancy power creates thermocline perturbations via convergence or divergence of the mass field. Source/sink terms do not play a major role in the tropical Pacific as APE changes result mainly from the adiabatic redistribution of mass within this large tropical (15N-15S) domain (Goddard and Philander, 2000).

When we consider the balance of terms associated with the climatological evolution of APE in the equatorial (3°S-3°N 60°W-15°E 0-400m) and tropical Atlantic (8°S-8°N 60°W-15°E 0-400m), we see that like the tropical Pacific the dominant term in the APE evolution equation is the buoyancy power term. Figures 5.14a and 5.14b depict the climatological balance of terms for the APE evolution equation (Equation 5.29), assessed over the equatorial Atlantic domain and the tropical Atlantic domain respectively. Clearly evident in these figures is the fact that seasonal APE changes are driven primarily by the buoyancy power term. The approximation that equatorial/tropical basin APE changes are driven primarily by buoyancy forcing associated with the horizontal redistribution of warm surface waters is therefore seen to hold when evaluating seasonal APE changes in the Atlantic having taken into account seasonal changes in BPE (Equation 5.34).

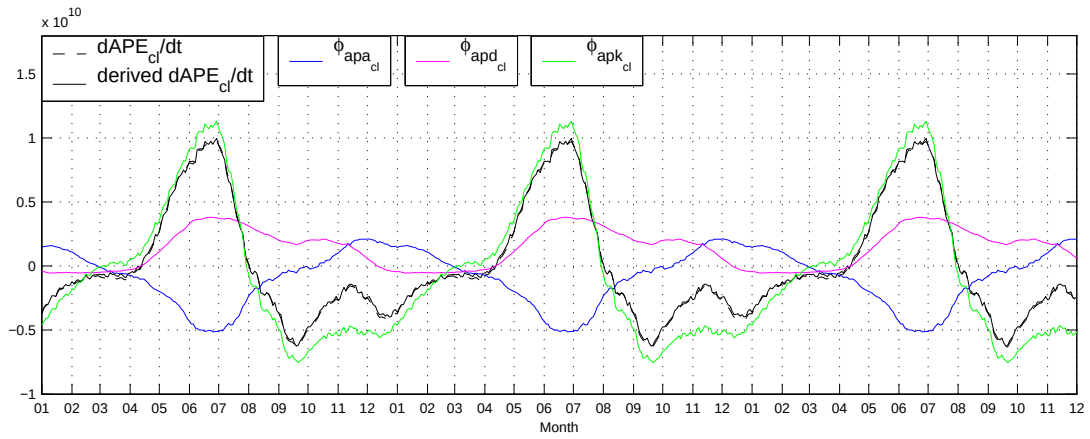
$$\frac{dAPE_{cl}}{dt} \approx \underbrace{\iiint g\tilde{\rho}_{cl}w_{cl}dV}_{\Phi_{apk_{cl}}} \quad (5.34)$$

The buoyancy power term is the reversible exchange term between the APE and KE evolution equations and so further insight can be gained into the processes responsible for buoyancy power fluctuations by considering the KE evolution equation (Equation 5.30).

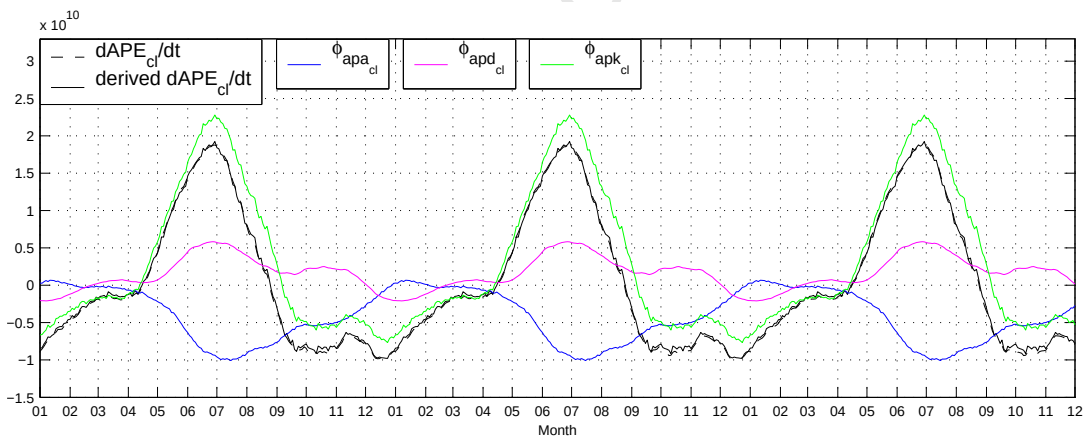
$$\begin{aligned} \frac{dKE_{cl}}{dt} &= -\Phi_{ka_{cl}} - \Phi_{pw_{cl}} - \Phi_{apk_{cl}} + \Phi_{ww_{cl}} - \Phi_{ss_{cl}} \\ \Phi_{apk_{cl}} &= -\frac{dKE_{cl}}{dt} - \Phi_{ka_{cl}} - \Phi_{pw_{cl}} + \Phi_{ww_{cl}} - \Phi_{ss_{cl}} \end{aligned} \quad (5.35)$$

In the case of the tropical Pacific, the source of buoyancy power fluctuations in the KE evolution equation is seen to be the wind power term (Goddard and Philander, 2000). While only a small portion of the work done by the wind drives changes in the KE of the tropical Pacific, a far larger portion goes toward driving changes in APE by doing work against pressure gradients and generating buoyancy power. The rate of change of APE in the tropical Pacific is seen to be approximately equal to the work done on the ocean by the wind minus dissipation and the flux of energy out of the domain (Fedorov, 2007). A quadrature relationship is seen to exist between wind power and APE in the Pacific, with inter-annual wind power fluctuations leading APE fluctuation as anomalous wind power drives buoyancy power anomalies resulting in anomalous APE evolution (Goddard and Philander, 2000).

Figure 5.15 shows the full climatological balance of terms for the KE evolution equation (Equation 5.30) assessed over the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m) respectively. In both domains, the wind power term ($\Phi_{ww_{cl}}$) is the dominant source term. As in the Pacific (Goddard and Philander, 2000; Brown and Fedorov, 2010), only a small



(a)



(b)

Figure 5.14: The climatological balance of terms (Js^{-1}) for the APE evolution equation (Equation 5.29), assessed over (a) the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0-400m), and (b) the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

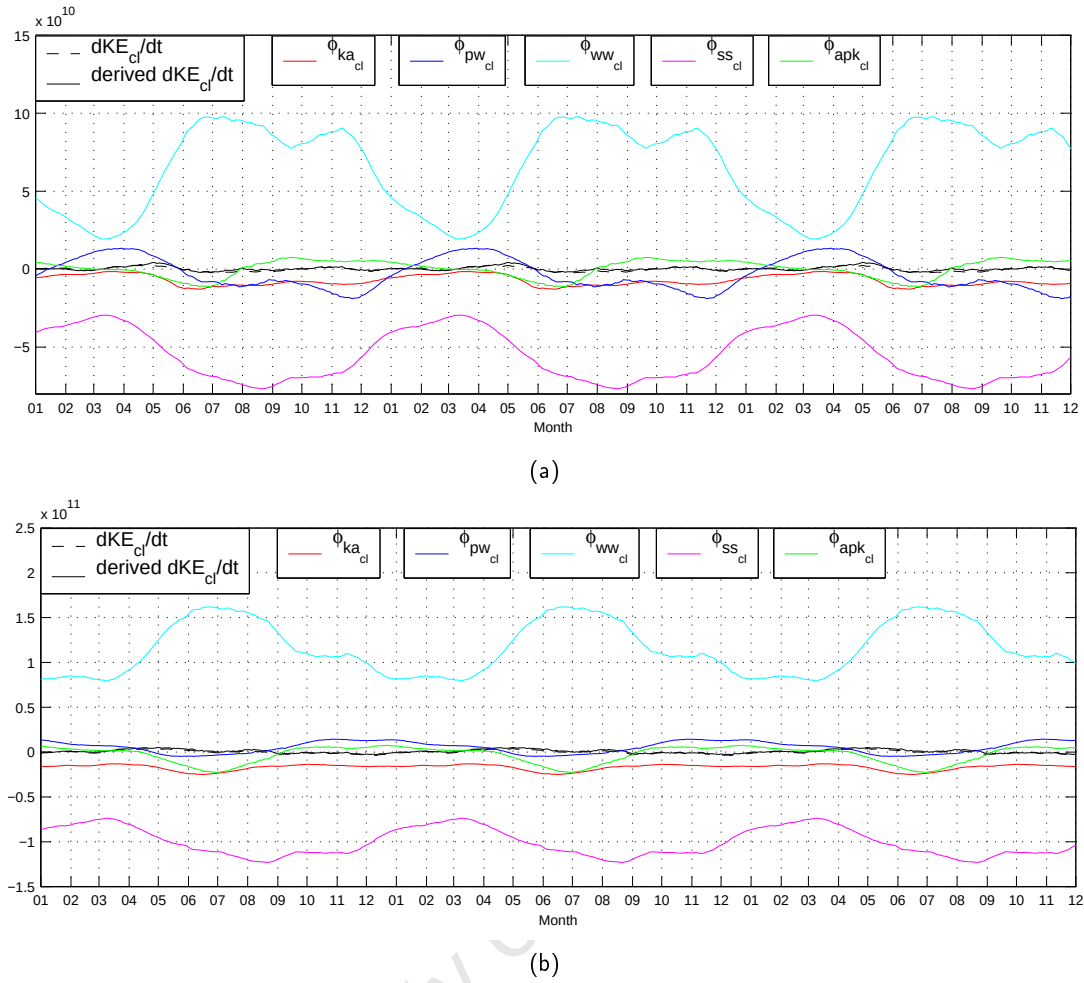


Figure 5.15: The climatological balance of terms (Js^{-1}) for the KE evolution equation (Equation 5.30), assessed over (a) the equatorial Atlantic domain (3°S - 3°N 60°W - 15°E 0 - 400m), and (b) the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0 - 400m). A 14 day running mean has been applied to the time series to smooth out the high frequency variability. Note $\Phi_{apk_{cl}}$ values in this KE evolution equation balance are the inverse of $\Phi_{apk_{cl}}$ values given for the APE evolution equation (Figure 5.14).

percentage of the work done on the ocean by the wind translates into KE changes ($d\text{KE}/dt$) and a larger portion of the wind power is dissipated. The dissipation of work done by the wind is seen in both domains to be primarily due to shear stresses of horizontal flows within each domain ($\Phi_{ss_{cl}}$) with the advection of KE out of each domain ($\Phi_{ka_{cl}}$) playing a smaller dissipative role. The pressure work term ($\Phi_{pw_{cl}}$) representing the work done against internal and surface pressure gradients by ageostrophic flow is found to dissipate the work done by the wind during certain months of the year. This term changes sign from an energy sink to an energy source term seasonally, however its seasonal fluctuations are significantly less than that of the primary source term, namely wind power and the primary dissipative term, the dissipation due to shear stresses of horizontal flows. Only a portion of wind power is converted into buoyancy power ($\Phi_{apk_{cl}}$) as the residual forces buoyancy power fluctuations.

As in the Pacific, one would expect the horizontal redistribution of warm surface waters, which gives rise to seasonal buoyancy power and hence seasonal APE changes in the Atlantic, to be primarily related to fluctuations in wind power over the domain. In Figure 5.16, seasonal fluctuations in the buoyancy power term are compared with seasonal fluctuations in the wind power term for both the equatorial and tropical

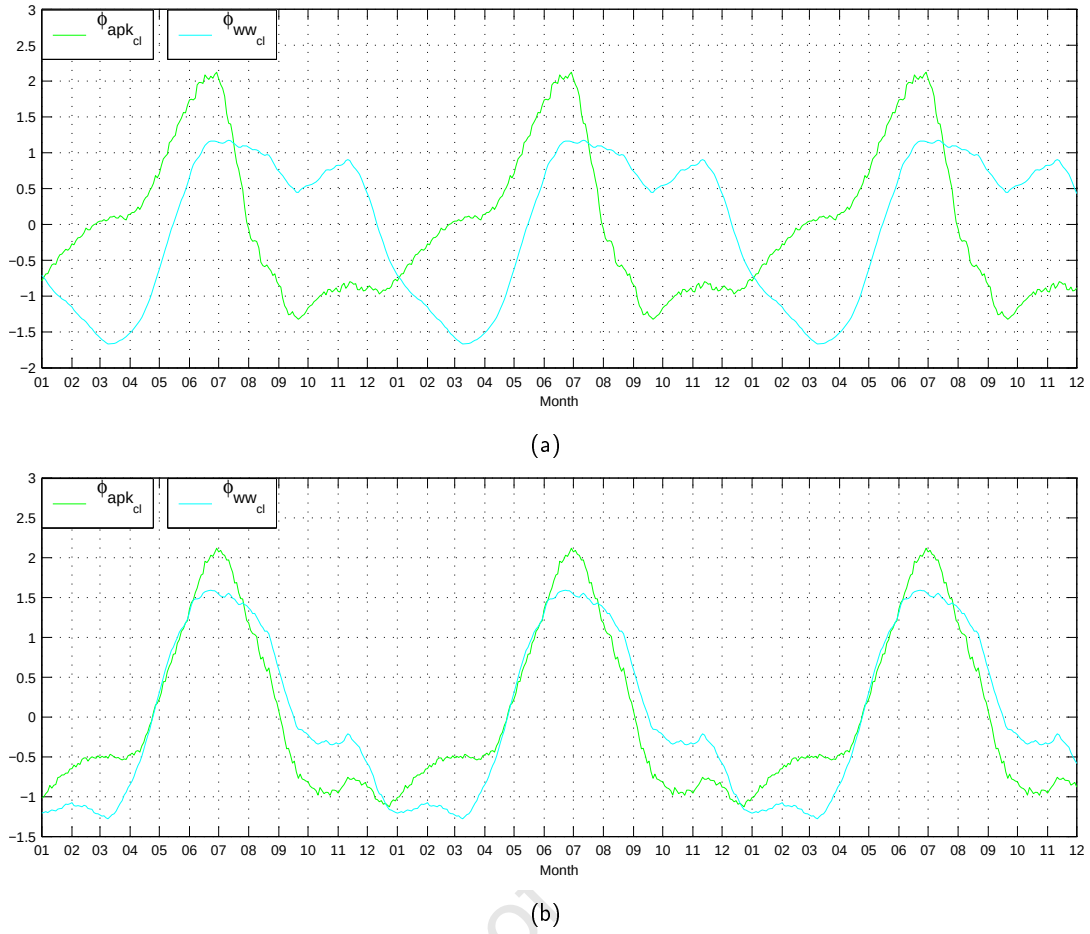
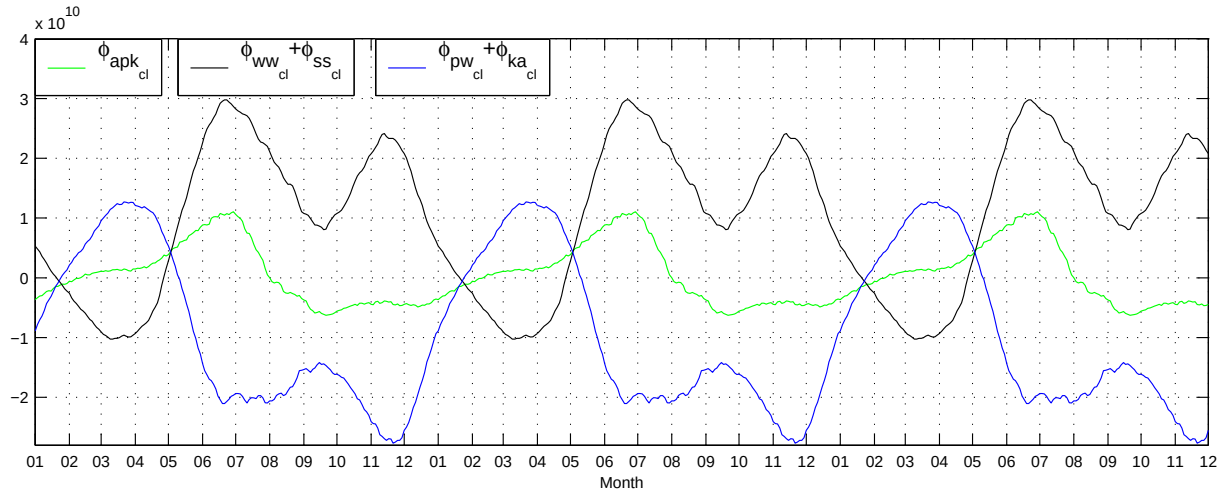


Figure 5.16: A comparison between seasonal fluctuations in buoyancy power ($\Phi_{apk_{cl}}$) and wind power ($\Phi_{ww_{cl}}$) for (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m), and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). Buoyancy power ($\Phi_{apk_{cl}}$) and wind power ($\Phi_{ww_{cl}}$) values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

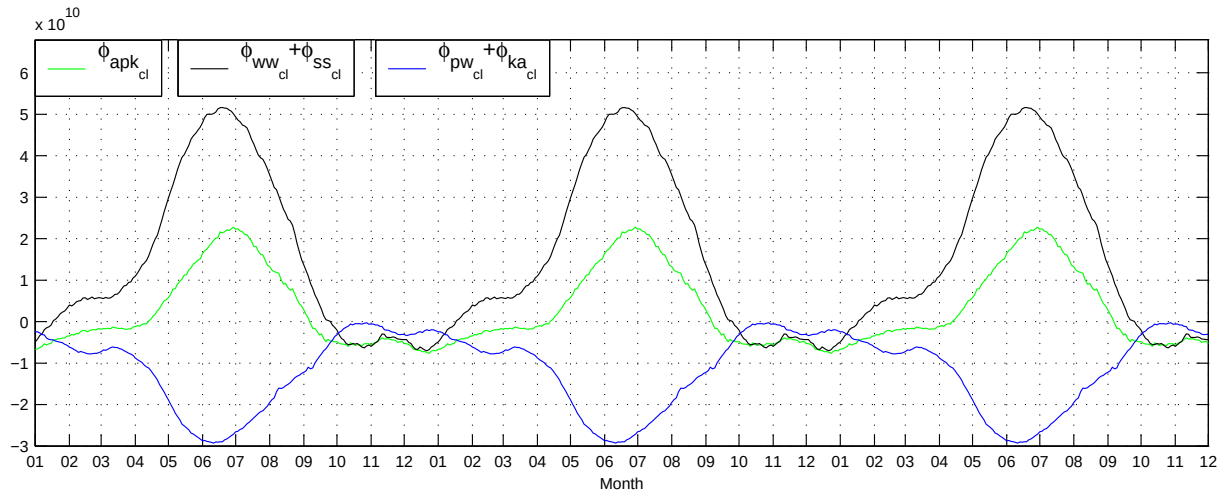
Atlantic domains. In Figure 5.16b, seasonal changes in the buoyancy power within the tropical Atlantic domain correspond closely with seasonal changes in the work done by the wind over this domain. This relationship is not observed however for the equatorial domain (Figure 5.16a), as wind power and seasonal buoyancy power changes are not totally in-phase. As shown in the following paragraphs, this difference is attributed to the larger influence of processes acting at the boundaries of the equatorial domain than in the tropical domain.

Figure 5.17 compares seasonal fluctuations in the buoyancy power term with seasonal fluctuations in the combined effect of terms associated with physical processes operating within the domain as well as at the boundaries of the domain. The combined effect of physical processes operating within the domain is represented by, $\Phi_{ww_{cl}} + \Phi_{ss_{cl}}$, the sum of work done by wind stress acting on surface currents within the domain and dissipation by shear stresses of horizontal flows within the domain (the KE evolution term (dKE/dt) is assumed negligible and neglected). The combined effect of terms associated with processes acting at the boundaries of the domain is represented by, $\Phi_{ka_{cl}} + \Phi_{pw_{cl}}$, the advection of KE across the surface of the domain and pressure work across the surfaces of the volume.

For the large tropical Atlantic domain, seasonal fluctuations in the buoyancy power term are in-phase with



(a)



(b)

Figure 5.17: A comparison between seasonal fluctuations in the buoyancy power term ($\Phi_{apk_{cl}}$); the combined effect of physical processes operating within the domain, $\Phi_{ww_{cl}} + \Phi_{ss_{cl}}$; and the combined effect of terms associated with boundary processes, $\Phi_{ka_{cl}} + \Phi_{pw_{cl}}$ for (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). Values are given in Js^{-1} . A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

seasonal fluctuations in the physical processes operating within the domain (Figure 5.17b) and do not correspond with processes acting at the boundaries, implying the lesser role of boundary processes in driving seasonal changes in the buoyancy power term within the tropical Atlantic domain. This result is consistent with Figure 5.16b in which seasonal changes in the buoyancy power within the tropical Atlantic domain correspond closely with seasonal changing in the work done by the wind over this domain. For the tropical Atlantic domain, seasonal changes in the buoyancy power term and hence APE, are seen to be primarily dictated by seasonal changes in the work done by wind stress acting on surface currents within the domain.

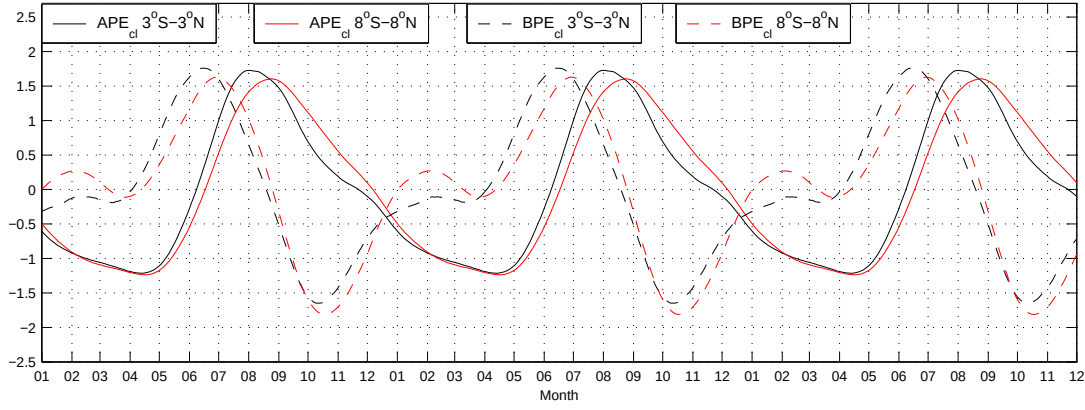
For the smaller equatorial Atlantic domain, seasonal changes in buoyancy power do not correspond closely with seasonal changes in wind power (Figure 5.16a) and seasonal buoyancy power changes do not correspond closely with physical processes operating within the domain nor processes acting at the boundaries (Figure 5.17a). Instead, buoyancy power changes are the residual of the sum of these processes, $\Phi_{ww_{cl}} + \Phi_{ss_{cl}} + \Phi_{ka_{cl}} + \Phi_{pw_{cl}}$. Both physical processes operating within the domain and processes acting at the boundaries appear to play a significant role in driving seasonal buoyancy power changes.

The larger influence of processes acting at the boundaries of the equatorial domain than the tropical domain points to the significant contribution of off-equatorial adjustment on buoyancy power within the equatorial domain, as seasonally excited transients enter and exit the equatorial domain. This finding is consistent with conclusions drawn in previous studies that boundary reflections play an important role in seasonal equatorial SSH variability (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009) and that the contribution of Kelvin and Rossby waves reflected at the eastern and western boundaries to simulated equatorial SSH variability is equal to directly forced waves (Ding et al., 2009).

Seasonal APE changes within the equatorial domain are therefore driven not only by seasonal changes in the work done by the wind within the equatorial domain but also by the response of the greater tropical Atlantic domain to seasonal fluctuations in wind forcing via its impact along the boundaries of the equatorial domain. Therefore to better understand the processes behind the seasonal redistribution of warm surface waters and associated APE changes within the equatorial Atlantic, the seasonal redistribution of warm surface waters driven by wind stress forcing over the tropical Atlantic region, and the oceanic response to this forcing, needs to be considered. The meridional extent of the tropical domain chosen, 8°S-8°N, is deemed large enough to encompass this wind forced, seasonally adjusting region as seasonal buoyancy power changes are predominantly forced by wind power fluctuations and boundary processes play a lesser role. Extending the domain over which APE is evaluated to include the tropical domain simplifies the dynamics governing seasonal APE variability down to one dominant process, namely the work done by the wind over the tropical Atlantic domain.

As shown in Figure 5.18, the extension of the meridional domain over which APE is evaluated to include APE changes due to the meridional redistribution of mass within the tropical domain, does not have a significant effect on the signature of seasonal APE variability with the exception that APE values peak approximately a month later for the tropical domain. This result is ascribed to the fact that this larger domain takes longer to adjust than the equatorial region to seasonal changes in the wind forcing.

Seasonal changes in the APE of the tropical Atlantic are seen to be primarily forced by wind power fluctuations ($\Phi_{ww_{cl}} = \iint_{z=0} \vec{v}_{cl} \cdot \tau_{s_{cl}} dS = \iint_{z=0} u_{cl} \tau_{s_{cl}}^x + v_{cl} \tau_{s_{cl}}^y dS$). Figure 5.19 compares the relative contribution of the zonal ($\Phi_{ww_{cl}}^x = \iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS$) and meridional ($\Phi_{ww_{cl}}^y = \iint_{z=0} v_{cl} \tau_{s_{cl}}^y dS$) components of the wind power term to climatological wind power values. Zonal wind stress acting on zonal surface currents is



(a)

Figure 5.18: Equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) climatological APE and BPE changes versus tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) climatological APE and BPE changes. APE and BPE values were calculated based on output from the ROMS-TatI simulation (1980-2004) following Equation 3.2. Values have been normalised to facilitate a better comparison. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

clearly responsible for the majority of work done on the ocean by the wind within the tropical Atlantic $\Phi_{ww_{cl}} \approx \Phi_{ww_{cl}^x}$.

Making the assumption that seasonal wind power fluctuation may be approximated by the work done by zonal winds acting on zonal surface currents, $\Phi_{ww_{cl}} \approx \Phi_{ww_{cl}^x}$, this term is decomposed into its mean, $\Phi_{ww_{cl}^x}^{mm} = \iint_{z=0} \overline{u_{cl}} \overline{\tau_{s_{cl}}^x} dS$, mean perturbation, $\Phi_{ww_{cl}^x}^{mp} = \iint_{z=0} u'_{cl} \overline{\tau_{s_{cl}}^x} dS + \iint_{z=0} \overline{u_{cl}} \tau_{s_{cl}}^{x'} dS$, and perturbation, $\Phi_{ww_{cl}^x}^{pp} = \iint_{z=0} u'_{cl} \tau_{s_{cl}}^{x'} dS$ components.

$$\begin{aligned}
 \Phi_{ww_{cl}^x} &= \iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS \\
 &= \iint_{z=0} (\overline{u_{cl}} + u'_{cl}) (\overline{\tau_{s_{cl}}^x} + \tau_{s_{cl}}^{x'}) dS \\
 &= \iint_{z=0} \overline{u_{cl}} \overline{\tau_{s_{cl}}^x} dS + \iint_{z=0} u'_{cl} \overline{\tau_{s_{cl}}^x} dS + \iint_{z=0} \overline{u_{cl}} \tau_{s_{cl}}^{x'} dS + \iint_{z=0} u'_{cl} \tau_{s_{cl}}^{x'} dS \quad (5.36)
 \end{aligned}$$

This decomposition is similar to that done by Goddard and Philander (2000) except that they evaluated the effect of inter-annual zonal wind stress and zonal current perturbations from climatological values.

$$\Phi_{ww_{cl}^x} = \Phi_{ww_{cl}^x}^{mm} + \Phi_{ww_{cl}^x}^{mup\tau} + \Phi_{ww_{cl}^x}^{m\tau pu} + \Phi_{ww_{cl}^x}^{pp} \quad (5.37)$$

In this context, $\overline{\tau_{s_{cl}}^x}$ and $\overline{u_{cl}}$ represent mean zonal wind stress and surface current values, while $\tau_{s_{cl}}^{x'}$ and u'_{cl} represent seasonal perturbations from these mean fields. The mean perturbation component of wind power, $\Phi_{ww_{cl}^x}^{mp}$, comprises of two terms - $\Phi_{ww_{cl}^x}^{mup\tau} = \iint_{z=0} \overline{u_{cl}} \tau_{s_{cl}}^{x'} dS$, which represents the effect of seasonal zonal wind stress fluctuations on mean surface currents and $\Phi_{ww_{cl}^x}^{m\tau pu} = \iint_{z=0} u'_{cl} \overline{\tau_{s_{cl}}^x} dS$, representing the effect of the mean wind field acting on seasonal surface current variations associated with the adjustment of the ocean.

In Figure 5.20, the relative contribution of mean perturbation terms $\Phi_{ww_{cl}^x}^{mup\tau}$ and $\Phi_{ww_{cl}^x}^{m\tau pu}$ of Equation 5.37

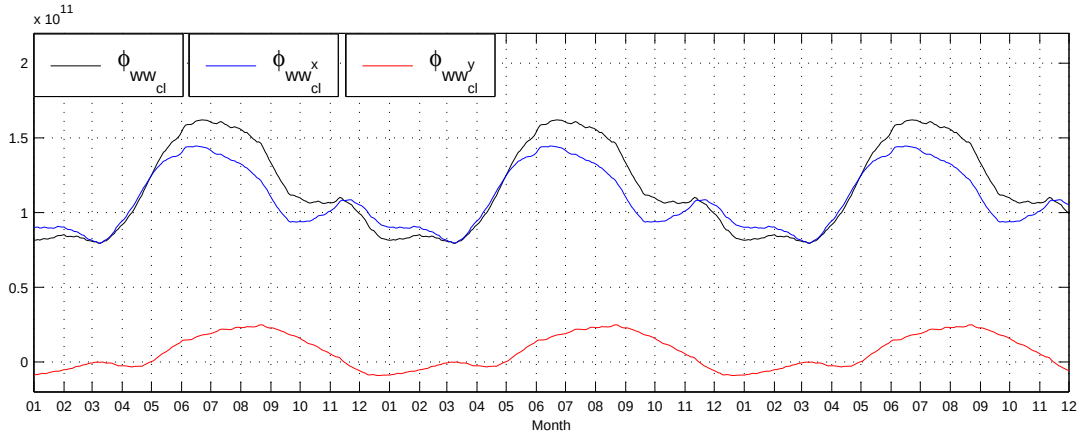


Figure 5.19: A comparison between seasonal fluctuations in the wind power term ($\Phi_{ww_{cl}}$), and its zonal ($\Phi_{ww_{cl}}^x$) and meridional ($\Phi_{ww_{cl}}^y$) components for the tropical Atlantic domain (8°S-8°N 60°W-15°E). Values are given in Js^{-1} . A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

as well as the perturbation term, $\Phi_{ww_{cl}}^{pp}$, to seasonal fluctuation in zonal wind power, $\Phi_{ww_{cl}}^x - \Phi_{ww_{cl}}^{mm}$, is assessed. Seasonal fluctuations in wind power are seen to be predominantly driven by the mean perturbation terms, $\Phi_{ww_{cl}}^{mup\tau}$ and $\Phi_{ww_{cl}}^{mtpu}$. This behaviour indicates that seasonal changes in zonal wind stress forcing as well as seasonal changes in surface currents, as the ocean adjusts to seasonal wind forcing, play an important role in influencing seasonal wind power fluctuations.

The nature of seasonal fluctuations in the mean perturbation term $\Phi_{ww_{cl}}^{mup\tau}$ reveals some interesting insight into the ability of wind stress forcing to do work on the ocean over different regions within the tropical domain. Figure 5.21 compares seasonal changes in the magnitude of the mean perturbation term, $\Phi_{ww_{cl}}^{mup\tau}$, with seasonal fluctuations in zonal wind stress integrated over the tropical Atlantic domain (8°S-8°N 60°W-15°E) and 8°S-2°N 50°W-0°E. Evident in Figure 5.21 is that the character of seasonal fluctuations in zonal wind stress integrated over the tropical Atlantic domain (8°S-8°N 60°W-15°E) differ substantially to that of the mean perturbation term. $\Phi_{ww_{cl}}^{mup\tau} = \iint_{z=0} \overline{u_{cl}} \tau_{s_{cl}}^x ' dS$ represents the contribution of seasonal zonal wind stress fluctuations (acting on the mean surface current field) to seasonal fluctuations in wind power term. As a result, one would expect seasonal changes in the intensity of zonal wind stress forcing integrated over the tropical Atlantic and this component of the wind power term to coincide. However, spatial variations in the mean surface velocity field affect the ability of zonal wind stress to do work on the ocean. This behaviour is shown by Figure 5.22, which depicts the mean zonal surface velocity field (u) for the tropical Atlantic domain, and Figure 5.21, which depicts the improved correspondence between seasonal fluctuations in the mean perturbation term $\Phi_{ww_{cl}}^{mup\tau}$ and zonal wind stress integrated over 8°S-2°N 50°W-0°E (as demarcated in Figure 5.22). A stronger westward component in the mean surface currents between 8°S-2°N and 50°W-0°E means that seasonal zonal wind stress variability in this region has a greater impact in driving seasonal APE variability.

In March/April, zonal wind stress integrated over 8°S-2°N 50°W-0°E is at its seasonal minimum (Figure 5.21), together with APE values, as the ITCZ is nearest the equator. From April onwards, zonal wind stress in this region increases as the ITCZ starts to migrate northwards. Wind power increases (Figure 5.20) resulting in positive buoyancy power values from May (Figure 5.17b) which increase APE as the cold tongue

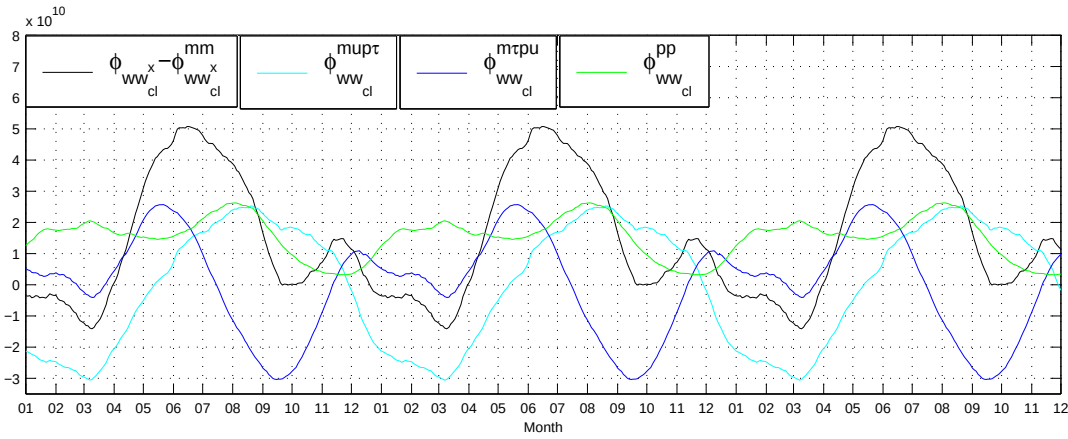


Figure 5.20: A comparison between seasonal fluctuations in the zonal component of the wind power term from its temporal mean ($\iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS - \iint_{z=0} \overline{u_{cl} \tau_{s_{cl}}^x} dS$), the mean perturbation terms, $\Phi_{ww_{cl}}^{mup\tau}$ and $\Phi_{ww_{cl}}^{m\tau pu}$ as well as the perturbation term, $\Phi_{ww_{cl}}^{pp}$ for the tropical Atlantic domain (8°S-8°N 60°W-15°E). Values are given in Js^{-1} . A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

develops. Westward surface currents accelerate during this time as the tropical Atlantic ocean adjusts to this sudden increase in zonal wind stress. This acceleration of westward surface currents is illustrated in Figure 5.23, which depicts the simulated seasonal perturbations in the zonal surface velocity averaged between 4°S and 4°N. The positive contribution of this enhanced westward flow on the wind's ability to do work on the ocean is expressed by term $\Phi_{ww_{cl}}^{m\tau pu}$ in Figure 5.20.

Zonal wind stress continues to increase term $\Phi_{ww_{cl}}^{mup\tau}$ in Figure 5.20 up until August when equatorial APE is at its maximum and eastern-central equatorial Atlantic SSTs are at their minimum. This enhanced wind stress does not however manage to further increase wind power as seasonal wind power values peak in June (Figures 5.16 and 5.20). This timing is due to the influence of surface currents on the ability of zonal wind stress changes to do work on the ocean (term $\Phi_{ww_{cl}}^{m\tau pu}$ in Figure 5.20).

The deceleration of westward surface flow between July-October is thought to be largely driven by the delayed response of the ocean to the sudden onset of the trades in May-June rather than a direct response of the ocean to wind variations (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009). As described in Chapter 2, transients are excited by the abrupt seasonal change in zonal wind stress which occurs in May (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009). The impact of these seasonally excited transients on equatorial surface currents is shown in Figure 5.23. Seasonal equatorial SSH and zonal surface current variability in the tropical Atlantic possesses characteristics of a basin mode (Ding et al., 2009). Firstly, a pronounced semi-annual cycle in equatorial SSH and zonal surface currents is evident even though the semi-annual cycle in surface winds is relatively weak in comparison with annual forcing, and secondly, there is an in quadrature relationship between the zonal SSH gradient and zonal surface currents from March to August. Ding et al. (2009) proposed that the semi-annual component in zonal winds excites the second baroclinic basin mode, thus forcing the prominent semi-annual component in SSH and surface zonal currents despite relatively weak forcing. Unlike the Pacific (Yu and McPhaden, 1999), boundary (coastal) reflections are seen to play an important role in seasonal SSH variability in the tropical Atlantic (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009). Seasonally,

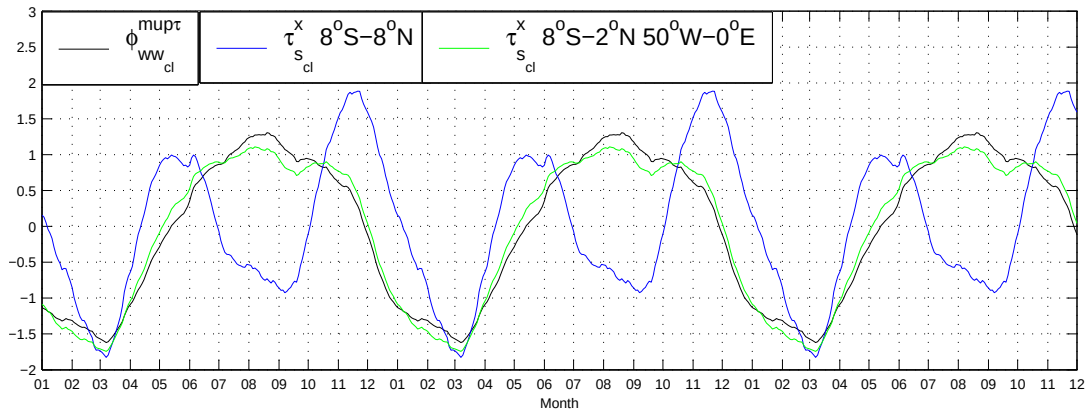


Figure 5.21: A comparison between seasonal fluctuations in the mean perturbation term, $\Phi_{ww_{cl}}^{mup\tau}$, seasonal fluctuations in zonal wind stress integrated over the tropical Atlantic domain (8°S-8°N 60°W-15°E), and seasonal fluctuations in zonal wind stress integrated over 8°S-2°N 50°W-0°E. Values have been normalised. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

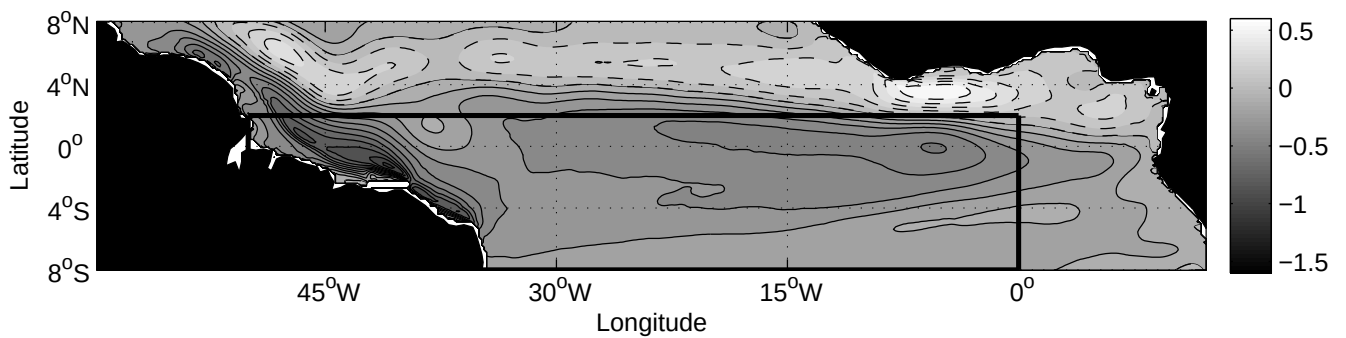


Figure 5.22: The mean zonal surface velocity field (u) for the tropical Atlantic domain, based on output from the ROMS-TAtl simulation (1980-2004). Zonal surface velocities are given ms^{-1} .

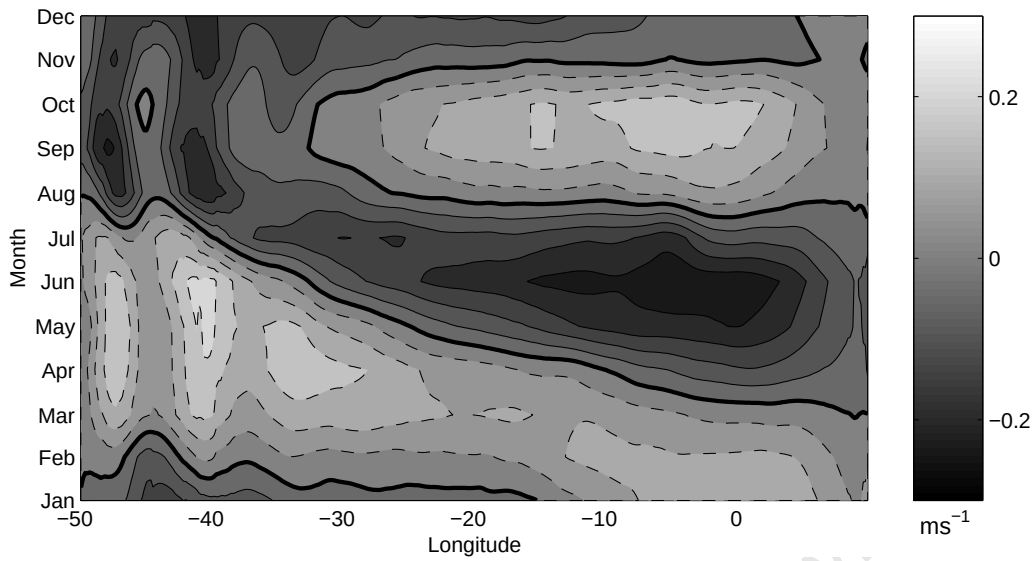


Figure 5.23: Seasonal zonal surface current perturbations from the annual mean averaged between 4°S-4°N. This figure is based on output from the ROMS-TAtl simulation (1980-2004). Zonal surface velocity perturbations are given ms^{-1} .

the contribution of Kelvin and Rossby waves reflected at the eastern and western boundaries to simulated SSH variability is equal to directly forced waves (Ding et al., 2009). Seasonally excited Kelvin and Rossby wave propagation results in the acceleration of westward surface flow between April-June followed by a deceleration of westward surface flow between July-October and the subsequent acceleration in November-December (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009). The effect of these seasonal surface current changes on the ability of the wind to do work on the ocean is seen in term $\Phi_{wwcl}^{m\tau pu}$ (Figure 5.20) as it peaks in May/June, reaches a minimum in September/October and then a secondary peak in December.

Figure 5.24 illustrates the spatial structure and temporal evolution of the climatological wind power field over the tropical Atlantic. The spatial pattern and temporal evolution associated with the seasonal zonal wind stress forcing and surface currents described in Chapter 2 are evident. Along the coast of South America, high wind power values are associated with the NBC throughout the year. In the north western tropical Atlantic, large negative wind power values are associated with the retroflection of the NBC and the NECC where easterly zonal winds act against eastward surface currents. Seasonal variations in the strength and eastward extent of these negative wind power values correspond with seasonal variability in the NECC. Despite the large magnitude of positive and negative wind power values in the north-western basin, their spatial extent is limited. It is seasonal variability in wind power values over the 8°S-2°N and 50°W-0°E region that correspond best with the domain integrated seasonal fluctuations in wind power seen in Figure 5.20. The increase in wind power values over this region between April and June (Figures 5.24d-5.24f) corresponds with the sudden increase in domain integrated wind power during this period (Figure 5.20), and is due to the acceleration of both the equatorial trades and westward surface currents within this region. Furthermore, the decline in domain integrated wind power between July and October (Figure 5.20) is associated with the westward propagation and decreased extent of positive seasonal wind power within this region (Figures 5.24g-5.24j). The second somewhat weaker increase in domain integrated wind power between November

and December (Figure 5.20) likewise corresponds to an increases in wind power within this region (Figures 5.24k and 5.24l).

Figures 5.25 and 5.26 show the zonal wind stress and surface current variability responsible for the wind power fluctuations (Figures 5.24d-5.24j) associated with the development and decay of the cold tongue between April and September. Figure 5.25 shows the spatial structure and temporal evolution of seasonal zonal wind stress perturbations with respect to their annual mean values, while Figure 5.26 shows the spatial structure and temporal evolution of seasonal surface current perturbations with respect to their annual mean values. The zonal wind stress variability shown in Figure 5.25 acts as an indicator of the atmospheric response associated with the development and decay of the cold tongue between April and September, while the zonal surface current variability shown in Figure 5.26 acts as an indicator of the oceanic response.

As continental convection moves from the equator into the northern hemisphere in April/May, it accelerates southerly winds over the Gulf of Guinea (Mitchell and Wallace, 1992; Okumura and Xie, 2004). Furthermore, the growth of asymmetric conditions about the equator is associated with the intensification of easterly trades in the eastern, equatorial-southern tropical Atlantic due to the penetration of the south east trades into the northern hemisphere (Figures 5.25a and 5.25b). The increased westward zonal wind stress appears first in the eastern equatorial Atlantic and moves westward. By June, this region of enhanced easterly trades has grown and is more equatorially symmetric (Figure 5.25c). The region of enhanced easterly trades grows and propagates westward between June and August (Figure 5.25d and 5.25e). This enhancement of the equatorial trades in the central-western Atlantic is thought to be related to the growth of a Bjerknes feedback, seasonally excited as the ITCZ migrates northwards allowing the trades to intensify on the equator.

As mentioned previously, the sudden onset of enhanced zonal wind stress forcing east of 30°W in May (Figure 5.25b), due to its semi-annual nature, occurs at a much shorter time scale than the adjustment time scale of the equatorial Atlantic (Philander and Pacanowski, 1986a). This sudden intensification is seen to excite upwelling equatorial Kelvin waves and downwelling off-equatorial Rossby waves (Schouten et al., 2005; Ding et al., 2009). The spatial structure of the effect of these transients on zonal surface currents is seen in Figure 5.26. Figures 5.26b, 5.26d, 5.26f, 5.26h, 5.26j and 5.26l, based on the observed surface current climatology product of Lumpkin and Garraffo (2005) have been added to validate the simulated fields shown in Figures 5.26a, 5.26c, 5.26e, 5.26g, 5.26i and 5.26k. In May, westward surface currents within the central-eastern equatorial Atlantic are accelerated (Figures 5.26c) and these positive zonal surface current anomalies propagate westwards (Figures 5.26c, 5.26e, 5.26g, 5.26i and 5.26k). This westward propagation of zonal surface current was attributed to first meridional mode Rossby waves by Ding et al. (2009), as within their simulated results, the Rossby wave contribution is seen to out-way the Kelvin wave contribution. The westward propagation of zonal wind forcing is also thought to reinforce this westward propagation (Ding et al., 2009). The upwelling equatorial Kelvin waves (downwelling off-equatorial Rossby waves) are then reflected at the eastern and western boundaries and act to decelerate westward surface currents near the equator between July and October (Figures 5.26g, 5.26j and 5.26k) (Ding et al., 2009).

Similarly in the Pacific, on inter-annual time scales, transient induced changes in surface currents affect the ability of the wind to do work on the ocean (Goddard and Philander, 2000). This ocean memory process, in accordance with the delayed oscillator theory, is seen to be responsible for the transition from La Niña to El Niño or vice versa. Surface current anomalies driven by the thermocline perturbations associated with off-equatorial Rossby waves arriving in the western equatorial Pacific affect the ability of the wind to do

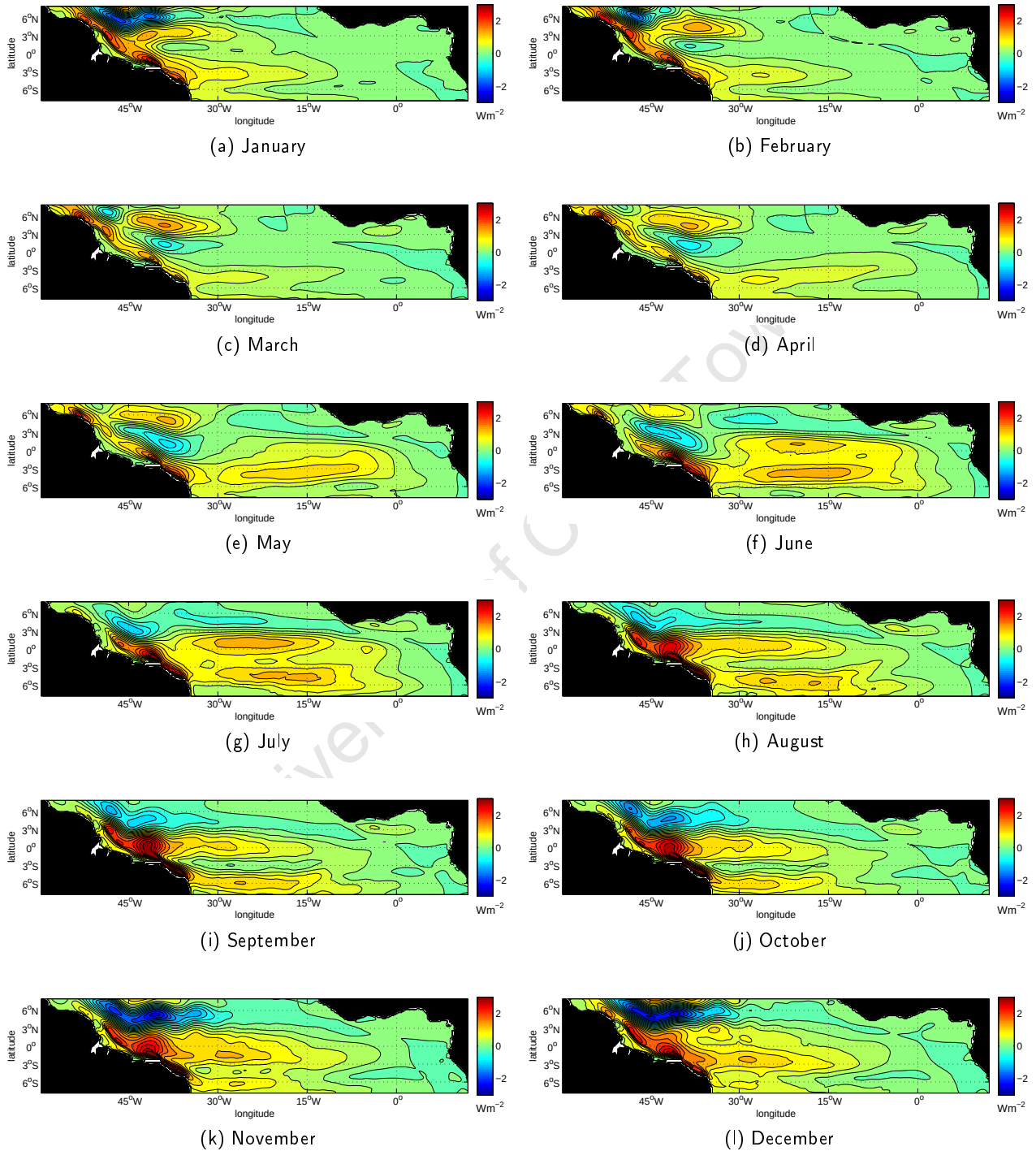


Figure 5.24: Monthly maps illustrating the spatial structure in the climatological wind power field (Wm^{-2}) over the tropical Atlantic domain (8°S - 8°N , 60°W - 15°E).

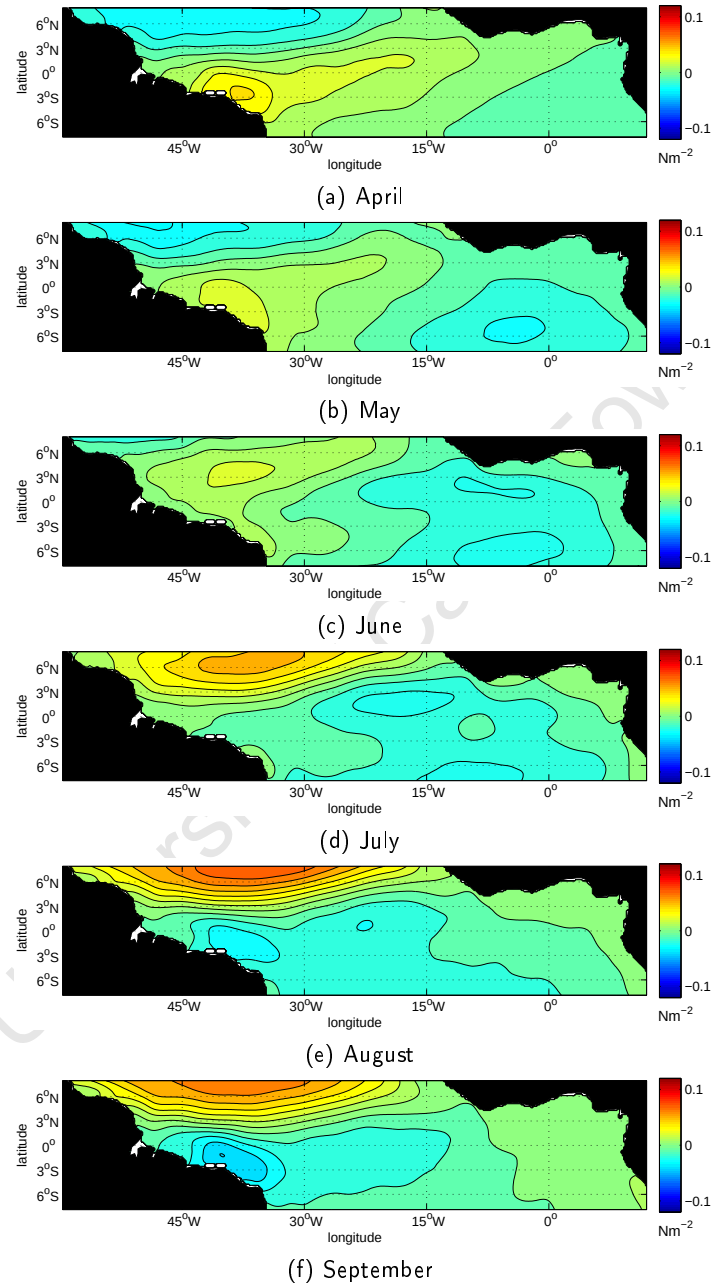


Figure 5.25: Monthly maps illustrating the spatial structure in climatological zonal wind stress perturbations from the annual mean (Nm^{-2}) over the tropical Atlantic domain (8°S - 8°N 60°W - 15°E) between April and September.

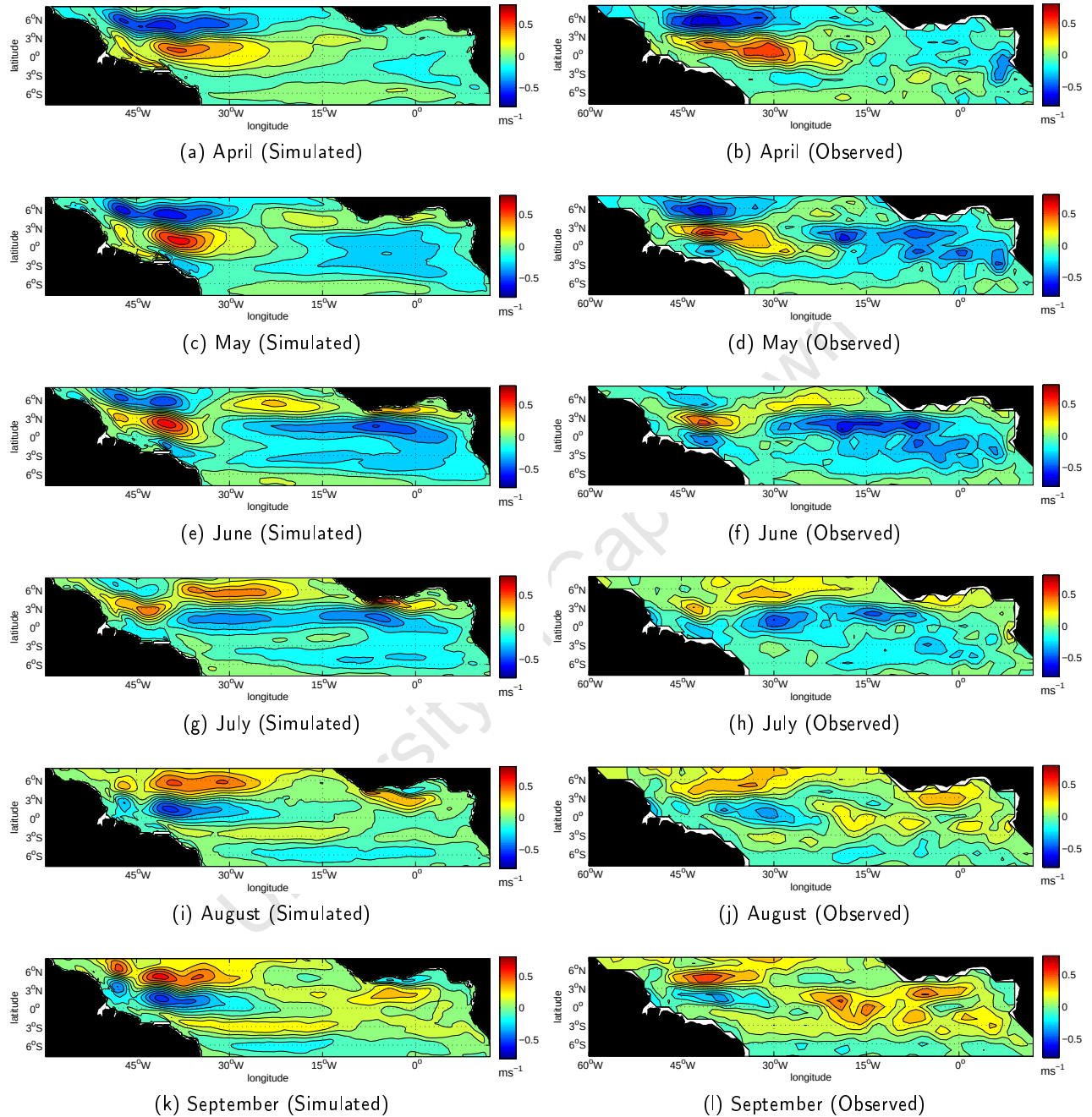


Figure 5.26: Monthly maps illustrating the spatial structure in climatological zonal surface current perturbations from the annual mean (ms^{-1}) over the tropical Atlantic domain (8°S–8°N 60°W–15°E) between April and September. Figures in the left column (a, c, e, g, i and k) are based on output from the ROMS-TAtI simulation (1980–2004), while figures in the right column (b, d, f, h, j and l) are based on the surface current climatology product of Lumpkin and Garraffo (2005) (www.aoml.noaa.gov/phod/dac/drifter_climatology.html).

work on the ocean (Goddard and Philander, 2000).

ENSO is a free mode of the coupled ocean-atmosphere system in the tropical Pacific that operates on inter-annual time scales with the Bjerknes feedback leading to the growth of events and the delayed oscillator mechanism setting the period of oscillations (Hirst, 1986; Suarez and Schopf, 1988). The seasonal cycle in the Atlantic is on the other hand a forced oscillation. However, between the months of April and October, seasonally excited coupled ocean-atmosphere interaction within the tropical Atlantic is responsible for a response which deviates from seasonal forcing (Figure 1.1). As outlined in Chapter 2, an asymmetric SST-mode excited in April/May leads to the growth of asymmetric conditions as the cold tongue-ITCZ complex develops in the Atlantic (Chang and Philander, 1994). The results presented above suggest that, in the tropical Atlantic, this process is accompanied by the excitation of a coupled mode, referred to as the thermocline mode, with similar properties to the delayed oscillator mode operating inter-annually in the Pacific. The Bjerknes feedback appears to be seasonally excited in April/May resulting in the steep seasonal increase in equatorial Atlantic APE and associated decline in central-eastern basin SST. This positive feedback is then damped by a delayed negative feedback mechanism, similar to the negative feedback mechanism operating inter-annually in the Pacific, as decreased APE observed in September-October is seen to be due to the influence of seasonal surface current fluctuations on the ability of seasonally enhanced surface winds to do work on the ocean. This seasonally excited coupled variability has the potential to oscillate. In November/December, this potential is evidenced as transient induced zonal current changes once again increase the ability of the wind to do work on the ocean (Figure 5.20) thereby increasing the buoyancy flux during these months (Figure 5.14). As mentioned in Section 5.2, the thermocline is seen to shoal slightly as the decrease in equatorial APE slows during these months. Furthermore, the zonal component of the trades intensifies slightly and the rate of increase in Atl3 SST is seen to decrease as a somewhat weaker second cooling event during the months of November-December is confined to a small near equatorial region within the central-eastern equatorial Atlantic (Okumura and Xie, 2006). Undamped, this coupled mode could potentially go on to shoal the thermocline as it does between May and August, however by January-March it is strongly damped by the forced seasonal cycle as solar forcing manages to warm south equatorial SST enough to weaken the cold tongue-ITCZ complex. As a result, the ITCZ migrates southwards back towards the equator, thereby suppressing the equatorial conditions which are necessary to promote and sustain the growth and oscillation of such a coupled mode.

Figure 5.27 summarises the seasonal behaviour in key variables of the above mentioned seasonally excited thermocline mode, namely wind power and APE. Figure 5.27a displays a phase diagram between seasonal APE values and the wind power term ($\Phi_{ww_{cl}}$), the dominant term driving seasonal changes in APE. Figure 5.27b compares seasonal fluctuations in APE, the APE tendency term ($dAPE_{cl}/dt$), and the wind power term ($\Phi_{ww_{cl}}$). In the Pacific, the presence of free oscillations associated with the delayed oscillator mode results in an out of phase relationship between wind power and APE on inter-annual timescales. This out of phase relationship is due to the fact that APE changes are driven predominately by wind power and wind power displays a dependence on APE via its relationship with surface current anomalies and the ability of the wind to do work on the ocean. This in quadrature relationship between wind power and APE observed inter-annually in the Pacific results in a circular trajectory that passes through all four quadrants of a phase diagram with wind power on the horizontal axis and APE on the vertical axis (Fedorov, 2007). The corresponding phase diagram for the seasonal cycle in the Atlantic is shown in Figure 5.27a. Only between April-September is a circular relationship between APE and the wind power term evident in Figure 5.27a, as

it is during these months that the seasonally excited thermocline mode of coupled variability plays an active role in the seasonal cycle.

Figure 5.18 depicts seasonal changes in the BPE of the equatorial and tropical domains. BPE changes within these two regions are seen to be approximately in-phase with the exception that tropical BPE values peak approximately a month later than that in the equatorial Atlantic. As with APE, this result is ascribed to the fact that this larger domain takes longer to adjust than the equatorial region to seasonal changes in the wind forcing. As pointed out in the methodology section of Chapter 3 (Section 3.2), seasonal changes in the mass and BPE of the equatorial and tropical Atlantic regions are in-phase with seasonal changes in their heat content. This relationship between heat content, mass and BPE suggests firstly, that seasonal variability in the density field appears to be primarily determined by seasonal variability in the temperature field, and secondly, that seasonal BPE change are primarily the result of seasonal changes in the mass of the equatorial and tropical domains rather than seasonal changes in vertical stratification due to mixing. These suggestions are reinforced by the results illustrated in Figure 5.28 which depict the climatological balance of terms for the BPE evolution equation given in Section 5.3.2 (Equation 5.19). For the tropical and equatorial domains, the relative contribution of BPE changes due the advection of the density field, $\Phi_{bpa_{cl}} = A_{BPE_{cl}}$ (Equation 5.19), versus BPE changes due to the combined effects of density diffusion and the penetrative solar radiation flux, $\Phi_{bpd_{cl}} = D_{BPE}$ (Equation 5.19), is shown in Figure 5.28. Also shown is the portion of $\Phi_{bpd_{cl}}$ due to seasonal changes in surface buoyancy forcing ($\Phi_{bpd_{cl}}^{surfacebuoyancy}$ - note this term includes the penetrative solar radiation flux). The rather small residual between $\Phi_{bpd_{cl}}$ and $\Phi_{bpd_{cl}}^{surfacebuoyancy}$ (not shown) represents seasonal changes in BPE due to mixing within the domain and diffusion at the horizontal and bottom surfaces of the domain.

For both the equatorial and tropical domains, seasonal BPE fluctuations are largely driven by fluctuations in the advection of the density field ($\Phi_{bpa_{cl}}$) (Figure 5.28). Advection of the density field ($\Phi_{bpa_{cl}}$) acts to increase BPE throughout the year for the tropical domain (with values reaching zero between July and October) and from November-July for the equatorial domain (decreasing BPE between July and October) (Figure 5.28). These seasonal changes in $\Phi_{bpa_{cl}}$ for the tropical domain are consistent with observed seasonal changes in the escape of heat across 8°N. As described in Chapter 2, escape of heat across 8°N is regulated by seasonal changes in the NBC/NECC current system in response to changes in wind stress forcing associated with the annual migration of the ITCZ. The 8°S-8°N region exports heat for most of the year therefore acting to increase BPE, except between July-October when northward transport is blocked as the NBC veers offshore to feed the NECC and increased wind stress curl causes the mixed layer to deepen south of the NECC (8°N) and shoal to the north (Philander and Pacanowski, 1986b; Lee and Csanady, 1999). Seasonal fluctuations in $\Phi_{bpd_{cl}}$ play a smaller role and act to decrease BPE throughout the year in both the tropical and equatorial domain. This term is negative as it is largely due to warming by surface buoyancy forcing ($\Phi_{bpd_{cl}}^{surfacebuoyancy}$ shown in Figure 5.28). In both domains, the surface buoyancy forcing term reaches a minimum in March/April when surface heating over the respective domain is at its annual maximum and a maximum in June/July when surface heating is at its minimum. These results suggest that seasonal changes in the BPE of the equatorial and tropical domains is controlled by seasonal changes in the warm water formation (in this case regarded in terms of surface buoyancy forcing rather than entrainment) and escape (largely controlled by off-equatorial wind stress curl) (Lee and Csanady, 1999; Bunge and Clarke, 2009).

Like seasonal APE, seasonal BPE changes therefore appear to be largely a product of the oceanic circulation

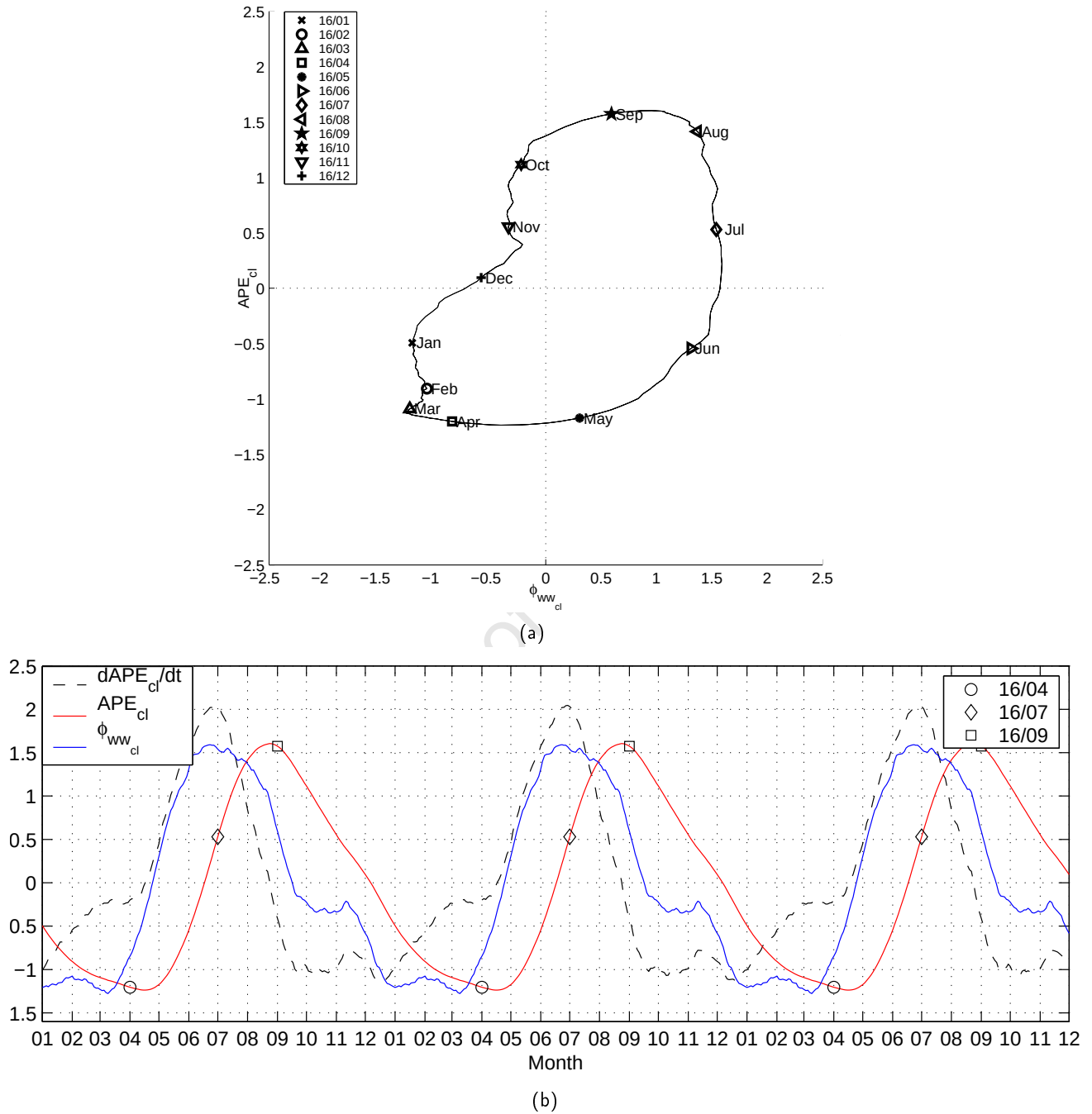


Figure 5.27: (a) APE verse wind power phase diagram associated with the seasonal cycle in the tropical Atlantic domain (8°S-8°N 60°W-15°E). (b) A comparison between seasonal fluctuations in APE (APE_{cl}), the APE tendency term ($dAPE_{cl}/dt$) and the wind power term ($\Phi_{ww_{cl}}$) for the tropical Atlantic domain (8°S-8°N 60°W-15°E). Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability.

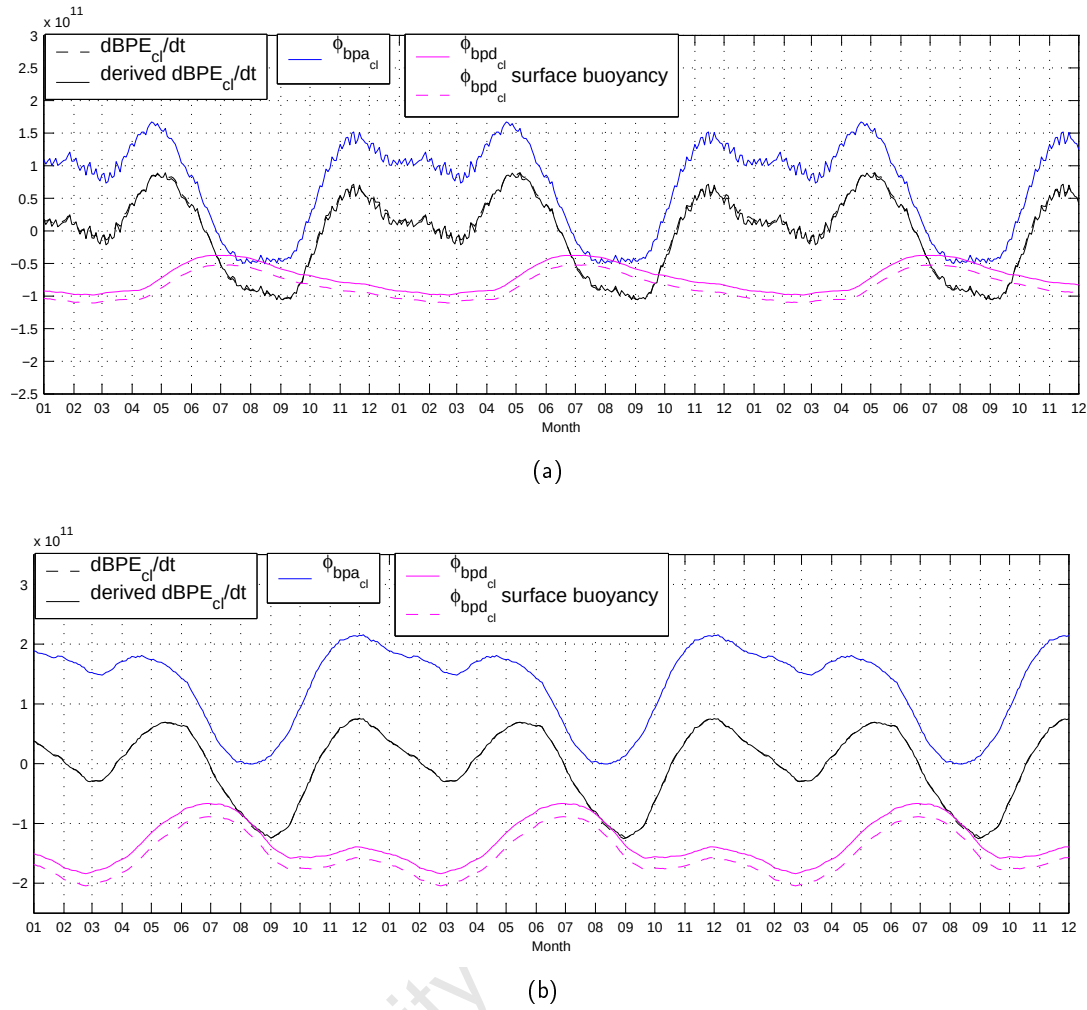


Figure 5.28: The climatological balance of terms (Js^{-1}) for the BPE evolution equation (Equation 5.19), assessed over (a) the equatorial Atlantic domain ($3^{\circ}\text{S}-3^{\circ}\text{N}$ $60^{\circ}\text{W}-15^{\circ}\text{E}$ $0-400\text{m}$) and (b) the tropical Atlantic domain ($8^{\circ}\text{S}-8^{\circ}\text{N}$ $60^{\circ}\text{W}-15^{\circ}\text{E}$ $0-400\text{m}$). A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

adjustments in response to fluctuations in seasonal zonal wind stress and its curl. As seen in Figure 5.18, seasonal APE and BPE changes are out of phase. This result is in agreement with the finding of Bunge and Clarke (2009) that changes in the slope of the equatorial thermocline are in quadrature with variations in the amount of warm water above the thermocline between $2^{\circ}\text{N}-2^{\circ}\text{S}$. The explanation for this relationship provided by Bunge and Clarke (2009) is akin to that for the equivalent relationship observed in the Pacific on inter-annual timescales, namely the equatorial heat content recharge-discharge mechanism of Jin (1997). The out of phase relationship seen between zonal slope of the equatorial thermocline (APE) and the mean depth of the equatorial thermocline (BPE) arises via the fact that they are both related to changes in the intensity of zonal wind stress at the equator: APE changes are directly forced by the associated wind power changes and BPE change are forced via the associated wind stress curl changes. Therefore, as revealed by the results of the seasonal APE evolution analysis described above, seasonal BPE changes do not directly influence seasonal APE changes and their out of phase relationship arises out of the fact that they are both related to seasonal changes in the intensity of zonal wind stress at the equator.

However, the shallow mean depth of the thermocline within the equatorial region in April, May and June could

potentially act in favour of the seasonal excitation of the Bjerknes feedback by preconditioning the region. Is it possible that anomalously high (low) BPE during April-May could lead to an enhanced (decreased) seasonally excited Bjerknes feedback? This raises the question: Can inter-annual SST variability associated with the zonal mode be explained in terms of a modulation of this newly identified, seasonally active thermocline mode and furthermore, could anomalous BPE particularly during April-May, be one of the factors responsible for the modulation of this seasonally active process? This question is addressed in the following chapter.

5.3.4 Conclusions

Taking into account seasonal changes in BPE when evaluating seasonal APE, the analysis conducted in this section shows that seasonal equatorial and tropical basin APE changes are driven primarily by buoyancy forcing associated with the horizontal redistribution of warm surface waters. The source of this buoyancy forcing is found in the KE evolution equation. For the equatorial domain, buoyancy power fluctuations are driven not only by seasonal changes in the work done by the wind within the equatorial domain but also by the response of the greater tropical Atlantic domain to seasonal fluctuations in wind forcing via its impact along the boundaries of the equatorial domain as seasonally excited transients enter and exit the equatorial domain. Therefore to better understand the processes behind the seasonal redistribution of warm surface waters and associated APE changes within the equatorial Atlantic, the seasonal redistribution of warm surface waters driven by wind stress forcing over the tropical Atlantic region, and the oceanic response to this forcing, are considered. For the tropical Atlantic domain, seasonal changes in the buoyancy power term and hence APE, are seen to be primarily dictated by seasonal changes in the work done by wind stress acting on surface currents within the domain. Extending the domain over which APE is evaluated to include the tropical domain simplifies the dynamics governing seasonal APE variability down to one dominant process, namely the work done by the wind over the tropical Atlantic domain.

In Chapter 3 and Section 5.2, evidence was offered to support the hypothesis that the Bjerknes feedback mechanism is excited seasonally and plays an active role in the seasonal development of the Atlantic cold tongue. This section provides further evidence that one of the key elements of the Bjerknes feedback is seasonally active in the Atlantic as zonal wind stress in the western-central equatorial Atlantic does work on the ocean driving thermocline depth changes in the east as APE increases and the cold tongue develops between April and August.

Furthermore, evidence of the decay mechanism behind this seasonally excited Bjerknes feedback mechanism is provided as transient induced changes in surface currents, associated with the delayed response of the tropical Atlantic to seasonal changes in wind stress forcing, affect the ability of the wind to do work on the ocean. Seasonal changes in surface currents as the ocean adjusts to seasonal wind forcing are seen to play an important role in dictating seasonal wind power fluctuations and hence the seasonal evolution of APE.

In the Pacific, inter-annual APE anomalies are grown by the Bjerknes feedback mechanism until the delayed response of the ocean affects the ability of the wind to do work on the ocean. The results of the energetics analysis presented above indicate that a similar process is seasonally excited in the Atlantic between April and October after which point, seasonal forcing regains control. The seasonally excited growth of asymmetric conditions about the equator in April appears to be accompanied by a seasonally excited Bjerknes feedback in the Atlantic which is then damped between August-October by a delayed, negative, ocean memory, feedback

mechanism whereby seasonal surface current anomalies effect the ability of seasonally enhance zonal winds to do work on the ocean.

As mentioned in Chapter 1, inter-annual variability in the equatorial Pacific can be viewed either in terms of the delayed oscillator paradigm or the recharge oscillator paradigm. The recharge oscillator idealisation differs from the delayed oscillator idealisation in its approximation of the phase transition mechanisms. The phase-transition mechanism in the case of the delayed oscillator idealisation is attributed to the delay associated with planetary wave propagation and western boundary wave reflection. In the case of the recharge oscillator idealisation, the ocean wave propagation process is not at the centre of the delayed negative feedback, but rather regarded as a part of the oceanic adjustment in which heat and mass are re-distributed under changing wind forcing. The recharge-discharge of equatorial zonal mean heat content is viewed as the phase-transition mechanism (Jin, 1997). The APE-wind power perspective of coupled equatorial variability in the Pacific lends itself to the delayed-oscillator paradigm as the effect of transients on surface currents and hence the ability of the wind to do work on the ocean is seen as the delayed negative feedback mechanism (Goddard and Philander, 2000). Similarly, the results of the energetics analysis of seasonal equatorial Atlantic variability presented in this section lend themselves to the perspective that the delayed negative feedback mechanism responsible for the decay of the cold tongue in the Atlantic is due to the delay associated with planetary wave propagation. However, as seen inter-annually in the Pacific, this seasonally active delayed negative feedback is associated with a recharge of equatorial zonal mean heat content (a decrease in BPE) in response to the intensification of the easterly trades due to the redistribution of mass between certain regions off the equator and the equatorial band. Only between April-September is a circular relationship in the phase diagram between APE and wind power evident, as it is during these months that a seasonally excited thermocline mode of coupled variability plays an active role in the seasonal cycle of the tropical Atlantic.

Finally, the question is raised: Can inter-annual SST variability associated with the zonal mode be explained in terms of a modulation of this newly identified, seasonally active thermocline mode? If so, what factors are responsible for the modulation of this seasonally active processes? One such factor might be inter-annual variability in BPE which appear to precondition the region for the seasonally excited Bjerknes feedback during April-May. The following Chapter 6, aims to address these questions.

Chapter 6

Inter-annual variability in the Equatorial Atlantic

6.1 Introduction

Given the supremacy of seasonal SST variability in the central-eastern equatorial Atlantic (Figure 1.4), one wonders whether the seasonal cycle in the Atlantic is so dominant that it is able to strongly influence the evolution of its inter-annual variability? Are the physical ocean processes involved in inter-annual SST variability different from those of the seasonal cycle, or are the processes the same except for a modulation in either phase or amplitude? The analysis carried out within this chapter aims to address these questions and the related questions posed in Chapter 1 pertaining to inter-annual variability associated with the zonal mode:

- Can inter-annual SST variability in the central-eastern equatorial Atlantic be explained in terms of a modulation of seasonally active coupled variability and, if so, what factors are responsible for the modulation of seasonally active processes?
- What is the role of ocean memory within central-eastern equatorial Atlantic SST variability?

The results obtained in Chapters 3 and 5 have established that remotely forced thermocline depth changes play a crucial role in driving seasonal SST changes, a process which only plays a significant role inter-annually in the Pacific. Furthermore, the relationship between seasonal SST, APE and zonal wind stress changes in the Atlantic supports the notion that the greater and steeper seasonal decrease in eastern-central basin SSTs observed in the equatorial Atlantic between April and August is due to the fact that the seasonally excited growth of asymmetric conditions about the equator is accompanied by a seasonally excited Bjerknes feedback as the thermocline shoals seasonally. This seasonally excited Bjerknes feedback is then damped between August-October by a delayed, negative, ocean memory feedback mechanism whereby seasonal surface current anomalies affect the ability of seasonally enhanced zonal winds to do work on the ocean. As a result, cool conditions do not persist as long as they do in the Pacific. Between April-September a circular relationship between APE and the work done on the ocean by the wind suggests that a seasonally excited thermocline mode of coupled variability driven by the seasonally excited Bjerknes feedback and its

subsequent decay, plays an active role in the tropical Atlantic's seasonal cycle. The difference in the time-scale of the largest SST fluctuations between the Pacific and Atlantic basins is therefore attributed to the fact in the Pacific the Bjerknes feedback and its associated delayed, negative feedback mechanism operates on inter-annual time-scales whereas in the Atlantic, it operates seasonally.

In light of these findings, the first of the two questions listed above is refined slightly: Can inter-annual SST variability associated with the zonal mode be explained in terms of a modulation of the seasonally active thermocline mode? If so, what factors are responsible for the modulation of this seasonally active processes?

As demonstrated by Goddard and Philander (2000), an assessment of the energetics of inter-annual equatorial ocean variability can shed light on the mechanisms governing ocean-atmosphere interactions i.e. the role of oceanic adjustment vs stochastic wind forcing such as wind bursts. Here, the tropical Atlantic ROMS-TAtl simulation is employed to assess inter-annual variability in the equatorial Atlantic from an energetics perspective.

The first results section of this chapter (Section 6.2), largely repeats the analysis conducted in Sections 3.3.3 and 3.3.4 of Chapter 3 based on SODA and ORCA data, except the results presented are based on ROMS-TAtl output. Results obtained in Section 6.2 are consistent with the results obtained based on SODA and ORCA data in Chapter 3 providing confidence in both the ROMS-TAtl simulation as well as the results obtained in Chapter 3. As in Chapter 3, the relationships between anomalous SST, APE and zonal wind stress obtained in Section 6.2 suggests that inter-annual Atl3 variability associated in the zonal mode may be best understood as the result of the modulation either in phase or amplitude of the Bjerknes feedback which is seasonally excited as the cold tongue develops between April and August.

The second results section (Section 6.3) explores the forcing behind inter-annual anomalies in equatorial and tropical Atlantic APE. The analysis of the processes driving APE anomalies within the tropical Atlantic conducted in Section 6.3 indicates that anomalous APE evolution within the tropical Atlantic is largely driven by anomalous zonal wind power over the tropical domain particularly during boreal summer months. Inter-annual APE anomalies like seasonal APE fluctuation are predominantly forced by fluctuations in wind power.

The third and final results section (Section 6.4) explores the relationship between anomalous wind power and Atl3 SST. It is shown that inter-annual tropical Atlantic wind power anomalies are largely the result of inter-annual modulations of seasonal wind power fluctuations associated with the seasonally excited thermocline mode identified in Section 5.3.3. The delayed negative feedback mechanism associated with the zonal mode events operates on much shorter time scales than that associated with ENSO. The ocean memory feedback mechanism associated with zonal mode events is seen in Section 6.4 as being largely associated within inter-annual modulations of the seasonally excited thermocline mode's delayed, negative feedback response between August and October.

The results obtained suggest that inter-annual variability in Atl3 SST correlated with anomalous APE can largely be regarded as inter-annual modulations of the seasonally excited Bjerknes feedback and its associated delayed, negative feedback. At the end of Section 5.3.3 it was pointed out that although seasonal APE changes were not directly forced by BPE changes, the shallow mean depth of the thermocline within the equatorial region between April and June potentially acts in favour of the seasonal excitation of the Bjerknes feedback by preconditioning the region. Based on this fact the following question was raised: Is it possible that anomalously high (low) BPE during April-May could lead to an enhanced (decreased) seasonally

excited Bjerknes feedback? The results obtained in Section 6.4 suggest that although there appears to be a relationship between the occurrence of anomalous BPE and zonal mode events it is an inconsistent one. While anomalous BPE may be one factor behind inter-annual modulations of the seasonally excited Bjerknes feedback, it is not the dominant process. Furthermore, results obtained in Section 6.4 suggest that tropical Atlantic BPE anomalies are primarily forced by inter-annual variability in the warm water escape process.

6.2 Anomalous APE-SST-Wind Relationship

Figure 6.1a depicts the relationship between inter-annual anomalies in the equatorial Atlantic APE (3°S-3°N 60°W-15°E 0-400m) and Atl3 SST based on ROMS-TAtl data from 1980-2004. The APE-SST relationship observed in Figure 6.1a is consistent with the APE-SST relationship revealed in Chapter 3 based on SODA and ORCA data from 1958-2004. The maximum correlation between equatorial Atlantic APE and Atl3 SST is observed at zero lag with a correlation coefficient of $r=-0.67$ ($r^*=-0.19$). This correlation is higher than the correlation coefficient of $r=-0.56$ ($r^*=-0.19$) based on SODA data (Figure 3.19) and that of $r=-0.62$ ($r^*=-0.19$) based on ORCA data (Figure A.11). The relationship observed in Figure 6.1a between anomalous APE and SST is indicative of the significant role remotely forced thermocline depth variations play in inter-annual Atl3 SST variability.

As highlighted in Chapter 3, this relationship between equatorial APE and central-eastern basin equatorial SST is considerably weaker than the corresponding relationship in the Pacific, a result that is in agreement with the weaker relationship observed between inter-annual SST anomalies and thermocline depth anomalies in the central-eastern equatorial Atlantic (Zebiak, 1993; Vauclair and du Penhoat, 2001; Keenlyside and Latif, 2007). This weaker relationship in the Atlantic than the Pacific is attributed to the fact that the Bjerknes feedback mechanism does not appear to be associated with all anomalous Atl3 SST events (Zebiak, 1993; Chang et al., 2006a). Unlike the Pacific, where inter-annual SST anomalies are predominantly forced by the Bjerknes feedback mechanism, a stronger contribution from external forcing and disparate local air-sea interactions (i.e. the WES feedback or the Ekman feedback) is thought to occur in the Atlantic (Chang et al., 2006a).

Figure 6.1b explores the seasonality in the relationship between inter-annual anomalies in equatorial Atlantic APE and Atl3 SST based on ROMS-TAtl data. The seasonal dependence in the correlation between inter-annual anomalies in equatorial Atlantic APE and Atl3 SST observed in Figure 6.1b is similar to that observed in Chapter 3 based on SODA (Figure 3.20) and ORCA (Figure A.12) data. Anomalous Atl3 SST correlates best with anomalous APE between April and August. This finding is in agreement with seasonality in the regression between SST and thermocline depth anomalies (assessed using sea level anomalies as a proxy) obtained by Keenlyside and Latif (2007). Also worth noting is the secondary peak in the correlation between anomalous equatorial APE and Atl3 SST during December and January. On the other hand, between January and April the correlation between inter-annual anomalies in equatorial Atlantic APE and Atl3 SST drops substantially.

Figure 6.2 illustrates the relationship between inter-annual anomalies in SST, the average temperature of the surface mixed layer, and the average temperature of the subsurface layer for the Atl3 region. As expected inter-annual anomalies in the mean temperature of the surface mixed layer correspond closely with SST anomalies. A correlation coefficient of $r=0.98$ ($r^*=0.2$) is observed between these two variables (Figure

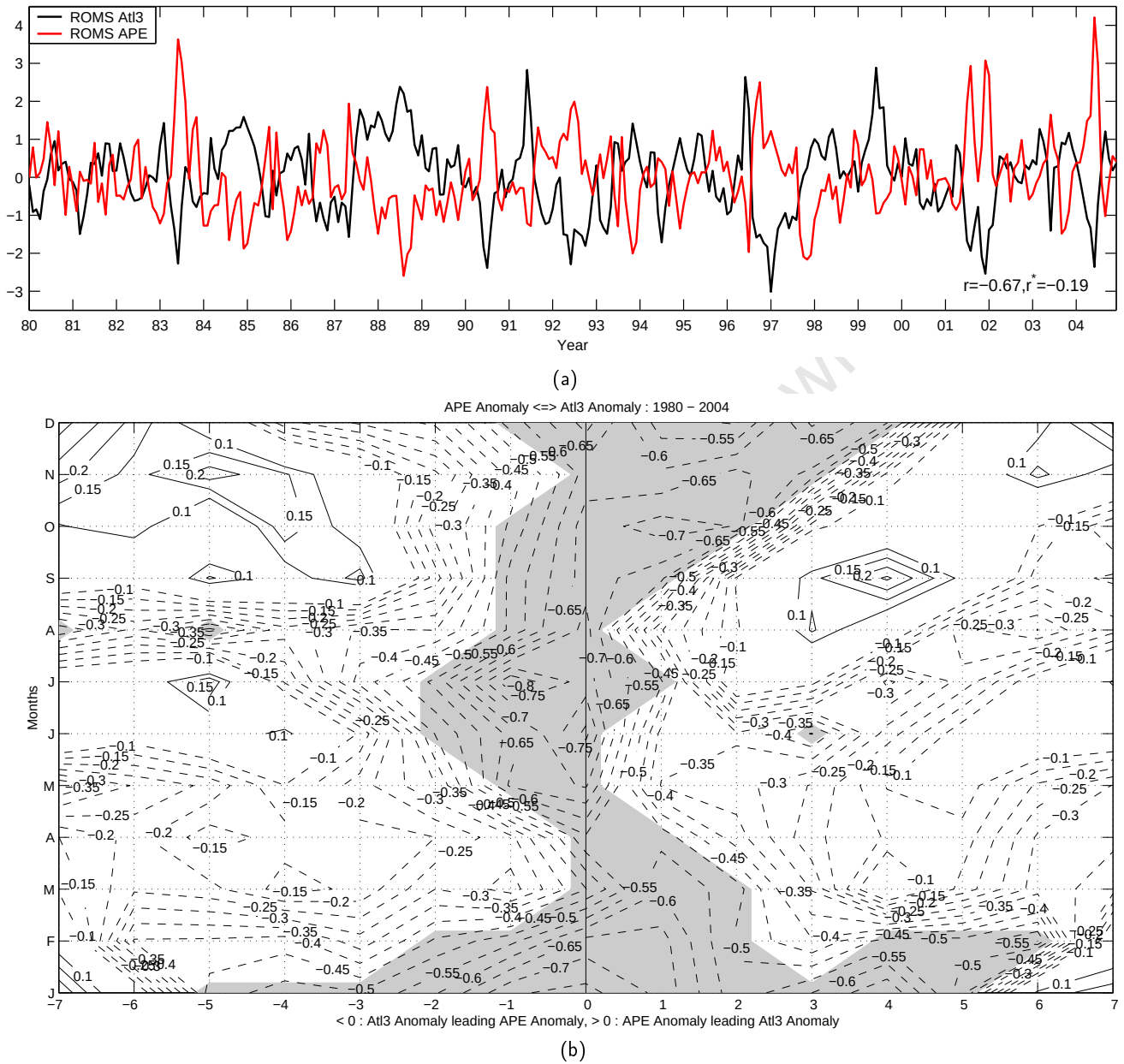


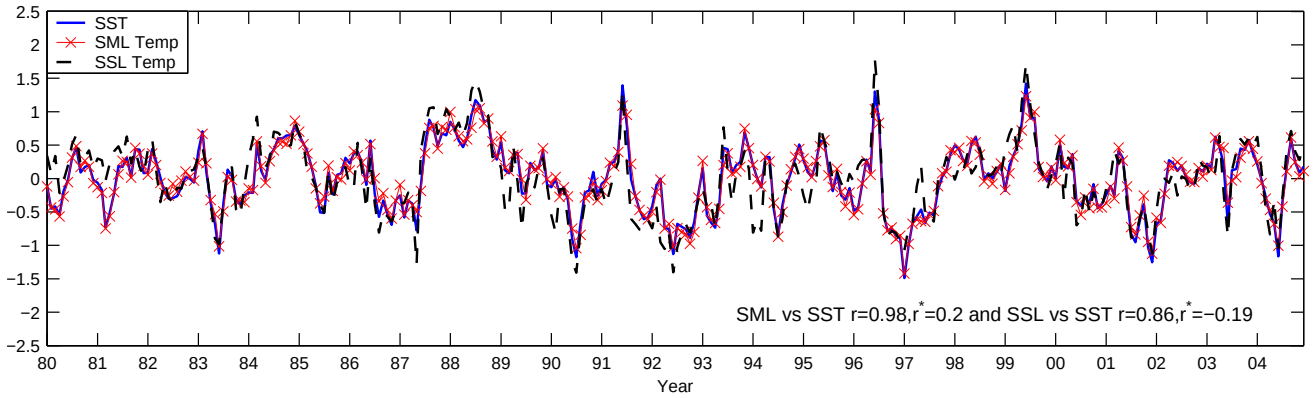
Figure 6.1: (a) A comparison between inter-annual Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised. (b) The seasonal dependence in correlations between Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST and APE anomalies were calculated based on ROMS-TAtl output. APE values were calculated following Equation 3.2.

6.2a). The relationship between inter-annual anomalies in the mean temperature of the subsurface layer and SST, with a correlation coefficient of $r=0.85$ ($r^*=0.19$), is slightly less robust (Figure 6.2a). There are clearly periods where subsurface temperature anomalies are not reflected in the SST field. To explore whether there is any seasonality in these periods when SST and subsurface temperature anomalies do not correspond, the seasonal dependence in the correlation between inter-annual anomalies in SST and the average temperature of the subsurface layer is shown in Figure 6.2b. Seasonally, inter-annual anomalies in the mean temperature of the subsurface layer do not correspond closely with SST anomalies between January and April (Figure 6.2b). This result is consistent with seasonality in the correlation between Atl3 SST and equatorial Atlantic APE anomalies observed in Figure 6.1b, suggesting that subsurface temperature anomalies associated with thermocline depth variability are less readily reflected in the SST field during these months. In Figure 6.3, seasonal variability in the average wind stress over the Atl3 region is shown. The weak relationship between inter-annual subsurface temperature variability and SST variability observed between January and April in Figure 6.2b is attributed to a decoupling of the surface mixed layer from the subsurface layer as climatological wind stress over the Atl3 region is at its weakest (Figure 6.3) and climatological atmospheric heating is at its strongest (Figure 5.1b).

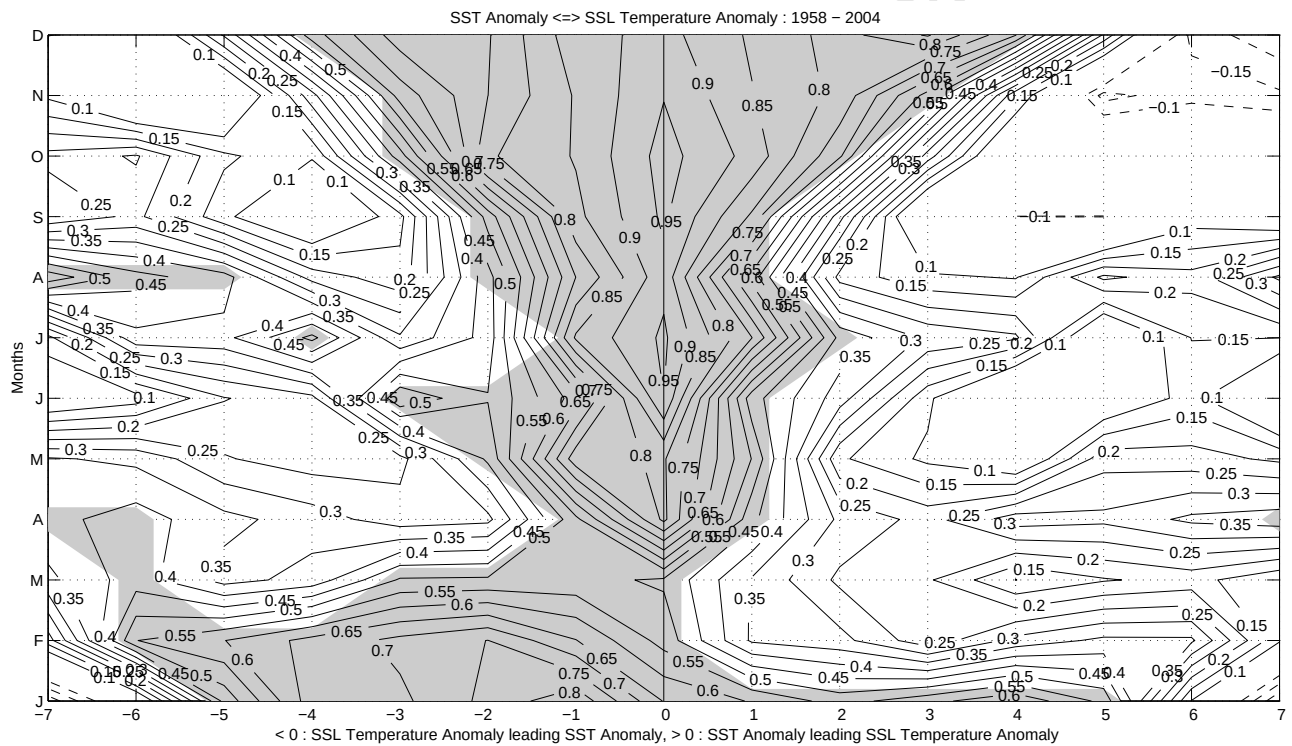
The strong relationship seen between inter-annual APE and SST anomalies between April and August is consistent with the fact that zonal mode events peak in boreal summer (Zebiak, 1993; Keenlyside and Latif, 2007). As shown in Section 5.2, seasonally these months coincide with the development of the cold tongue when subsurface cooling is the dominant process regulating Atl3 SST (Figure 5.1) and the translation of subsurface anomalies into surface anomalies is more effective (Figure 6.2b) due to enhanced wind stress (Figure 6.3) and equatorial upwelling (Figures 5.4, Latif and Grötzner, 2000; Peter et al., 2006; Keenlyside and Latif, 2007).

Furthermore, the strong relationship seen between inter-annual APE and SST anomalies during boreal summer is consistent with the observationally based findings of Keenlyside and Latif (2007) which suggest that the Bjerknes feedback is strongest in boreal summer. The seasonal dependence in the correlation between inter-annual Atl3 SST anomalies and APE anomalies proposes that anomalous Atl3 SST events forced by an anomalous Bjerknes feedback occur predominantly during boreal summer, while SST anomalies which occur outside of this season, in particular between January and April, are more likely to be the result of remote forcing or a different local air-sea interaction process. Analysis of the seasonality in the relationship between inter-annual variability in Atl3 SST, equatorial APE and zonal wind stress conducted in Chapter 3 based on SODA data, supports this notion that anomalous SST events in the Atl3 region forced by the Bjerknes feedback mechanism occur predominantly during boreal summer. Similar results are found here in the relationship between Atl3 SST, equatorial APE and zonal wind stress anomalies based on the ROMS-TAtl output.

Using ROMS-TAtl data from 1980-2004, Figure 6.4 explores the relationship between monthly anomalies in zonal wind stress and Atl3 SST, and Figure 6.5 the relationship between monthly anomalies in zonal wind stress and equatorial Atlantic APE. In Figure 6.4a, anomalous zonal wind stress to the west of the Atl3 region correlates best with anomalous Atl3 SST. This result is in agreement with the results obtained in Chapter 3 and previous literature on the zonal mode (Zebiak, 1993; Ruiz-Barradas et al., 2000; Keenlyside and Latif, 2007). Furthermore, the significant relationship between inter-annual anomalies in equatorial Atlantic APE and zonal wind stress anomalies in the western-central equatorial Atlantic depicted in Figure 6.5a supports the idea that the Bjerknes feedback mechanisms is the cause of the wind-SST relationship in



(a)



(b)

Figure 6.2: (a) A comparison between ROMS-TAtl, monthly, inter-annual anomalies in SST, the average temperature of the Surface Mixed Layer (SML) (varying depth) and the average temperature of the Sub-Surface Layer (SSL) (fixed depth = 70m) for the Atl3 region. The correlation between variables (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). (b) The seasonal dependence in the correlation between inter-annual anomalies in SST and the average temperature of the subsurface layer for the Atl3 region. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey.

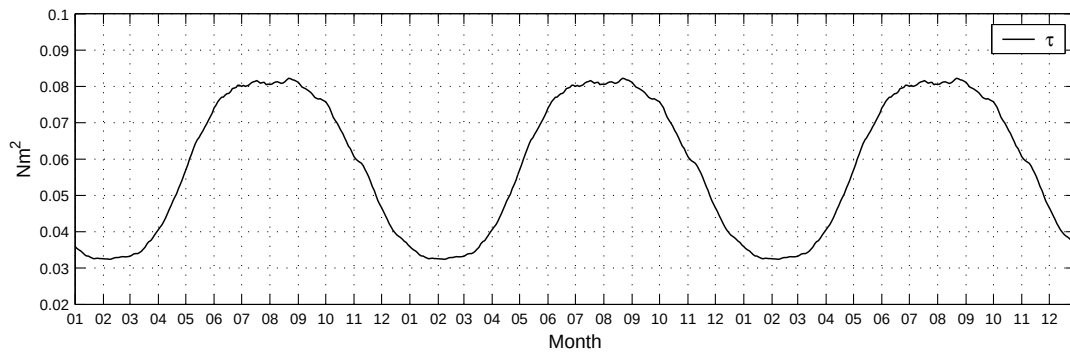


Figure 6.3: Climatological wind stress values for the Atl3 region. Climatological wind stress values were derived from the ROMS-TAtl simulation's wind stress forcing.

Figure 6.4a. Inter-annual Atl3 SST anomalies correlated with zonal wind stress anomalies to the west are the surface expressions of anomalous equatorial APE correlated with western-central Atlantic zonal wind stress anomalies.

The correlation analysis depicted in Figures 6.4a and 6.5a suggests that the Bjerknes feedback is associated with a significant amount of the observed inter-annual Atl3 variability. Figure 6.4b shows the seasonal dependence in the correlation between western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies and Atl3 SST anomalies and Figure 6.5b the seasonal dependence in the correlation between western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies and equatorial Atlantic APE anomalies. As seen in Chapter 3, a distinct seasonal dependence exists in the correlation between anomalous zonal wind stress and Atl3 SST (Figure 6.4b) as well as anomalous zonal wind stress and equatorial APE (Figure 6.5b) with the highest correlation between these variables observed during the boreal summer months. However, unlike the results obtained in Chapter 3 using SODA data, the relationship between anomalous western Atlantic zonal wind stress and Atl3 SST peaks in June rather than May. A similar discrepancy in the month during which the relationship between anomalous western Atlantic zonal wind stress and Atl3 SST is strongest was found by Keenlyside and Latif (2007) when using different observational datasets. Consistent with the observational results from Keenlyside and Latif (2007), the strongest relationship between anomalous western Atlantic zonal wind stress and thermocline depth anomalies (equatorial APE) is observed in June-July.

As seen in Keenlyside and Latif (2007) based on observational data, and in Chapter 3 based on SODA data, the correlation between anomalous Atl3 events and thermocline depth, as well as the correlation between anomalous Atl3 events and zonal wind anomalies, peaks in boreal summer. This consistent timing provides confidence in the ROMS-TAtl simulation.

Figure 6.6 shows the time lag-lead correlations between anomalous, (a) Atl3 SST and equatorial Atlantic APE, (b) Atl3 SST and western Atlantic zonal wind stress, and (c) western Atlantic zonal wind stress and equatorial Atlantic APE based on ROMS-TAtl data from 1980-2004. The lag-lead correlations shown in Figure 6.6 are all fairly symmetric about zero lag. These symmetric lag-lead correlations suggest that the covariability observed is the result of a positive feedback with anomalies in each variable reinforcing one another (Frankignoul and Hasselmann, 1977). Asymmetric or antisymmetric lag-lead correlations about zero lag would have suggested that they were the result of purely one-way forcing or a negative feedback (Frankignoul and Hasselmann, 1977). Figures 6.1b, 6.4b, and 6.5b illustrate the seasonal dependence in

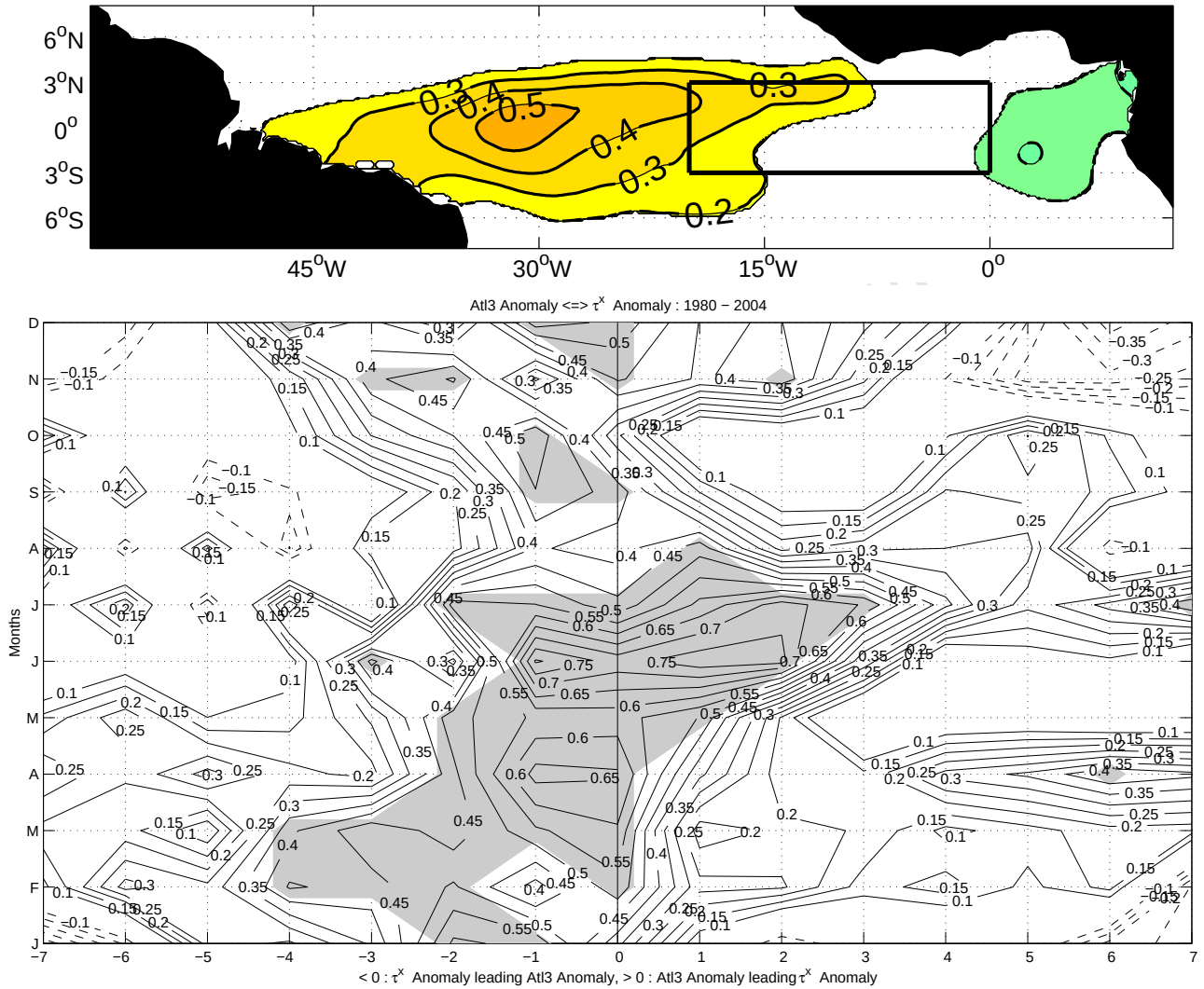


Figure 6.4: (a) A spatial map based on ROMS-TAtl data of the correlation between monthly anomalies in zonal wind stress values and AtI3 SST spanning 1980-2004. Correlation values plotted are those significance at the 95% level. (b) The seasonal dependence in correlation between ROMS-TAtl western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies and AtI3 SST anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey.

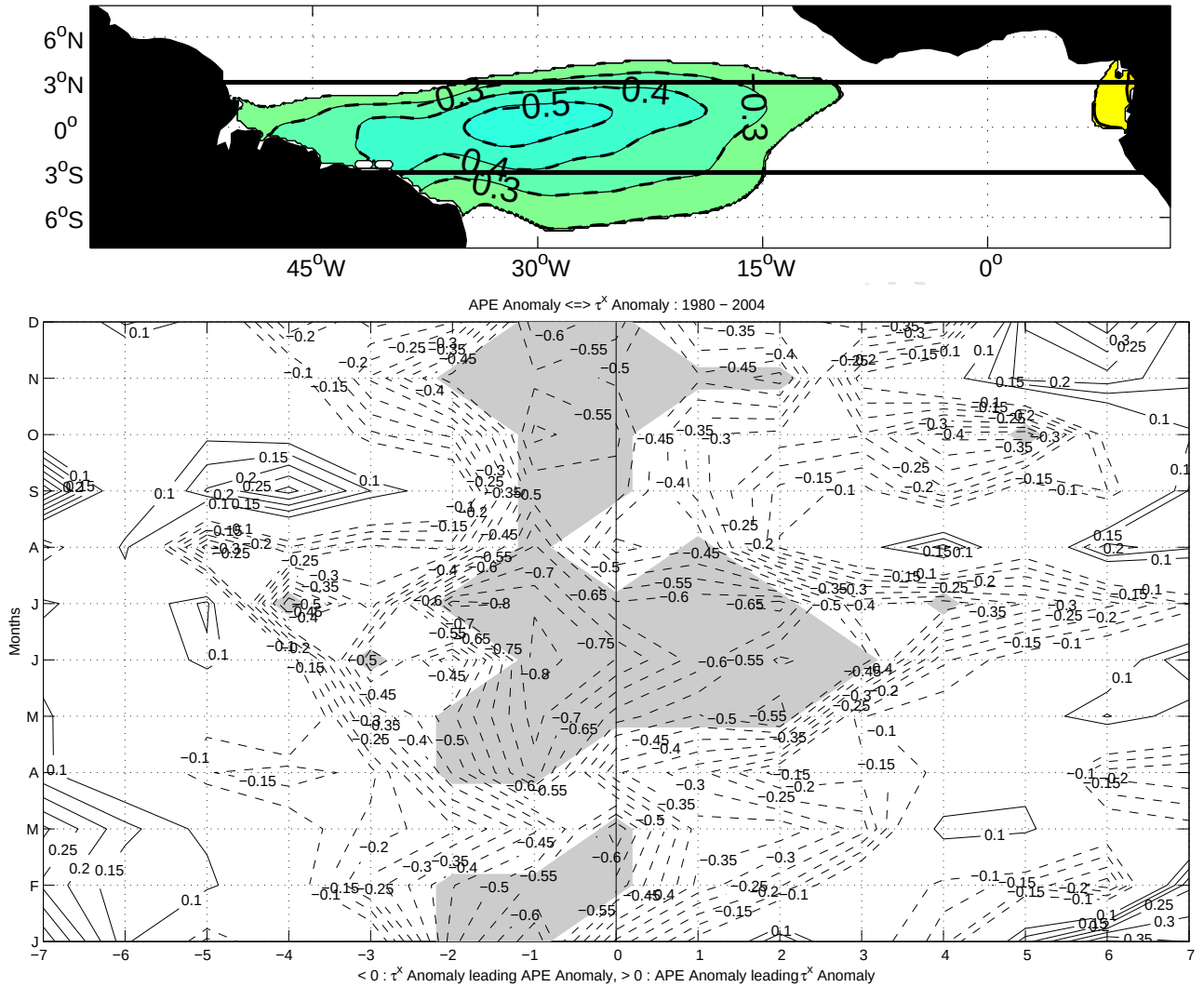
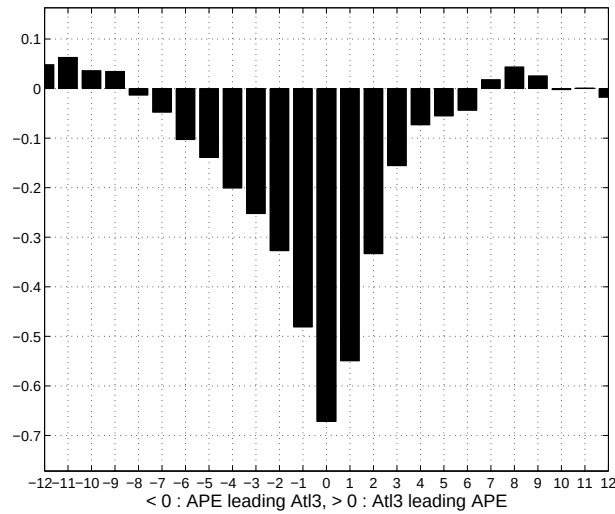
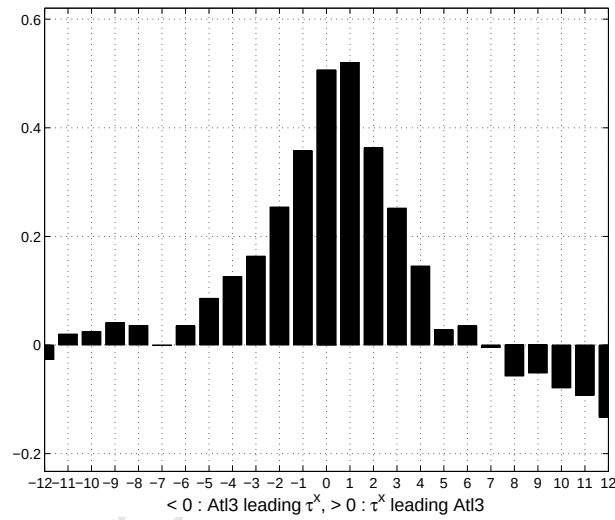


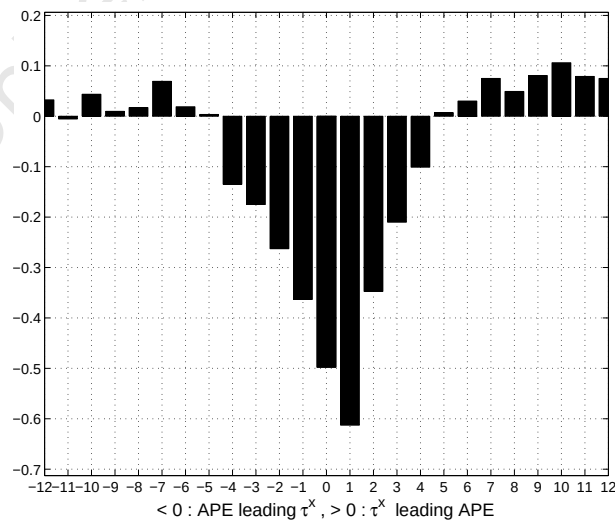
Figure 6.5: (a) A spatial map based on ROMS-TAtl data of the correlation between monthly anomalies in zonal wind stress values and equatorial APE (3°S-3°N 60°W-15°E 0-400m) spanning 1980-2004. Correlation values plotted are those significance at the 95% level. (b) The seasonal dependence in correlation between ROMS-TAtl western equatorial Atlantic (5°N-5°S 40°W-20°W) zonal wind stress anomalies and equatorial APE anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey.



(a)



(b)



(c)

Figure 6.6: Time lag correlations between anomalous (a) Atl3 SST and equatorial Atlantic APE (3°S - 3°N 60°W - 15°E 0-400m), (b) Atl3 SST and western equatorial Atlantic (5°N - 5°S 40°W - 20°W) zonal wind stress, and (c) equatorial Atlantic APE and western equatorial Atlantic zonal wind stress. SST, APE and zonal wind stress anomalies were derived from 1980-2004 ROMS-TAtl data. APE values were calculated following Equation 3.2.

correlations between Atl3 SST and equatorial Atlantic APE anomalies, zonal wind stress and Atl3 SST anomalies, and zonal wind stress and equatorial APE anomalies respectively. In these figures lag-lead correlations of the same sign are roughly symmetric about zero lag between April and August. These results are consistent with the findings of Keenlyside and Latif (2007) who suggest that the Bjerknes feedback is active inter-annually in the equatorial Atlantic and that seasonal variations in the influence of SSTs on surface winds, as well as seasonal variability in subsurface–surface coupling, point to the fact that the Bjerknes feedback is predominately active between April and August. Here however, in light of results presented in Chapters 3 and 5 suggesting that the Bjerknes is seasonally active between April and August, we suggest that anomalous SST events in the Atl3 region forced by the Bjerknes feedback mechanism occur predominantly during boreal summer as a modulation of an already seasonally active Bjerknes feedback.

As in Chapter 3, it is also worth noting the secondary peak between November and January in the correlations between Atl3 SST and equatorial Atlantic APE anomalies, anomalous zonal wind stress and Atl3 SST, and anomalous zonal wind stress and equatorial APE (Figures 6.1b, 6.4b and 6.5b). This secondary peak corresponds with that in inter-annual central equatorial Atlantic SST variability identified by Okumura and Xie (2006) as being associated with a November–December zonal mode of coupled variability.

6.3 Inter-annual Anomalies in Equatorial and Tropical Atlantic APE

In this section, the forcing behind inter-annual anomalies in equatorial and tropical Atlantic APE is explored. Following the methodology outlined in Section 5.3.2, the contribution of each term to the evolution of APE (Equation 5.29) has been estimated and then climatological values (as defined in Section 5.3) have been subtracted to establish inter-annual anomalies. Figure 6.7 shows the relationship between anomalous APE evolution and anomalous buoyancy power for, (a) the equatorial Atlantic (3°S–3°N 60°W–15°E 0–400m), and (b) the tropical Atlantic (8°S–8°N 60°W–15°E 0–400m). As observed inter-annually in the Pacific (Goddard and Philander, 2000), and seasonally in the Atlantic (Section 5.3), the dominant term driving inter-annual anomalies in equatorial and tropical Atlantic APE is seen to be the buoyancy power term (Figure 6.7). A correlation coefficient of $r=0.78$ ($r^*=0.19$) is seen between monthly anomalies in buoyancy power and APE evolution for the equatorial Atlantic and a correlation coefficient of $r=0.89$ ($r^*=0.19$) for the tropical Atlantic. The approximation that equatorial/tropical basin APE changes are driven primarily by buoyancy forcing associated with the horizontal redistribution of warm surface waters is therefore seen to hold when evaluating anomalous APE changes in the Atlantic (Equation 6.1)

$$\frac{dAPE_{ano}}{dt} \approx \underbrace{\iiint g\tilde{\rho}wdV - \iiint g\tilde{\rho}_{cl}w_{cl}dV}_{\Phi_{apk_{ano}}} \quad (6.1)$$

The buoyancy power term is the reversible exchange term between the APE and KE evolution equations and so, as in Section 5.3.3, further insight can be gained into the processes responsible for anomalous buoyancy power and hence anomalous APE evolution by considering the KE evolution equation (Equation 5.30).

$$\begin{aligned} \frac{dKE_{ano}}{dt} &= -\Phi_{ka_{ano}} - \Phi_{pw_{ano}} - \Phi_{apk_{ano}} + \Phi_{ww_{ano}} - \Phi_{ss_{ano}} \\ \Phi_{apk_{ano}} &= -\frac{dKE_{ano}}{dt} - \Phi_{ka_{ano}} - \Phi_{pw_{ano}} + \Phi_{ww_{ano}} - \Phi_{ss_{ano}} \end{aligned} \quad (6.2)$$

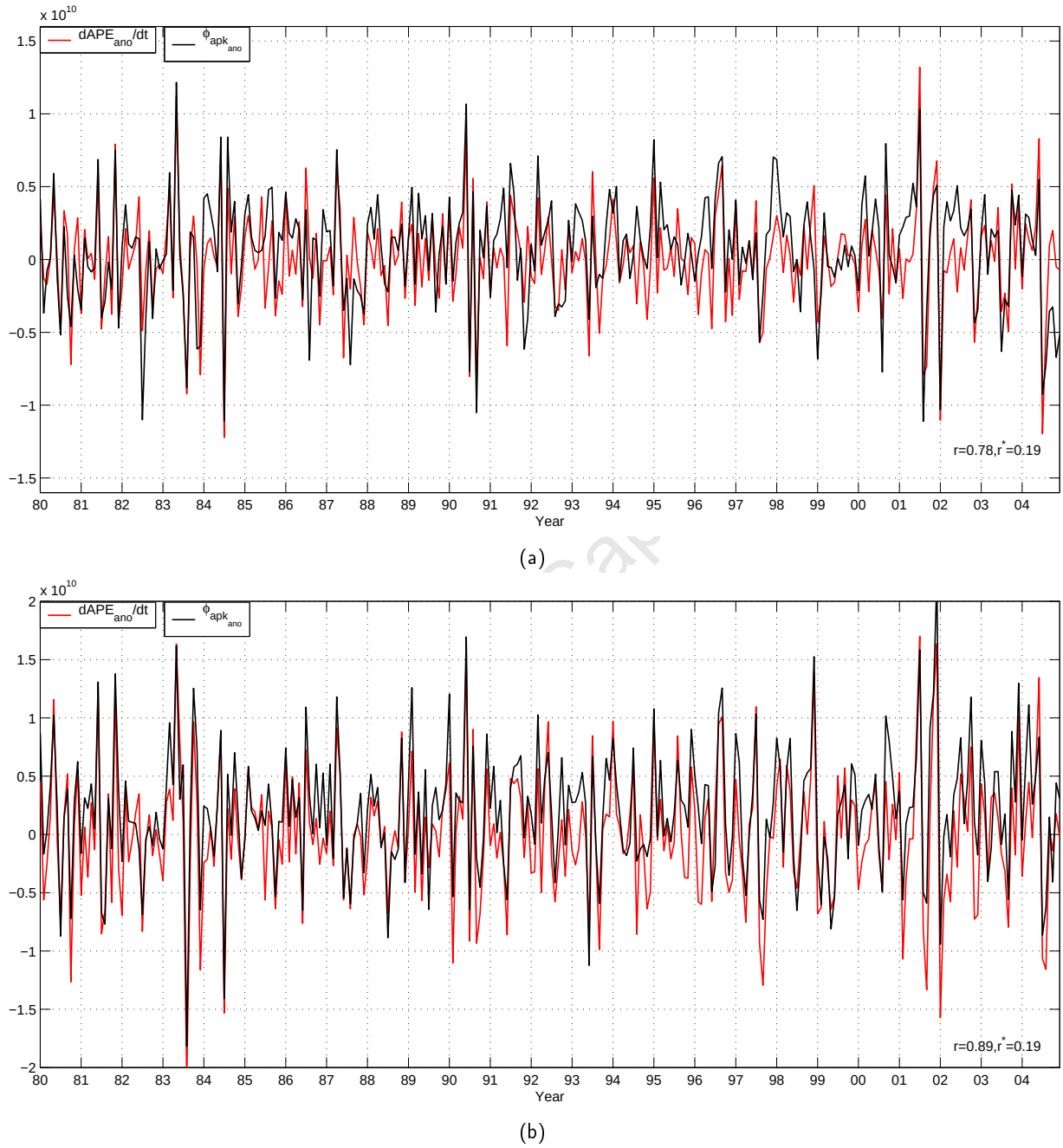


Figure 6.7: A comparison between monthly, inter-annual anomalies in the buoyancy power term ($\Phi_{apk_{ano}}$ in $J s^{-1}$), and APE evolution ($dAPE_{ano}/dt$ in $J s^{-1}$), assessed over (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*).

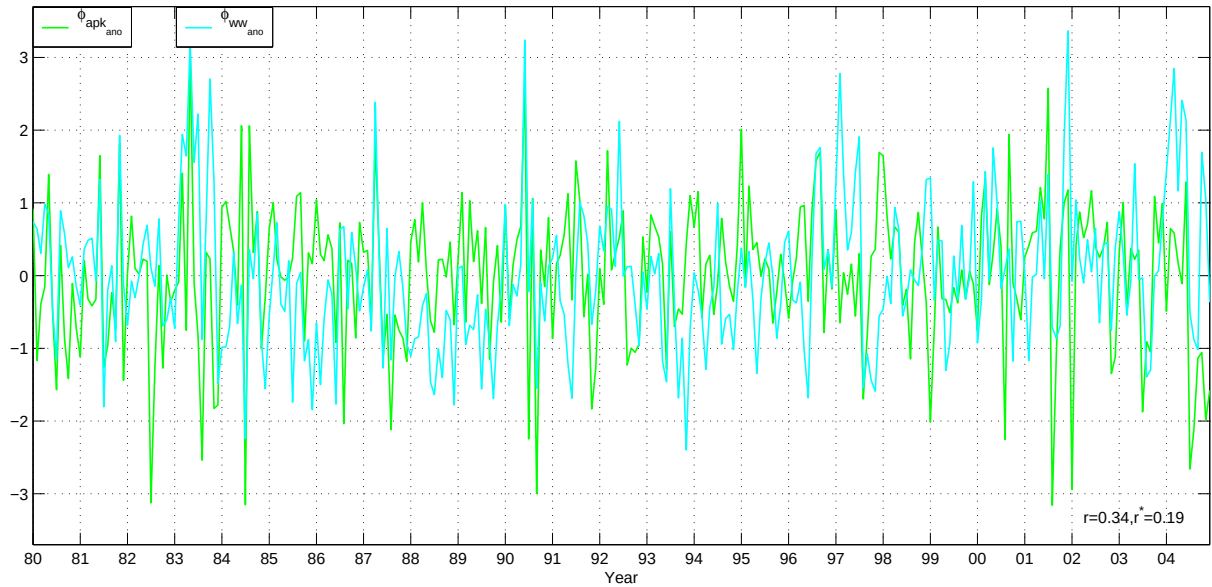
Following the methodology outlined in Section 5.3.2 the contribution of each term to the KE evolution equation (Equation 5.30) has been estimated and then climatological values (as defined in Section 5.3) have been subtracted to establish the terms driving KE evolution on inter-annual timescales. As seen inter-annually in the Pacific (Goddard and Philander, 2000), and seasonally in the Atlantic (Section 5.3), one would expect the horizontal redistribution of warm surface waters, which gives rise to anomalous buoyancy power and hence anomalous APE changes in the Atlantic, to be primarily related to fluctuations in wind power over the domain. In Figure 6.8, anomalous fluctuations in the buoyancy power term are compared with anomalous fluctuations in the wind power term for both the equatorial and tropical Atlantic domains. Like the results obtained in Section 5.3.3 with respect to the seasonal relationship between buoyancy power fluctuations and wind power changes within the equatorial and tropical domains (Figure 5.16), Figure 6.8 shows that anomalous buoyancy power corresponds closely with anomalous wind power (Figure 6.8b) within the tropical Atlantic domain but not within the equatorial domain (Figure 6.8a). A correlation coefficient of $r=0.34$ ($r^*=-0.19$) is found between monthly anomalies in buoyancy power and wind power for the equatorial Atlantic and a correlation coefficient of $r=0.67$ ($r^*=0.19$) for the tropical Atlantic. As in Section 5.3.3, this difference is thought to be due the larger influence of processes acting at the boundaries of the equatorial domain than the tropical domain, as transients (Rossby and Kelvin waves), excited by wind stress fluctuations, enter and exit the equatorial domain.

For the equatorial domain, changes in APE are driven not only by anomalous work done by the wind within the equatorial domain but also by anomalous processes acting at the boundaries. On the other hand, anomalous APE changes within the tropical domain are seen to be largely due to anomalous work done by the wind within the tropical domain. Extending the domain over which APE is evaluated polewards therefore simplifies the dynamics governing inter-annual APE variability down to a single dominant process, namely the work done by the wind over the tropical Atlantic.

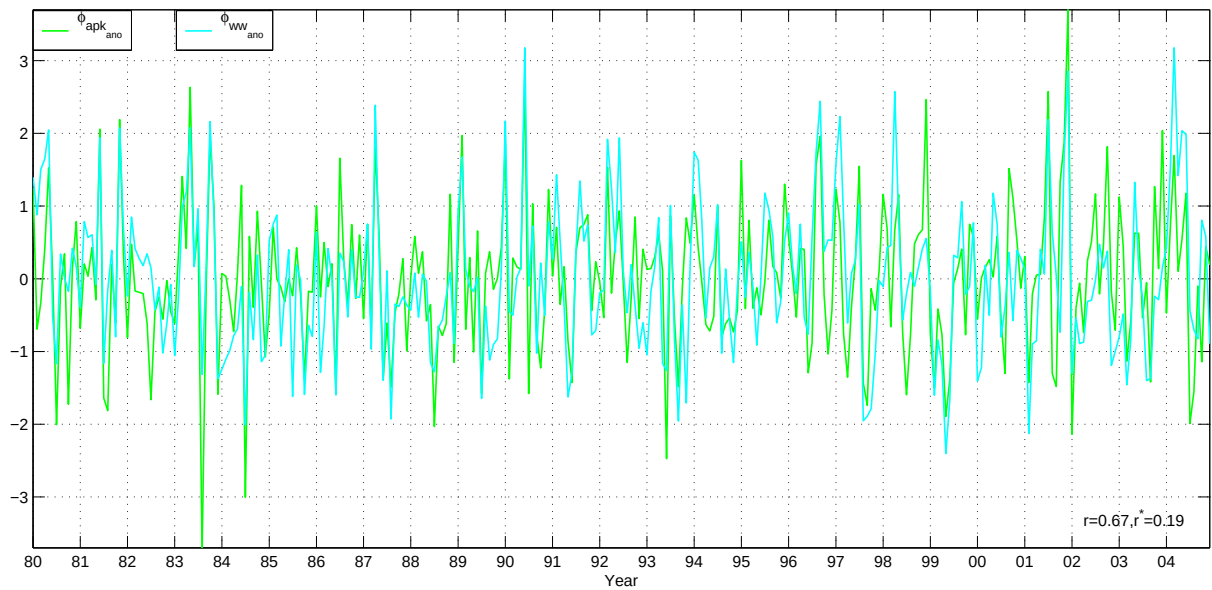
Figure 6.9 explores the implication of extending the domain over which APE is evaluated for the Atl3 SST-APE relationship. Although this meridional extension of the domain over which APE is evaluated allows for a meridional redistribution of mass within the domain, there is no obvious effect on the signature of inter-annual APE variability (Figure 6.9a). Inter-annual anomalies in tropical Atlantic APE are highly correlated with those in equatorial Atlantic APE ($r=0.86$ ($r^*=-0.19$)), with the maximum correlation occurring at zero lag (Figure 6.9b). Subsequently, the relationship between Atl3 SST anomalies and APE anomalies decreases only slightly (Figure 6.9c) when one extends the domain over which APE is evaluated to include the tropical domain. The maximum correlation between tropical Atlantic APE and Atl3 SST anomalies is observed at zero lag with a correlation coefficient of $r=-0.59$ ($r^*=-0.19$).

Figure 6.10 illustrates the relationship between anomalous APE evolution and anomalous wind power over the tropical Atlantic. A correlation coefficient of $r=0.66$ ($r^*=-0.19$) is seen between anomalous APE evolution and anomalous wind power. This result is consistent with the fact that anomalous APE evolution within the tropical domain is driven predominantly by buoyancy power anomalies (Figure 6.7b, Equation 6.1) and that the source of these buoyancy power anomalies in the KE energy evolution equation (Equation 6.2) is largely due to anomalous wind power (Figure 6.8b).

In Section 5.3 (Figure 5.19), zonal winds acting on zonal surface currents were found to be responsible for the majority of the work done seasonally by the wind on the ocean within the tropical Atlantic. The zonal component is also seen to be the dominant component within the anomalous wind power term. Figure 6.11a



(a)



(b)

Figure 6.8: A comparison between monthly inter-annual anomalies in the buoyancy power term ($\Phi_{apk_{ano}}$), and wind power ($\Phi_{ww_{ano}}$) term, for (a) the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m) and (b) the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). Anomalous buoyancy power ($\Phi_{apk_{ano}}$) and wind power ($\Phi_{ww_{ano}}$) values have been normalised.

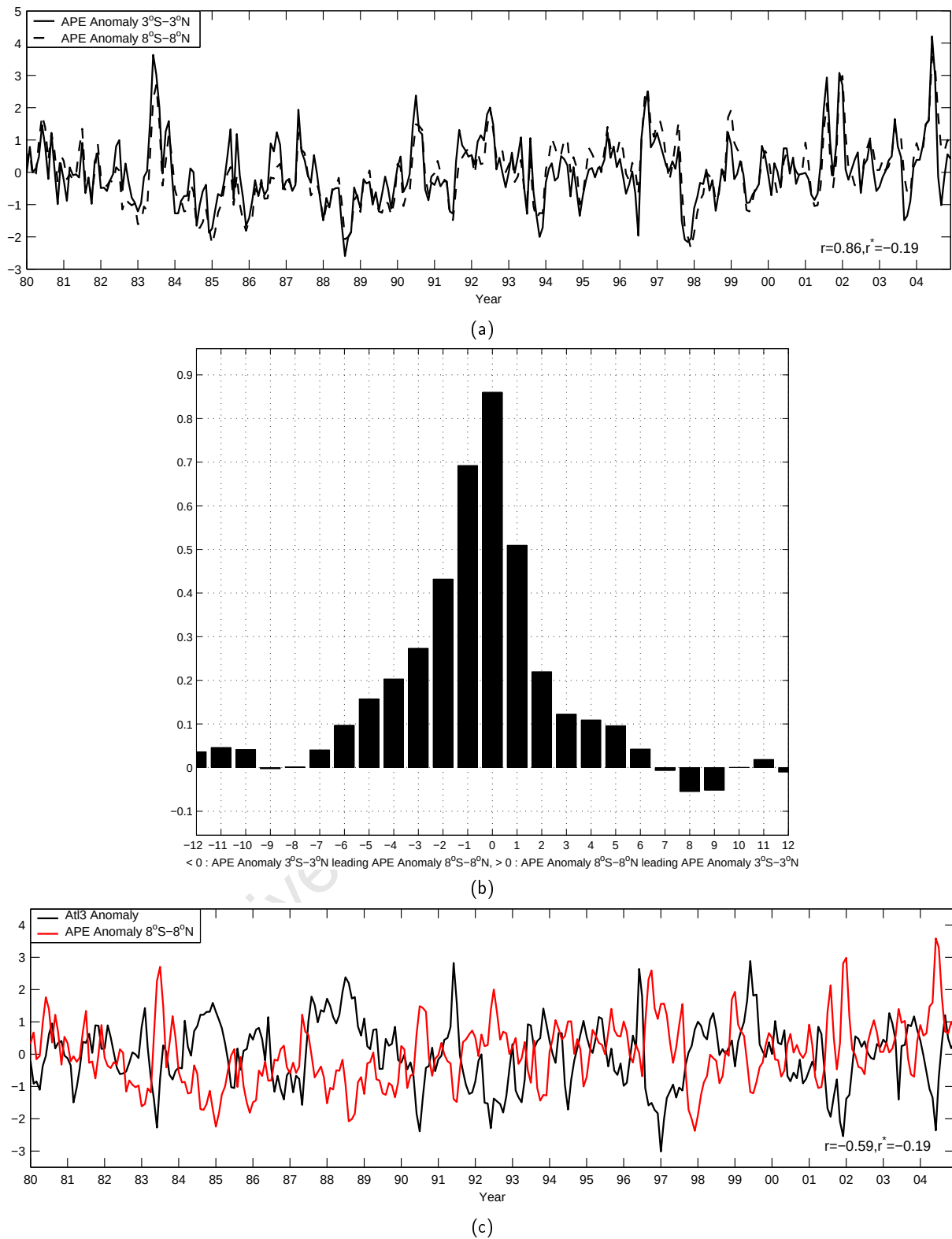


Figure 6.9: (a) A comparison between inter-annual equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE anomalies and tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE anomalies. (b) Time lag correlations between equatorial and tropical Atlantic APE anomalies. (c) A comparison between inter-annual Atl3 SST and tropical Atlantic (8°S-8°N 60°W-15°E 0-400m) APE anomalies. SST and APE anomalies were calculated based on ROMS-TAtl output. SST and APE values have been normalised. APE values were calculated following Equation 3.2. In a and c, the correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*).

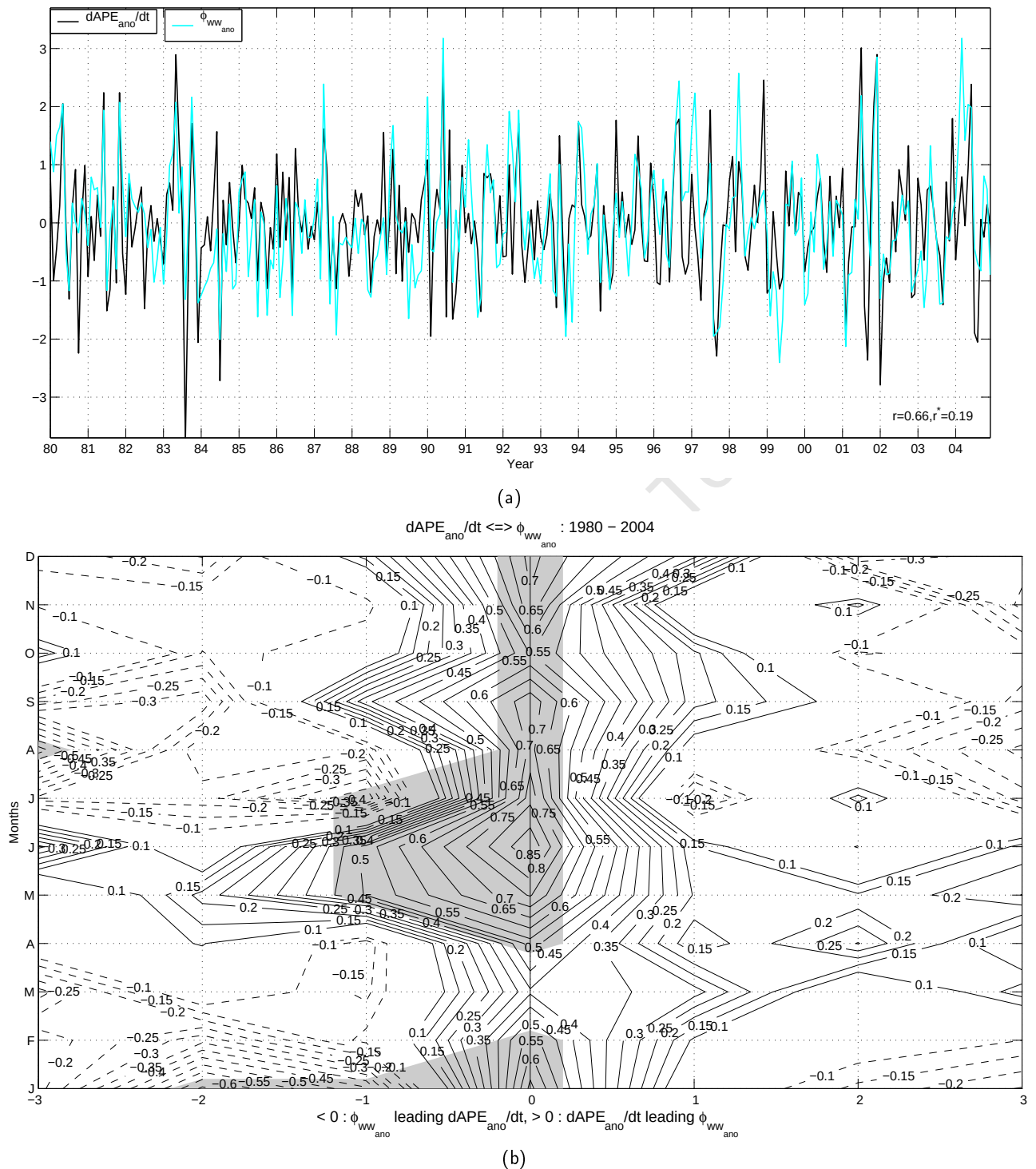


Figure 6.10: A comparison between monthly, inter-annual anomalies in APE evolution ($dAPE_{ano}/dt$) and wind power ($\Phi_{ww_{ano}}$) for the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). Anomalous APE evolution ($dAPE_{ano}/dt$) and wind power ($\Phi_{ww_{ano}}$) values have been normalised. (b) The seasonal dependence in correlation between monthly inter-annual anomalies in APE evolution ($dAPE_{ano}/dt$) and wind power ($\Phi_{ww_{ano}}$) for the tropical Atlantic domain (8°S-8°N 60°W-15°E 0-400m). Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey.

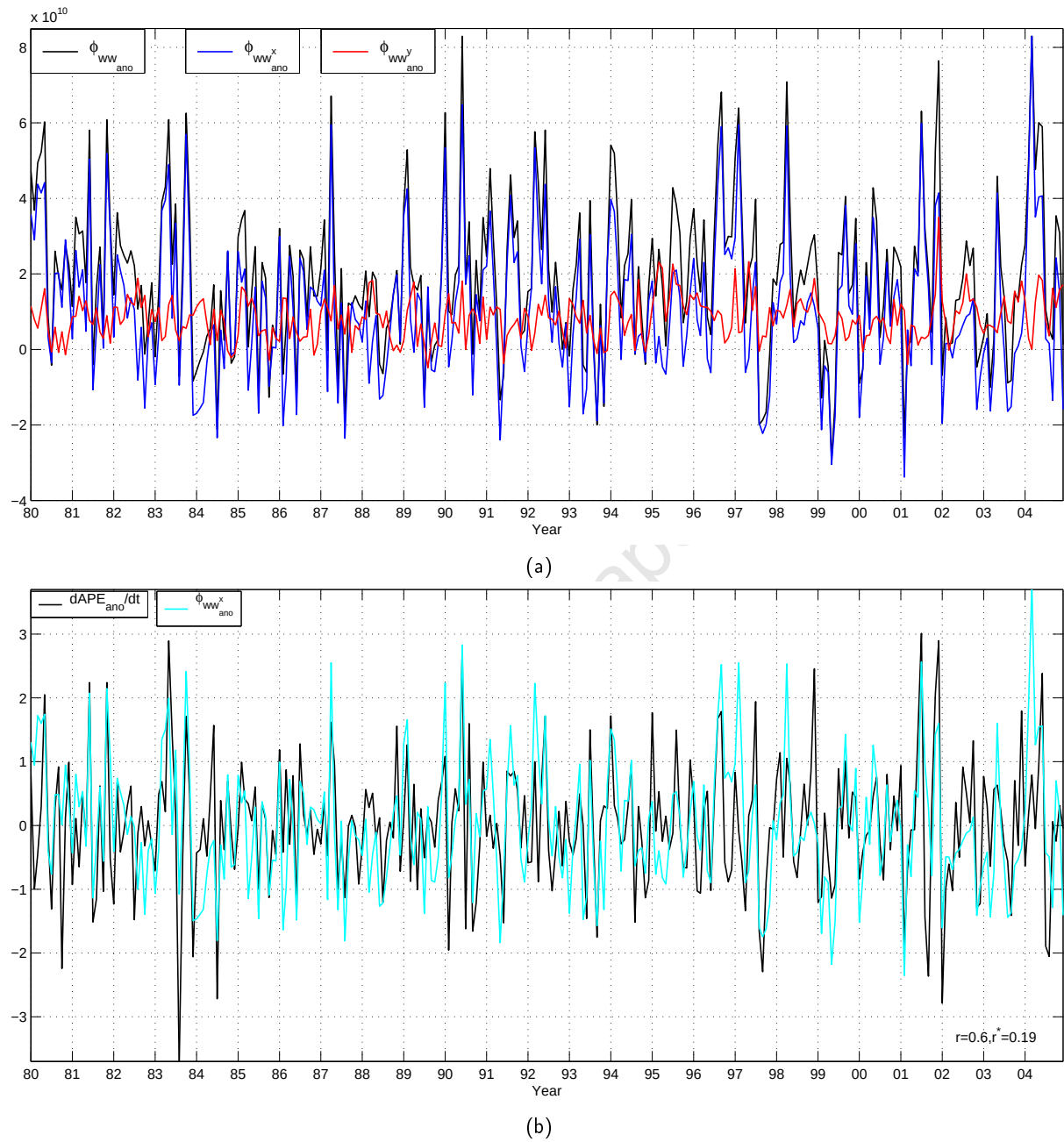


Figure 6.11: (a) A comparison between monthly, inter-annual anomalies in the wind power term ($\Phi_{ww_{ano}}$), and its zonal ($\Phi_{ww^x_{ano}}$) and meridional ($\Phi_{ww^y_{ano}}$) components for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). Values are given in Js^{-1} . (b) A comparison between monthly, inter-annual anomalies in the zonal component of the wind power term ($\Phi_{ww^x_{ano}}$), and APE evolution ($dAPE_{ano}/dt$) assessed over the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). Values have been normalised.

compares the relative contribution of the zonal ($\Phi_{ww_{ano}}^x = \iint_{z=0} u \tau_s^x dS - \iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS$) and meridional, ($\Phi_{ww_{ano}}^y = \iint_{z=0} v \tau_s^y dS - \iint_{z=0} v_{cl} \tau_{s_{cl}}^y dS$) components of anomalous wind power term to anomalous wind power values. Zonal wind stress acting on zonal surface currents is clearly responsible for the majority of the anomalous work done by the wind on the ocean within the tropical Atlantic, $\Phi_{ww_{ano}} \approx \Phi_{ww_{ano}}^x$. Figure 6.11b compares monthly anomalies in the zonal component of the wind power term with those in APE evolution for the tropical Atlantic domain. A correlation coefficient of $r=0.6$ ($r^*=-0.19$) is found between APE evolution and zonal wind power anomalies.

This analysis indicates that anomalous APE evolution within the tropical Atlantic is largely driven by anomalous wind power over the tropical domain (Figure 6.10a), particularly during boreal summer (Figure 6.10b). Inter-annual APE anomalies within the Atlantic are therefore predominantly forced by the same process as seasonal APE fluctuations: namely, by fluctuations in wind power. This result emphasises the fact that unlike in the Pacific where seasonal APE fluctuations are small with wind power fluctuations only forcing large APE fluctuations on inter-annual time scales, in the Atlantic, wind power fluctuations force large seasonal changes in APE with inter-annual fluctuations in wind power resulting in inter-annual APE anomalies. As illustrated in the following section, these inter-annual anomalies in tropical Atlantic wind power appear to be largely due to inter-annual modulations of the seasonal wind power changes associated with the thermocline mode seasonally excited between April and October.

6.4 Inter-annual modulation of the seasonally excited thermocline mode

The analysis conducted below suggests that the inter-annual variability in Atl3 SST which is related to anomalous APE can largely be regarded as inter-annual modulations of the seasonally excited Bjerknes feedback and its associated thermocline mode.

Considering the significant relationship between anomalous tropical Atlantic APE and Atl3 SST (Figure 6.9c) and the fact that anomalous tropical Atlantic APE is largely forced by tropical Atlantic wind power anomalies (Figure 6.10), one would expect anomalous wind power events to lead anomalous Atl3 SST events. Figure 6.12a shows the lag-lead relationship between anomalous tropical Atlantic wind power and anomalous Atl3 SST (note a 3 month running mean has been applied to each time series). Tropical Atlantic wind power anomalies are seen to lead Atl3 SST anomalies by 1-2 months (Figure 6.12a). A maximum anti-correlation of $r=-0.58$ is found between tropical Atlantic wind power and Atl3 SST when the former leads by 1 month.

Figure 6.12b explores whether there is any seasonal dependence in the lag-lead correlation between monthly Atl3 SST and tropical Atlantic wind power anomalies. The weakest correlation between wind power and Atl3 anomalies regardless of the lag or lead time occurs between January and April (Figure 6.12b). Consistent with the results obtained in Section 6.2, it is during these months that subsurface temperature anomalies associated with thermocline depth variability are less readily reflected in the SST field. The correlation between Atl3 SST anomalies and equatorial Atlantic APE anomalies is at its weakest then (Figure 6.1b) as Atl3 SST anomalies do not correspond closely with subsurface temperature anomalies (Figure 6.2b). Seasonally, the surface mixed layer is decoupled from subsurface variability during these months as the

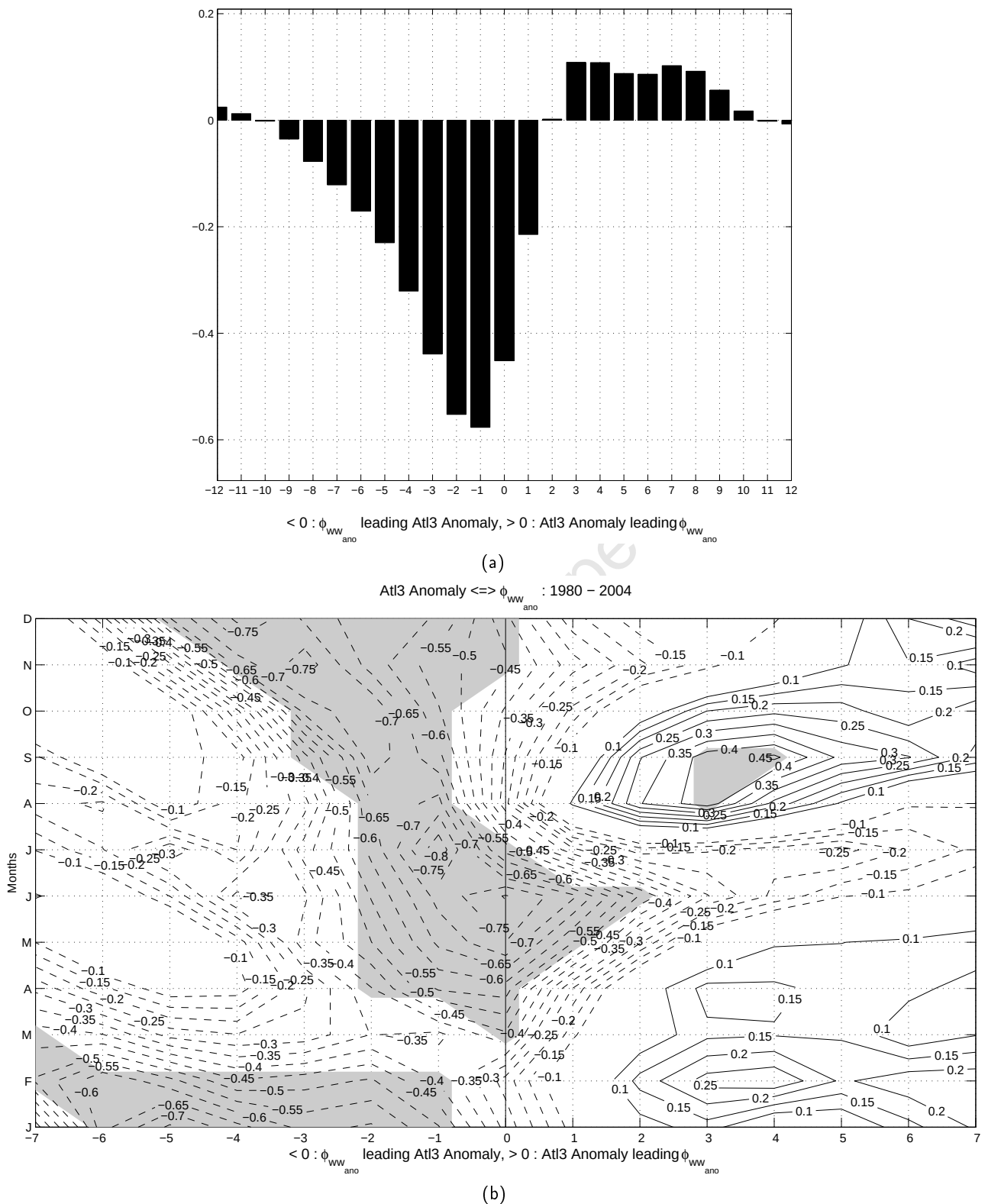


Figure 6.12: (a) Time lag correlations between inter-annual Atl3 SST anomalies and tropical Atlantic wind power anomalies (Φ_{wwano}). (b) The seasonal dependence in correlations between monthly inter-annual Atl3 SST anomalies and tropical Atlantic wind power anomalies (Φ_{wwano}). Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies and wind power anomalies were calculated based on ROMS-TATl output and a three month running mean has been applied to the time series.

climatological wind stress over the Atl3 region is at its weakest (Figure 6.3) and climatological atmospheric heating is at its strongest (Figure 5.1b).

Significant and roughly symmetric, lag-lead correlations between anomalous Atl3 SST and wind power are seen in Figure 6.12b between April and July with anti-correlation values reaching a maximum at zero lag ($r < -0.75$). This symmetric lag-lead relationship suggests that anomalies in each variable reinforce one another as opposed to purely a one-way forcing. As discussed in more depth below, anomalous wind power during these months is believed to be predominantly due to anomalous - Bjerknes feedback induced - zonal wind stress anomalies acting on climatological zonal surface currents. Atl3 SST and wind power anomalies typically reinforce one another between April and July, as an anomalous evolution of the Bjerknes feedback which is seasonally excited during these months, thus causing the growth of Atl3 SST anomalies (Chapter 3 and Section 6.2).

Between July and September, the highest correlation between Atl3 SST anomalies and wind power anomalies occurs when wind power anomalies lead by 1-2 months (Figure 6.12b). As elaborated on below, anomalous wind power during these months is seen to be predominantly due to climatological wind stress acting on anomalous zonal surface currents. Furthermore, a significant portion of the anomalous wind work occurring between August-September is positively correlated with Atl3 SST anomalies occurring 3-4 months earlier (Figure 6.12b). Climatological zonal wind stress acting on zonal surface current anomalies is responsible for this delayed opposing wind work.

From September to February, the lead correlation between wind power and Atl3 anomalies increases from 2 months in September to 5 months in January. The exact reason for this growth in the lag-lead relationship between wind work and Atl3 SST anomalies is uncertain. One suggestion could be the greater persistence of Atl3 SST anomalies during these months. Figure 6.13, which shows the seasonal dependence in Atl3 SST auto correlation, indicates that anomalous Atl3 condition generally persist between September and February. This result suggests that anomalous, central-eastern basin SST variability associated with the November-December zonal mode of Okumura and Xie (2006) is related to anomalous wind power forcing during the preceding August to October months.

As in Goddard and Philander (2000) wind power, approximated by its zonal component ($\Phi_{ww} \approx \iint_{z=0} u \tau_s^x dS = \Phi_{ww^x}$), can be decomposed into its climatological (as defined in Section 5.33, $\Phi_{ww_{cl}}^x = \iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS$), mean perturbation, ($\Phi_{ww_{ano}}^{mp} = \iint_{z=0} u' \tau_{s_{cl}}^x dS + \iint_{z=0} u_{cl} \tau_s^{x'} dS$) and perturbation ($\Phi_{ww_{ano}}^{pp} = \iint_{z=0} u' \tau_s^{x'} dS$) components.

$$\begin{aligned}
 \Phi_{ww^x} &= \iint_{z=0} u \tau_s^x dS \\
 &= \iint_{z=0} (u_{cl} + u') (\tau_{s_{cl}}^x + \tau_s^{x'}) dS \\
 &= \iint_{z=0} u_{cl} \tau_{s_{cl}}^x dS + \iint_{z=0} u' \tau_{s_{cl}}^x dS + \iint_{z=0} u_{cl} \tau_s^{x'} dS + \iint_{z=0} u' \tau_s^{x'} dS
 \end{aligned} \tag{6.3}$$

This decomposition is similar to the decomposition of the climatological zonal wind power term done in Section 5.3 except that here the effects of inter-annual zonal wind stress and zonal current perturbations from climatological values are examined. The mean perturbation component, $\Phi_{ww_{ano}}^{mp}$, comprises of two terms:

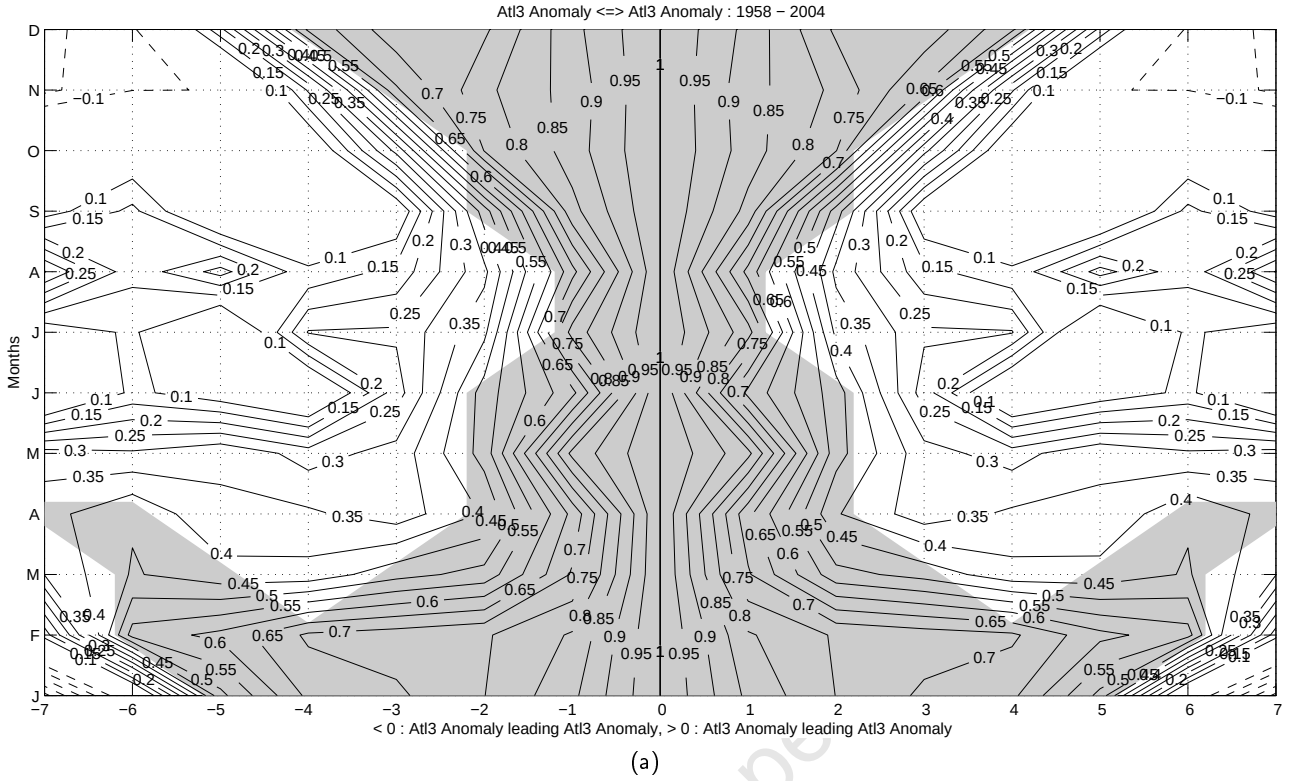


Figure 6.13: The seasonal dependence in monthly Atl3 SST auto correlation. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies were calculated based on ROMS-TAtl output.

$\Phi_{wwano}^{mup\tau} = \iint_{z=0} u_{cl} \tau_s^x dS$, which represents the effects of anomalous zonal wind stress fluctuations acting on climatological surface currents and $\Phi_{wwano}^{m\tau pu} = \iint_{z=0} u' \tau_{sc}^x dS$, representing the effects of climatological winds acting on anomalous surface current variations. τ_{sc}^x and u_{cl} represent the climatological zonal wind stress and surface current values, while τ_s^x and u' represent inter-annual perturbations from these climatological fields.

$$\begin{aligned}
 \Phi_{ww^x} &= \Phi_{ww_{cl}^x} + \Phi_{wwano}^{mup\tau} + \Phi_{wwano}^{m\tau pu} + \Phi_{wwano}^{pp} \\
 \Phi_{ww^x} - \Phi_{ww_{cl}^x} &= \Phi_{wwano}^{mup\tau} + \Phi_{wwano}^{m\tau pu} + \Phi_{wwano}^{pp} \\
 \Phi_{wwano}^x &= \Phi_{wwano}^{mup\tau} + \Phi_{wwano}^{m\tau pu} + \Phi_{wwano}^{pp}
 \end{aligned} \tag{6.4}$$

Following Equation 6.4, anomalous wind power approximated by its zonal component is decomposed into three components. In Figure 6.14, the relative contribution of the mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ as well as the perturbation term, Φ_{wwano}^{pp} , to wind power anomalies, Φ_{wwano}^x , is shown. Anomalous tropical Atlantic wind power is determined primarily by the mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ with fluctuations in the perturbation term, Φ_{wwano}^{pp} , playing a much smaller secondary role (Figure 6.14).

Figure 6.15a shows the lag-lead relationship between the mean perturbation term $\Phi_{wwano}^{mup\tau}$ and anomalous Atl3 SST (note a 3 month running mean has been applied to each time series). $\Phi_{wwano}^{mup\tau} = \iint_{z=0} u_{cl} \tau_s^x dS$ represents the contribution of anomalous zonal wind stress fluctuations acting on climatological surface currents. Fluctuations in the mean perturbation term $\Phi_{wwano}^{mup\tau}$ lead Atl3 SST anomalies by 1 month (Figure

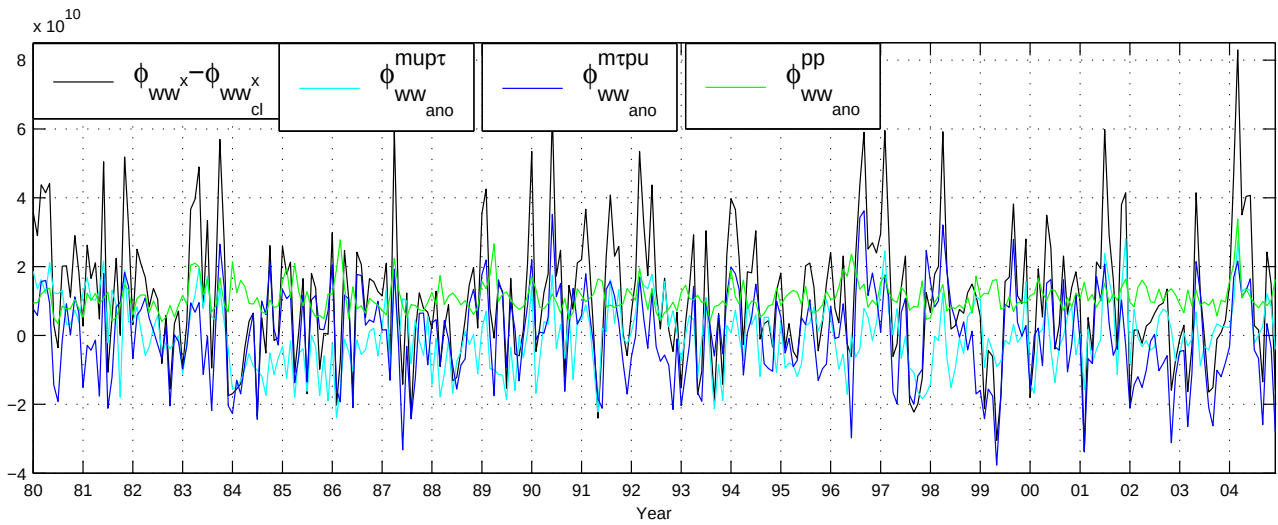


Figure 6.14: A comparison between monthly anomalies in the zonal component of the wind power term, Φ_{ww}^x ; the mean perturbation terms, $\Phi_{ww}^{mup\tau}$ and $\Phi_{ww}^{m\tau pu}$; and the perturbation term, Φ_{ww}^{pp} for the tropical Atlantic domain (8°S - 8°N 60°W - 15°E 0-400m). Values are given in Js^{-1} .

6.15a), with maximum anti-correlation of $r=-0.55$. Note that lag-lead correlations between $\Phi_{ww}^{mup\tau}$ and Atl3 SST (Figure 6.15a) are more symmetric about zero lag than those between anomalous wind power and Atl3 SST (Figure 6.12a). This more symmetric lag-lead relationship suggests that $\Phi_{ww}^{mup\tau}$ and anomalous Atl3 SST reinforce one another as opposed to purely a one-way forcing.

Figure 6.15b explores whether there is any seasonal dependence in the lag-lead correlation between monthly Atl3 SST anomalies and $\Phi_{ww}^{mup\tau}$. Correlation values between anomalous Atl3 SST and $\Phi_{ww}^{mup\tau}$ peak between April and July with anti-correlation values reaching a maximum at zero lag ($r < -0.75$). Lag-lead correlations are roughly symmetric during these months suggesting that anomalies in each variable reinforce one another as opposed to only one variable influencing the other. The seasonal dependence in the lag-lead correlation depicted in Figure 6.15b, largely resembles that of Figure 6.4b (Section 6.2) which depicts the seasonal dependence in the correlation between Atl3 SST anomalies and western equatorial Atlantic (5°N - 5°S 40°W - 20°W) zonal wind stress anomalies. As discussed in Section 6.2, seasonality in the occurrence of anomalous Atl3 events correlated with thermocline depth and zonal wind anomalies peaks between April and August. Furthermore, lag-lead correlations between these variables are roughly symmetric during these months suggesting that anomalous boreal summer Atl3 SST events are predominantly forced by an anomalous Bjerknes feedback. $\Phi_{ww}^{mup\tau}$ therefore represents the contribution of anomalous zonal wind stress, associated with inter-annual variability in the seasonally excited Bjerknes feedback, to wind power anomalies. Atl3 SST anomalies and $\Phi_{ww}^{mup\tau}$ typically reinforce one another between April and July as an anomalous evolution of the Bjerknes feedback that is seasonally excited during these months (Chapter 3 and Section 6.2).

Interestingly, the lag-lead relationship between Atl3 SST anomalies and $\Phi_{ww}^{mup\tau}$ from November to December is largely one sided despite the suggestion by Okumura and Xie (2006) that the Bjerknes feedback is associated with the generation of November-December zonal mode events. Figure 6.15b suggests that the anomalous, central-eastern basin SST variability associated with the November-December zonal mode of Okumura and Xie (2006) is related to wind work anomalies driven by anomalous zonal wind stress acting on climatological surfaced currents during the previous August to September.

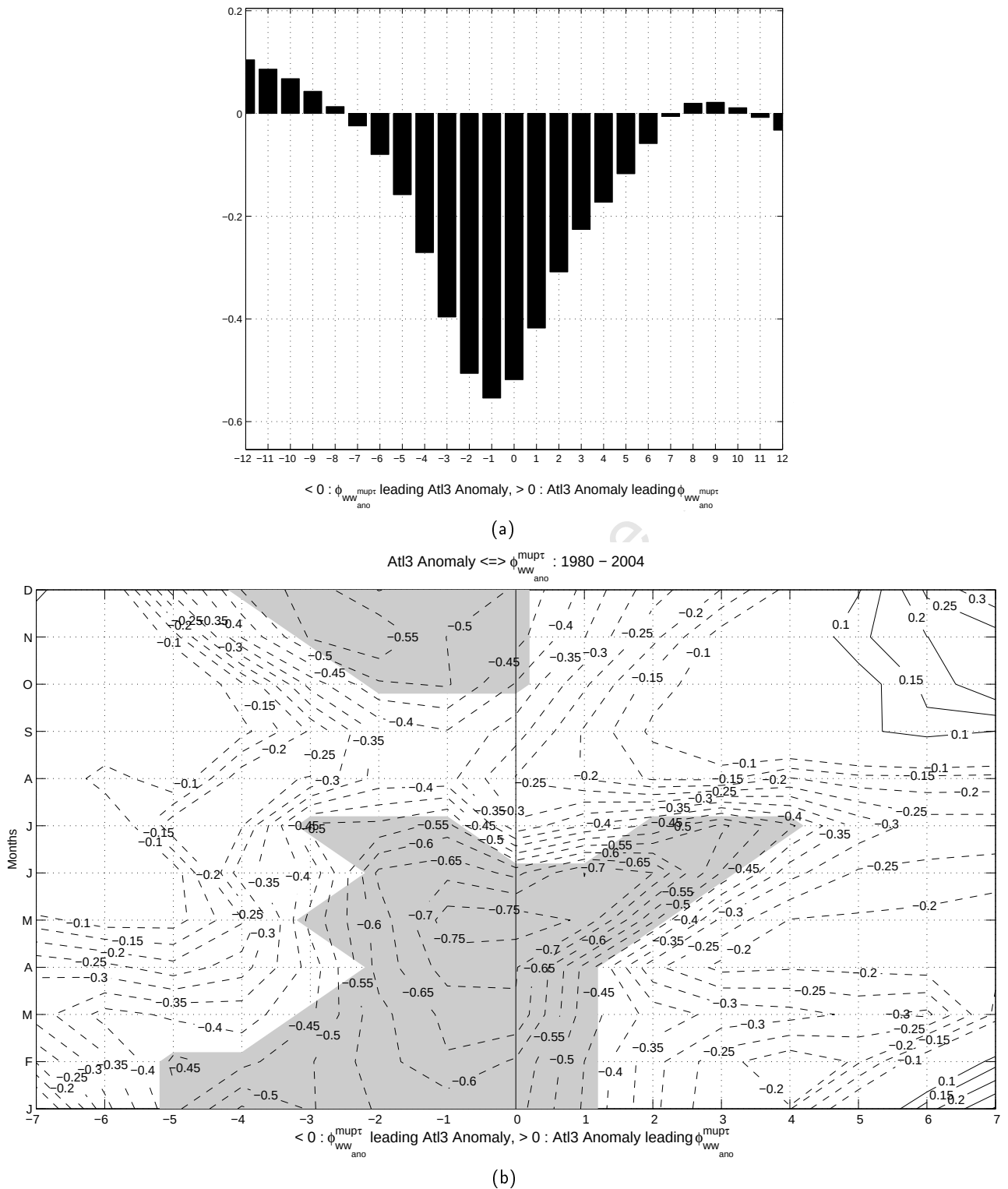


Figure 6.15: (a) Time lag correlations between inter-annual Atl3 SST anomalies and the mean perturbation term Φ_{ww}^{mupr} . (b) The seasonal dependence in correlations between monthly inter-annual Atl3 SST anomalies and the mean perturbation term Φ_{ww}^{mupr} . Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies and Φ_{ww}^{mupr} were calculated based on ROMS-TAtl output and a three month running mean has been applied to the time series.

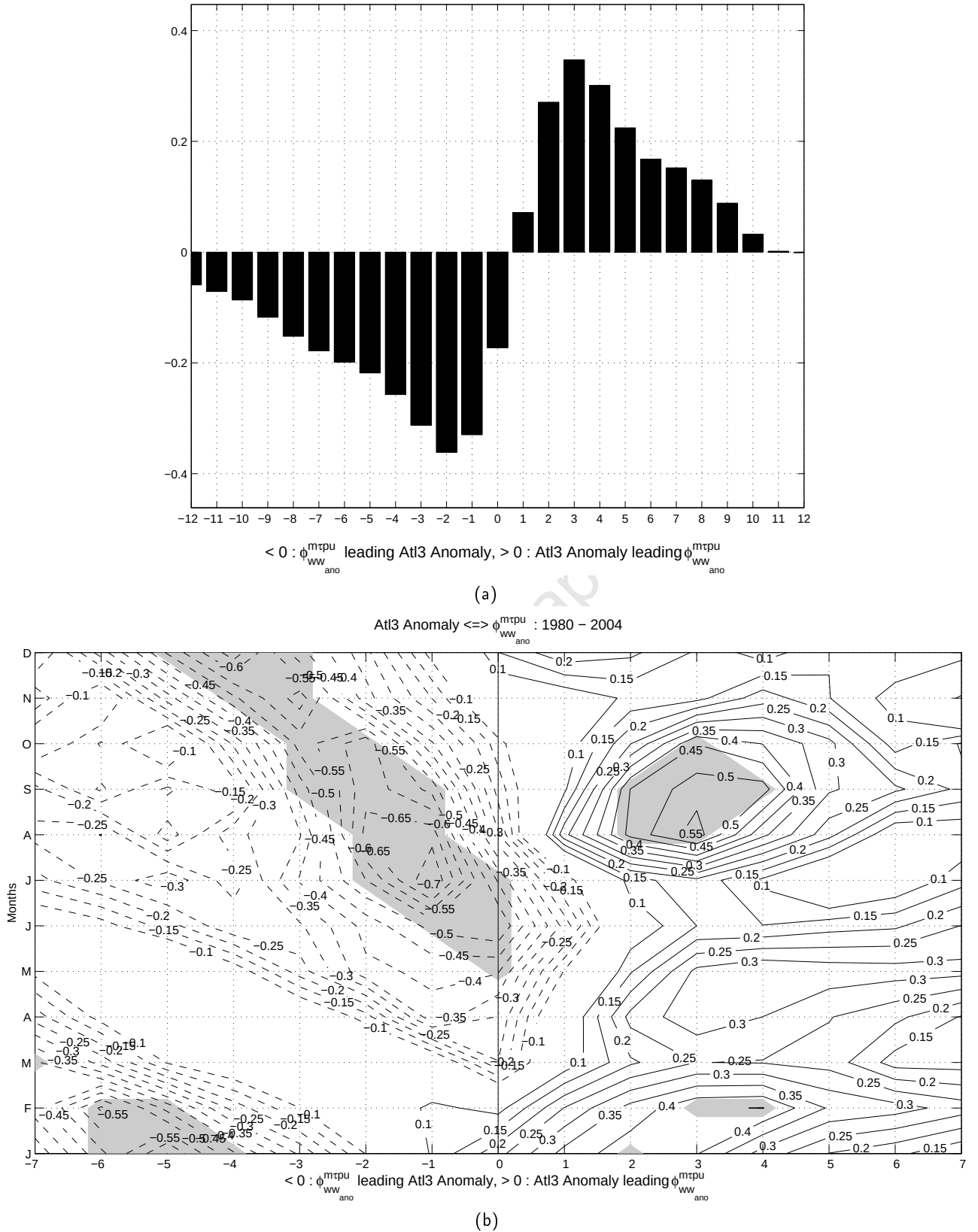


Figure 6.16: (a) Time lag correlations between inter-annual Atl3 SST anomalies and the mean perturbation term $\Phi_{ww}^{m\tau pu}$. (b) The seasonal dependence in correlations between monthly inter-annual Atl3 SST anomalies and the mean perturbation term $\Phi_{ww}^{m\tau pu}$. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. SST anomalies and $\Phi_{ww}^{m\tau pu}$ were calculated based on ROMS-TATl output and a three month running mean has been applied to the time series.

Figure 6.16a shows the lag-lead relationship between the mean perturbation term $\Phi_{wwano}^{m\tau pu}$ and anomalous Atl3 SST (note a 3 month running mean has been applied to each time series). $\Phi_{wwano}^{m\tau pu} = \iint_{z=0} u' \tau_{sc}^x dS$ represents the contribution of climatological zonal wind stress fluctuations acting on anomalous surface currents. Fluctuations in this term lead Atl3 SST anomalies by 2 months with maximum anti-correlation of $r=-0.36$ (Figure 6.16a). Unlike the relationship between $\Phi_{wwano}^{m\tau pt}$ and Atl3 SST seen in Figure 6.15a, a positive correlation of 0.35 between $\Phi_{wwano}^{m\tau pu}$ and Atl3 SST occurs when $\Phi_{wwano}^{m\tau pu}$ lags Atl3 SST anomalies by 3 months (Figure 6.16a). As seen seasonally in the Atlantic (Section 5.3.3) and inter-annually in the Pacific (Goddard and Philander, 2000), $\Phi_{wwano}^{m\tau pu}$ ($\Phi_{wwcl}^{m\tau pu}$ in the case of the seasonal cycle) represents the effect of transient induced changes in surface currents on the ability of the wind to do work on the ocean. From an energetics perspective, this term captures the effects of the delayed oscillator's negative, ocean memory feedback mechanism in the Pacific (Goddard and Philander, 2000). On inter-annual time scales in the Pacific, surface current anomalies driven by the thermocline perturbations associated with off-equatorial Rossby waves arriving in the western equatorial Pacific affect the ability of the wind to do work on the ocean (Goddard and Philander, 2000). This ocean memory process, in accordance with the delayed oscillator theory, is seen to be responsible for the transition from La Niña to El Niño or vice versa. Similarly, results obtained in Section 5.3.3 suggested that seasonal surface current changes - associated with transients excited by the abrupt seasonal change in zonal wind stress which occurs in May-June (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009) - effect the ability of the wind to do work on the ocean. Seasonally excited Kelvin and Rossby wave propagation results in accelerated westward surface flow in June-July followed by a decelerated westward surface flow between August-October and the subsequent acceleration in November-December (Philander and Pacanowski, 1986a; Schouten et al., 2005; Ding et al., 2009). The anti-correlation of $r=-0.36$ found between $\Phi_{wwano}^{m\tau pu}$ and Atl3 SST when $\Phi_{wwano}^{m\tau pu}$ leads by 2 months is thought to be the result of the westward equatorial surface current acceleration (deceleration) that accompanies the anomalous Bjerknes feedback associated with cold (warm) Atl3 SST events. On the other hand, the positive (negative) contribution of $\Phi_{wwano}^{m\tau pu}$ to anomalous wind work ~ 3 months after a warm (cold) event represents the role of the delayed, negative, ocean memory feedback mechanism. This delayed, negative, feedback mechanism operates on much shorter time scales than in the Pacific, where according to Goddard and Philander (2000) $\Phi_{wwano}^{m\tau pu}$ contributes positively (negatively) to anomalous wind work 9-12 months after a warm (cold) event.

Figure 6.16b investigates whether there is any seasonal dependence in the lag-lead correlation between monthly inter-annual Atl3 SST anomalies and $\Phi_{wwano}^{m\tau pu}$. Anti-correlation values between anomalous Atl3 SST and $\Phi_{wwano}^{m\tau pu}$ peak in July/August with anti-correlation values reaching a maximum ($r < -0.7$) when $\Phi_{wwano}^{m\tau pu}$ leads by one month. From September to February the lead grows from 2 months in September to 5 months in January. As mentioned earlier, a possible cause of this growth in the lag-lead relationship between wind work and Atl3 SST anomalies is thought to be the greater persistence in Atl3 SST anomalies seen during these months (Figure 6.13). The consistent growth in the lag-lead relationship between wind work and Atl3 SST from August to January (Figure 6.12b) appears to be primarily associated with the $\Phi_{wwano}^{m\tau pu}$ component of anomalous wind power (Figure 6.16b) and to a lesser extent with $\Phi_{wwano}^{m\tau pt}$ (Figure 6.15b). This result reinforces the hypothesis that anomalous conditions in December/January are largely associated with a persistence of anomalous Atl3 conditions originating in August/September as the $\Phi_{wwano}^{m\tau pu}$ (Figure 6.16b) rather than the $\Phi_{wwano}^{m\tau pt}$ (Figure 6.15b) component of anomalous wind power is more strongly correlated with anomalous Atl3 conditions in August/September.

Evident in Figure 6.16b between August and October are the significant positive correlations between $\Phi_{ww_{ano}}^{m\tau pu}$ and Atl3 SST anomalies of >0.45 when Atl3 SST leads by 2-4 months. This feature, also evident in Figure 6.12b, is only associated with climatological zonal wind stress fluctuations acting on anomalous surface currents ($\Phi_{ww_{ano}}^{m\tau pu}$ - Figure 6.16b) and not with anomalous zonal wind stress fluctuations acting on climatological surface currents ($\Phi_{ww_{ano}}^{mup\tau}$ - Figure 6.15b). Like the positive correlation of 0.35 between $\Phi_{ww_{ano}}^{m\tau pu}$ and Atl3 SST shown in Figure 6.16a when $\Phi_{ww_{ano}}^{m\tau pu}$ lags Atl3 SST anomalies by 3 months, these positive correlations represent the effects of the ocean memory, negative feedback mechanism. These significant positive correlations between $\Phi_{ww_{ano}}^{m\tau pu}$ and Atl3 SST (6.16b) coincide with the time of the year when the seasonally active ocean memory mechanism associated with the deceleration of surface currents acts to decay the seasonally excited Bjerknes feedback mechanism (Figure 5.20) as part of the seasonally active thermocline mode (Figure 5.27). They therefore offer further evidence in support of the hypothesis that anomalous conditions associated with the zonal mode in the tropical Atlantic are typically the result of a modulation of the seasonally active thermocline mode. While an amplified (suppressed) seasonally excited Bjerknes feedback between April-June leads to the growth of anomalous conditions, the seasonally active decay mechanism associated with the delayed deceleration of surface currents between August-October is also amplified (suppressed), hereby leading to the decay of anomalous conditions.

6.4.1 Composite Analysis and Individual Events

To establish a clearer picture of the typical processes associated with warm and cold zonal mode events, composites of years when cold/warm Atl3 SST events occurred during boreal summer months (June-August) have been created. Here, warm/cold event years are defined as years in which ROMS-TAtl Atl3 SST anomalies exceed one standard deviation (0.5°C) between June and August of that year, and correspond well with observed Hadley OI Atl3 SST. Figure 6.17 shows when monthly, ROMS-TAtl Atl3 SST anomalies exceed/fall below one standard deviation. Plotted in black in Figure 6.17 are the periods which are excluded due to poor correspondence with observed Hadley OI SST: namely, January 1981 - April 1983, September 1984 - August 1985, May 1990 - September 1990, August 1996 - August 1997 and June 2001 - April 2002. Within the remaining periods, cold event years have been identified as 1983, 1986, 1992, 1994 and 2004 and warm event years as 1984, 1987, 1988, 1991, 1995, 1996, 1997, 1998 and 1999. The 1984, 1987, 1988, 1993, 1995 and the 1998 warm events have been investigated in previous studies based on observed and model data (Carton and Huang, 1994; Vauclair and du Penhoat, 2001; Vauclair et al., 2004).

The 1983, 1986, 1992, 1994 and 2004 cold event years have been composited to gain insight into the typical evolution of pertinent variables such as APE and wind power during years which have experienced boreal summer cold events. Figure 6.18 shows the cold event year composite of (a) anomalous Atl3 SST, equatorial Atlantic APE, tropical Atlantic APE, and tropical Atlantic BPE; (b) total versus climatological Atl3 SST, tropical Atlantic APE, and tropical Atlantic BPE values; (c) total versus climatological values for wind work (Φ_{ww}) and advection of the density field (Φ_{bpa}); and (d) the relative contribution of mean perturbation terms $\Phi_{ww_{ano}}^{mup\tau}$ and $\Phi_{ww_{ano}}^{m\tau pu}$ to anomalous wind power. Typical of the cold events are anomalously high equatorial and tropical APE values peaking in June/July (Figure 6.18a). Considering these anomalous conditions in the context of total versus climatological conditions, Figure 6.18b shows that both APE and Atl3 SST anomalies are typically the results of an amplified and early onset of SST and APE changes driven by the seasonal development of the equatorial cold tongue and its associated Bjerknes feedback.

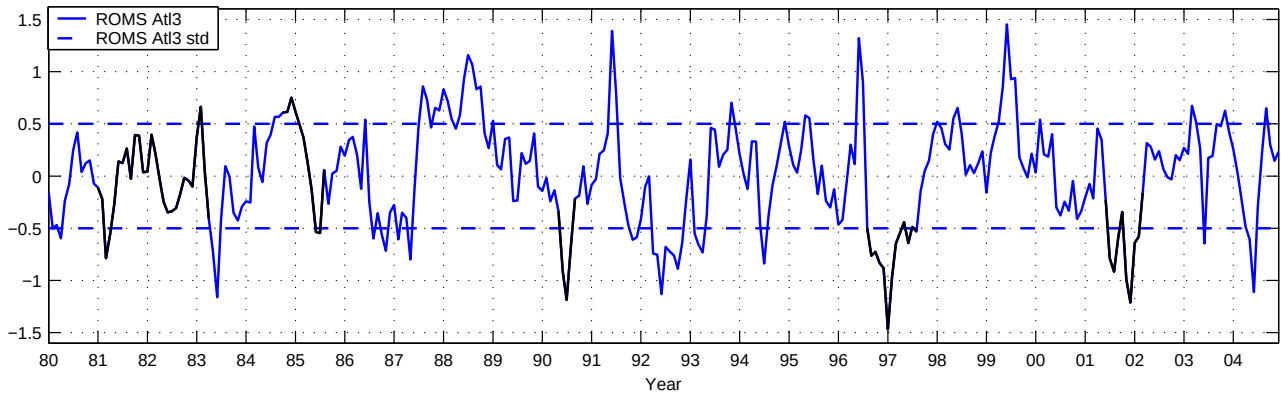


Figure 6.17: Monthly anomalies in ROMS-TATl Atl3 SST plotted in relation to one standard deviation. Periods where simulated Atl3 SST anomalies correspond poorly with observed Hadley OI Atl3 SST anomalies are plotted in black.

Anomalous high APE in June/July is seen in Figures 6.18c and 6.18d to be the result of anomalously strong wind power during the preceding months (February-July). The relative contribution of mean perturbation terms $\Phi_{ww_{ano}}^{mup\tau}$ and $\Phi_{ww_{ano}}^{m\tau pu}$ to anomalous wind power depicted in Figure 6.18d indicates that the anomalously strong wind power between February and August is due to the combined effect of anomalously strong wind stress acting on climatological surface currents ($\Phi_{ww_{ano}}^{mup\tau}$) and climatological wind stress acting on anomalously strong westward surface currents ($\Phi_{ww_{ano}}^{m\tau pu}$). Although $\Phi_{ww_{ano}}^{m\tau pu}$, representing the effect of climatological winds acting on anomalous surface currents, initially reinforces $\Phi_{ww_{ano}}^{mup\tau}$ between February and July, it reverses in August. This reversal results in anomalously weak wind power forcing between August and September that acts to decay the APE anomalies.

Figure 6.19 illustrates the typical spatial structure and temporal evolution of the anomalous wind power field during cold events in comparison with the spatial structure and temporal evolution of climatological wind power anomalies from the annual mean. Between March and July (Figures 6.19a, 6.19c, 6.19e, 6.19g and 6.19i), anomalously strong wind power values are observed over the 8°S-2°N and 50°W-0°E region. The anomalously strong wind power within this region corresponds with the anomalously strong domain integrated wind power seen during this period in Figures 6.18c and 6.18d. The spatial structure and temporal evolution of the positive wind power anomalies seen within this region (Figures 6.19e, 6.19g and 6.19i) are similar to that associated with the seasonal cycle (Figures 6.19f, 6.19h and 6.19j). The off-equatorial Rossby wave influence is evident as regions of enhanced wind power are situated either side of the equator and propagate westward. Likewise, the spatial structure associated with the subsequent negative wind power anomaly within this region, that acts to decrease APE and decay the cold event between July and October (Figures 6.18c and 6.18d), resembles that associated with the seasonal cycle during these months.

In Section 5.3.3 the behaviour of the seasonally excited thermocline mode was summarised by a phase diagram between seasonal APE values and the wind power term ($\Phi_{ww_{cl}}$). Figure 6.20 compares the cold year composite, APE versus wind power phase diagram with the seasonal, APE versus wind power phase diagram shown in Section 5.3.3 (Figure 5.27). Figure 6.20 clearly expresses how anomalously strong wind power during the early months of the year typically leads to an amplification of the seasonally excited thermocline mode as both the seasonally excited Bjerknes and the delayed, negative, ocean memory feedback mechanisms are amplified (6.18).

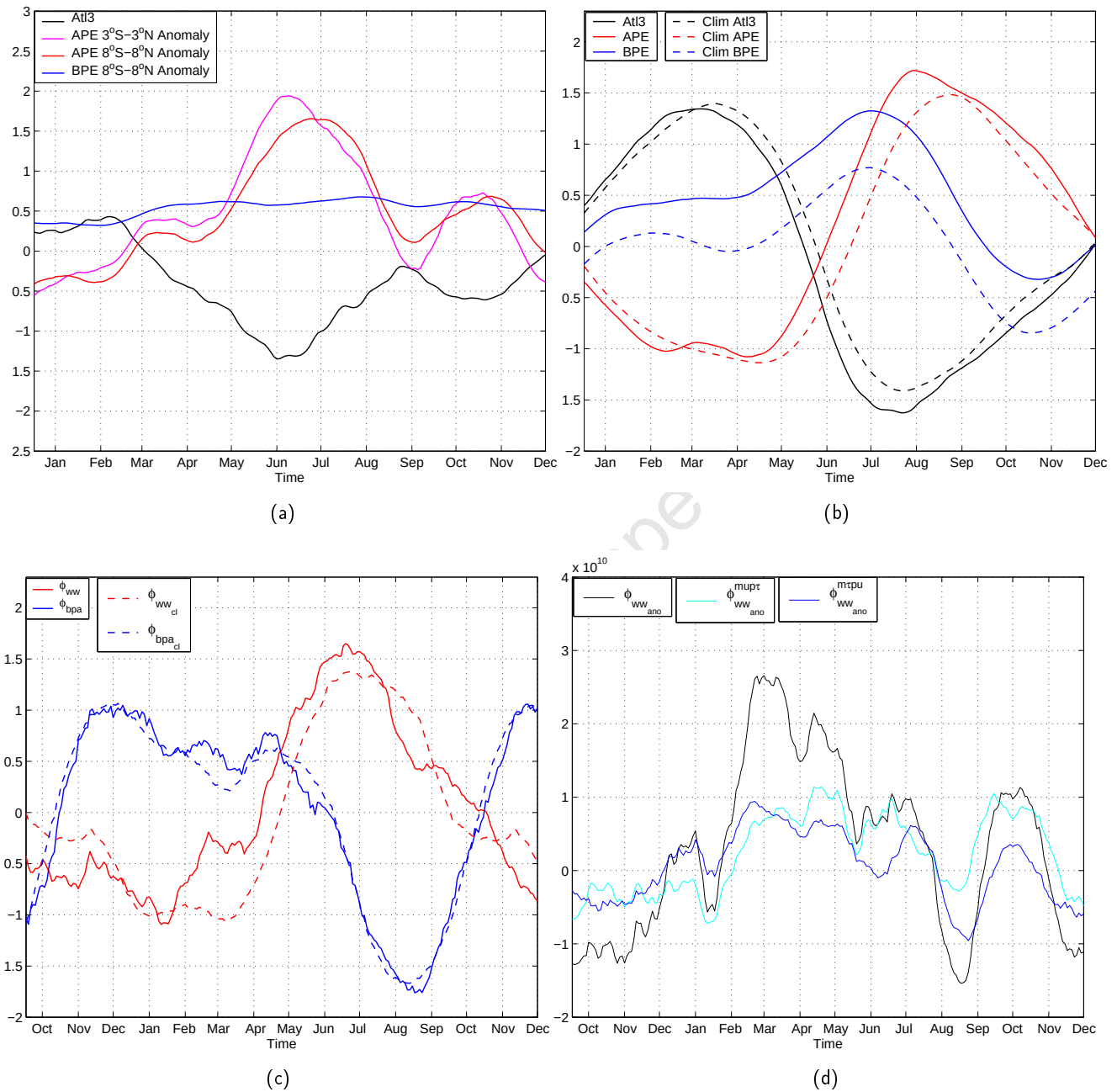


Figure 6.18: Cold year (1983, 1986, 1992, 1994 and 2004) composites of (a) anomalous Atl3 SST, equatorial Atlantic APE (3°S-3°N 60°W-15°E 0-400m), tropical Atlantic APE (8°S-8°N 60°W-15°E 0-400m) and tropical Atlantic BPE (8°S-8°N 60°W-15°E 0-400m); (b) total versus climatological values of Atl3 SST, tropical Atlantic APE and tropical Atlantic BPE; (c) total versus climatological values for wind work (Φ_{ww}) and advection of the density field (Φ_{bpa}); and (d) the relative contribution of mean perturbation terms Φ_{ww}^{mupt} and Φ_{ww}^{mrpu} to anomalous wind power. In (a)-(c) values have been normalised while in (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability.

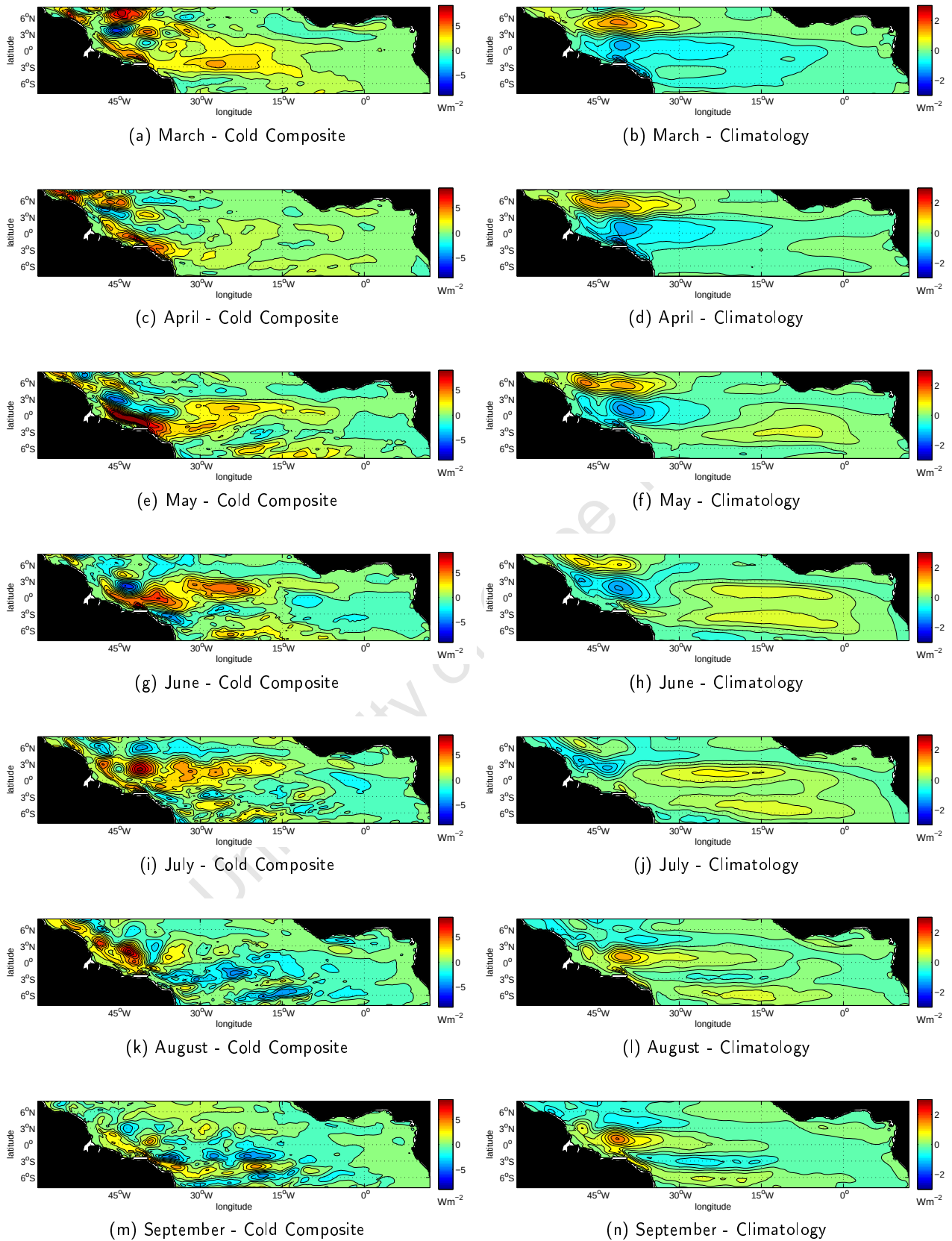


Figure 6.19: Cold year (1983, 1986, 1992, 1994 and 2004) composite monthly maps illustrating the spatial structure in wind power anomalies (Wm^{-2}) over the tropical Atlantic domain ($8^{\circ}S-8^{\circ}N$ $60^{\circ}W-15^{\circ}E$) between April and September are shown in the left column (a, c, e, g, i, k and m). These are compared with the spatial structure in climatological wind power anomalies from the annual mean in the right column (b, d, f, h, j, l and n).

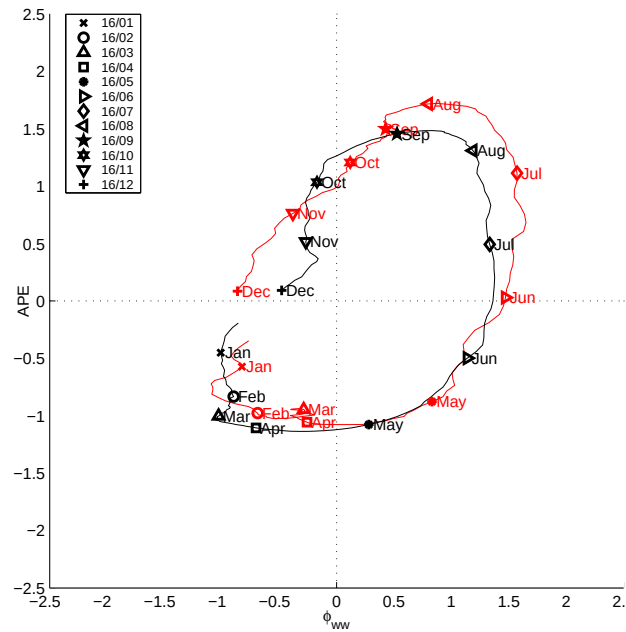


Figure 6.20: Cold year (1983, 1986, 1992, 1994 and 2004) composite of tropical Atlantic APE versus wind power phase diagram. The phase diagram based on composited total APE and wind power values is shown in red while the seasonal phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability.

Although during some cold event years, the delayed, negative, ocean memory feedback mechanism only acts to dissipate the anomalies, during others it manages to completely reverse them. This difference is illustrated by comparing the 1992 and 2004 cold events (Figures 6.21, 6.22 and 6.23). Figure 6.21 compares anomalous Atl3 SST, equatorial Atlantic APE, tropical Atlantic APE and tropical Atlantic BPE in 1992 against 2004, as well as the relative contribution of mean perturbation terms $\Phi_{ww_{ano}}^{mup\tau}$ and $\Phi_{ww_{ano}}^{m\tau pu}$ to anomalous 1992 and 2004 wind power. Figure 6.22 compares 1992 and 2004 total versus climatological values for Atl3 SST and equatorial Atlantic APE. Figure 6.23 compares the 1992 and 2004 tropical Atlantic APE versus wind power phase diagrams. In 1992, anomalously cold Atl3 SST and high APE conditions during boreal summer are damped between August and December (Figures 6.21a and 6.23a). Whereas in 2004, anomalously cold Atl3 SST and high APE conditions during boreal summer are reversed in August/September (Figures 6.21b and 6.23b). When comparing Figures 6.21c and 6.21d, it becomes apparent that the delayed negative feedback associated with $\Phi_{ww_{ano}}^{m\tau pu}$ manages to reverse Atl3 SST and equatorial Atlantic APE anomalies in 2004, yet in 1992 anomalously strong winds acting on climatological surface currents ($\Phi_{ww_{ano}}^{mup\tau}$) persist until December 1992 and therefore anomalously weak wind power due to $\Phi_{ww_{ano}}^{m\tau pu}$ only manages to damp and not reverse anomalous APE conditions. When viewing SST and APE anomalies from a total versus climatology perspective (Figure 6.22), it appears as though the 2004 event was predominantly the result of a shift in the phase of the seasonally excited thermocline mode while the 1992 event was an amplification of the seasonally excited thermocline mode.

To gain insight into the typical evolution of pertinent variables during years which have experienced boreal summer warm events, the composite of 1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999 warm event years was created. Figure 6.24 shows the warm year composite of (a) anomalous Atl3 SST, equatorial Atlantic APE, tropical Atlantic APE and tropical Atlantic BPE; (b) total versus climatological Atl3 SST,

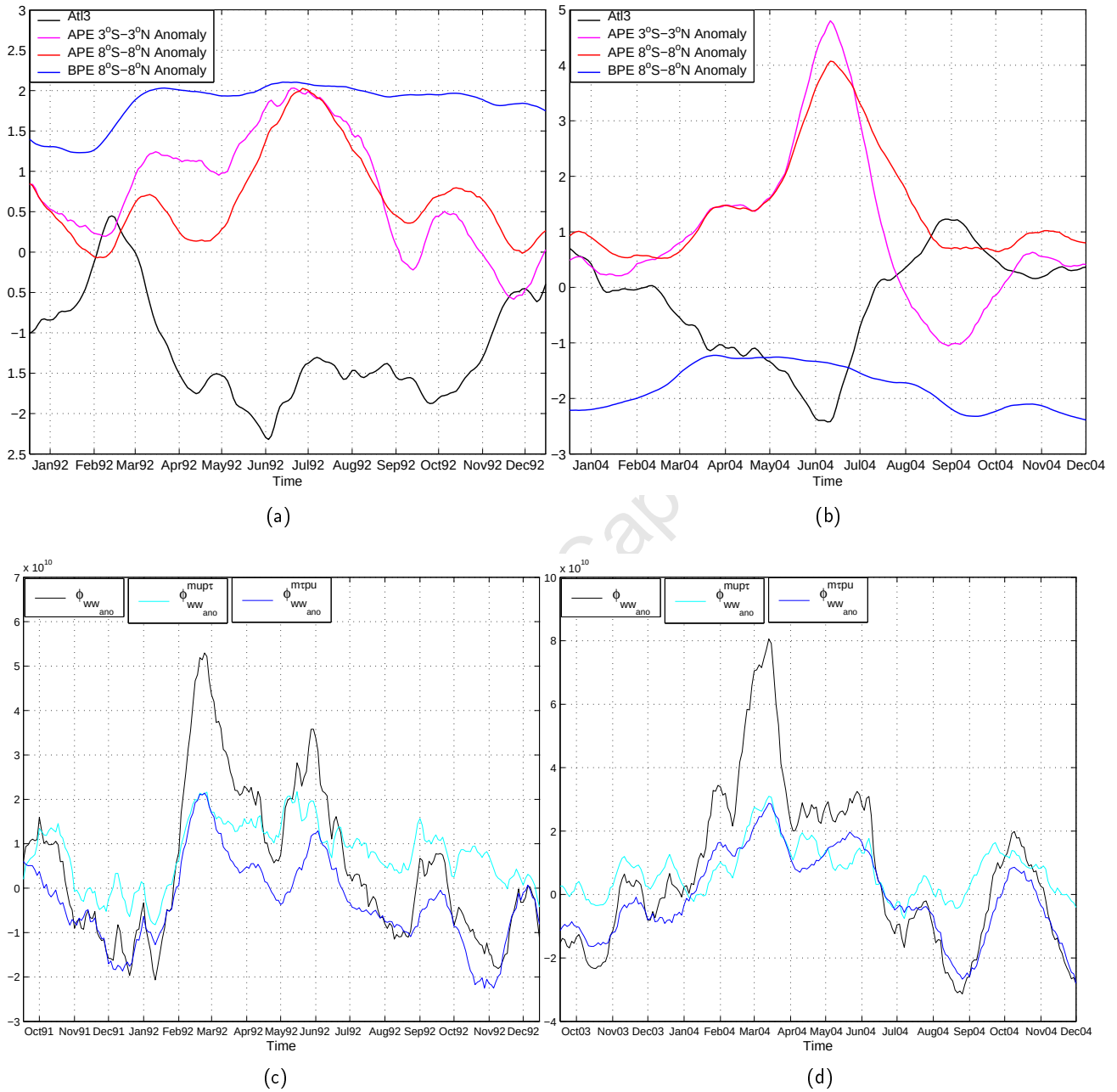


Figure 6.21: (a) 1992 and (b) 2004 anomalous Atl3 SST, equatorial Atlantic APE (3°S–3°N 60°W–15°E 0–400m), tropical Atlantic APE (8°S–8°N 60°W–15°E 0–400m) and tropical Atlantic BPE (8°S–8°N 60°W–15°E 0–400m). (c) The 1992 and (d) the 2004 relative contribution of mean perturbation terms $\Phi_{ww\,ano}^{mpt}$ and $\Phi_{ww\,ano}^{mtpu}$ to anomalous wind power. In (a) and (b) values have been normalised while in (c) and (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability.

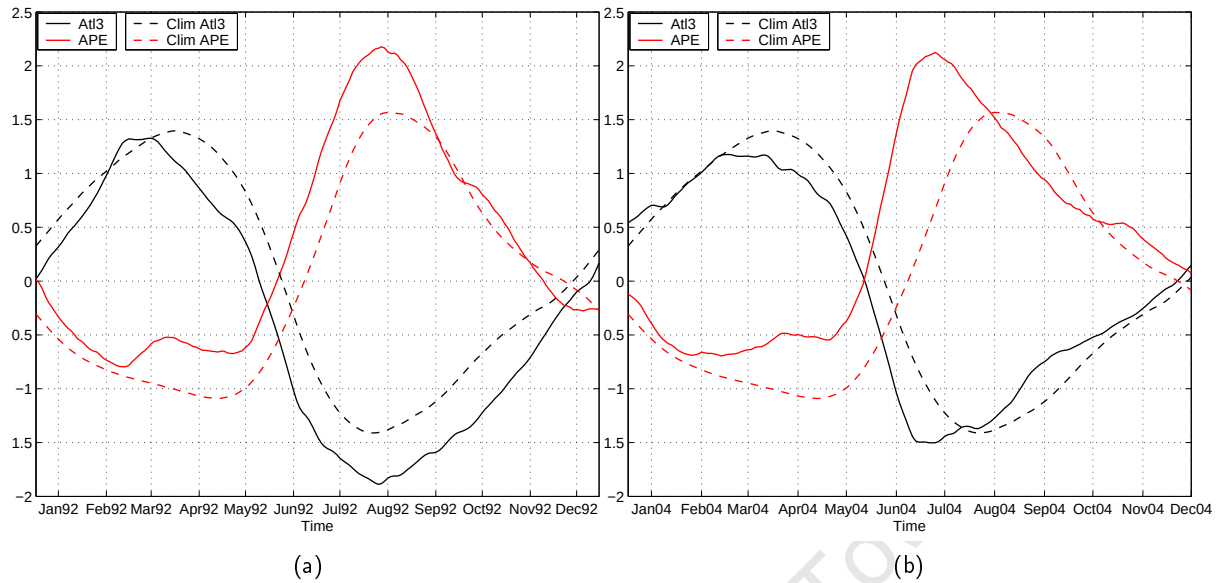


Figure 6.22: (a) 1992 and, (b) 2004, total versus climatological values for AtI3 SST and equatorial Atlantic APE. Values have been normalised. A 14 day running mean has been applied to each time series to smooth out the high frequency variability.

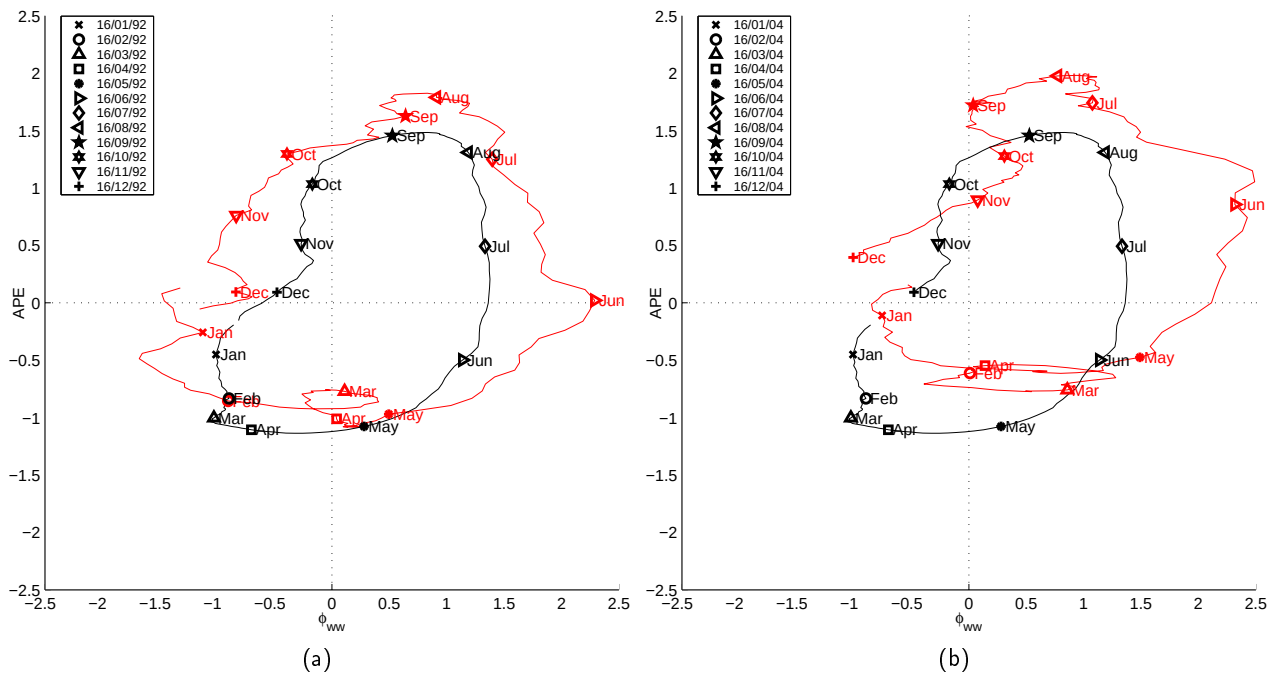


Figure 6.23: Tropical Atlantic APE versus wind power phase diagram for (a) 1992 and (b) 2004. The phase diagram based on total APE and wind power values is shown in red while the climatological phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability.

tropical Atlantic APE and tropical Atlantic BPE values; (c) total versus climatological values for wind work (Φ_{ww}) and advection of the density field (Φ_{bpa}); and (d) the relative contribution of mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ to anomalous wind power. Figure 6.24a shows that the zonal mode warm events are typified by anomalously low equatorial and tropical APE values which reach a minimum in July/August. The size of APE anomalies is greater for the cold composite (Figure 6.18a) than the warm (Figure 6.24a), but warm events are typically associated with much larger BPE anomalies which act to reinforce the effects of APE anomalies on SST. Considering these anomalous conditions in the context of total versus climatological conditions (Figure 6.24b) illustrates that both boreal summer SST and APE result from the decrease of the seasonal decline (increase) in SST (APE) typically associated with the development of the equatorial cold tongue and its associated Bjerknes feedback.

Anomalously low APE between June and August results from anomalously weak wind power between April and July (Figures 6.24c and 6.24d). The relative contribution of mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ to anomalous wind power depicted in Figure 6.24d indicates that the anomalously weak wind power between February and July is due to the combined effect of anomalously weak wind stress acting on climatological surface currents ($\Phi_{wwano}^{mup\tau}$) and climatological wind stress acting on anomalously weak westward surface currents ($\Phi_{wwano}^{m\tau pu}$). Although $\Phi_{wwano}^{m\tau pu}$ initially reinforces $\Phi_{wwano}^{mup\tau}$ between May and July, it reverses in July resulting in anomalously strong wind power forcing between August and November that acts to weaken the APE anomalies.

Figure 6.25 illustrates the typical spatial structure and temporal evolution of the anomalous wind power field during warm events in comparison with the spatial structure and temporal evolution of climatological wind power anomalies from the annual mean. Between March and July (Figures 6.25a, 6.25c, 6.25e, 6.25g and 6.25i), anomalously weak wind power values are observed over the 8°S-2°N and 50°W-0°E region and correspond with the anomalously weak domain integrated wind power seen during this period in Figures 6.24c and 6.24d. The spatial structure and temporal evolution of these weak wind power anomalies (Figures 6.25e, 6.25g and 6.25i) is similar to that of the seasonally enhanced wind power associated with the development of the cold tongue (Figures 6.25f, 6.25h and 6.25j). As with cold events, the off-equatorial Rossby wave influence is evident as regions of anomalously weak wind power are situated either side of the equator and propagate westward. Likewise, the spatial structure associated with the positive wind power anomalies within this region between July and October (Figures 6.24c and 6.24d), which act to increase APE and decay the warm event, is the opposite of the decreased wind power associated with the seasonal cycle during these months.

Figure 6.26 compares the warm year composite, APE versus wind power, phase diagram with the climatological phase diagram. Anomalously weak wind power during the early months of the year suppresses the seasonal increase in APE and the associated seasonally excited Bjerknes feedback between April and August (Figure 6.26). However as a result of this suppressed Bjerknes feedback, the delayed, seasonal surface current deceleration between August and October is also suppressed and therefore the seasonally active delayed, negative, ocean memory feedback mechanism associated with the seasonally excited thermocline mode is anomalously weak resulting in a return to near normal condition by October (6.24).

As with cold events, the delayed, ocean feedback mechanism simply weakens anomalous conditions during some warm event years, while during others it manages to completely reverse anomalous conditions. This difference is illustrated by comparing the 1988 and 1991 warm events in Figures 6.27, 6.28 and 6.29.

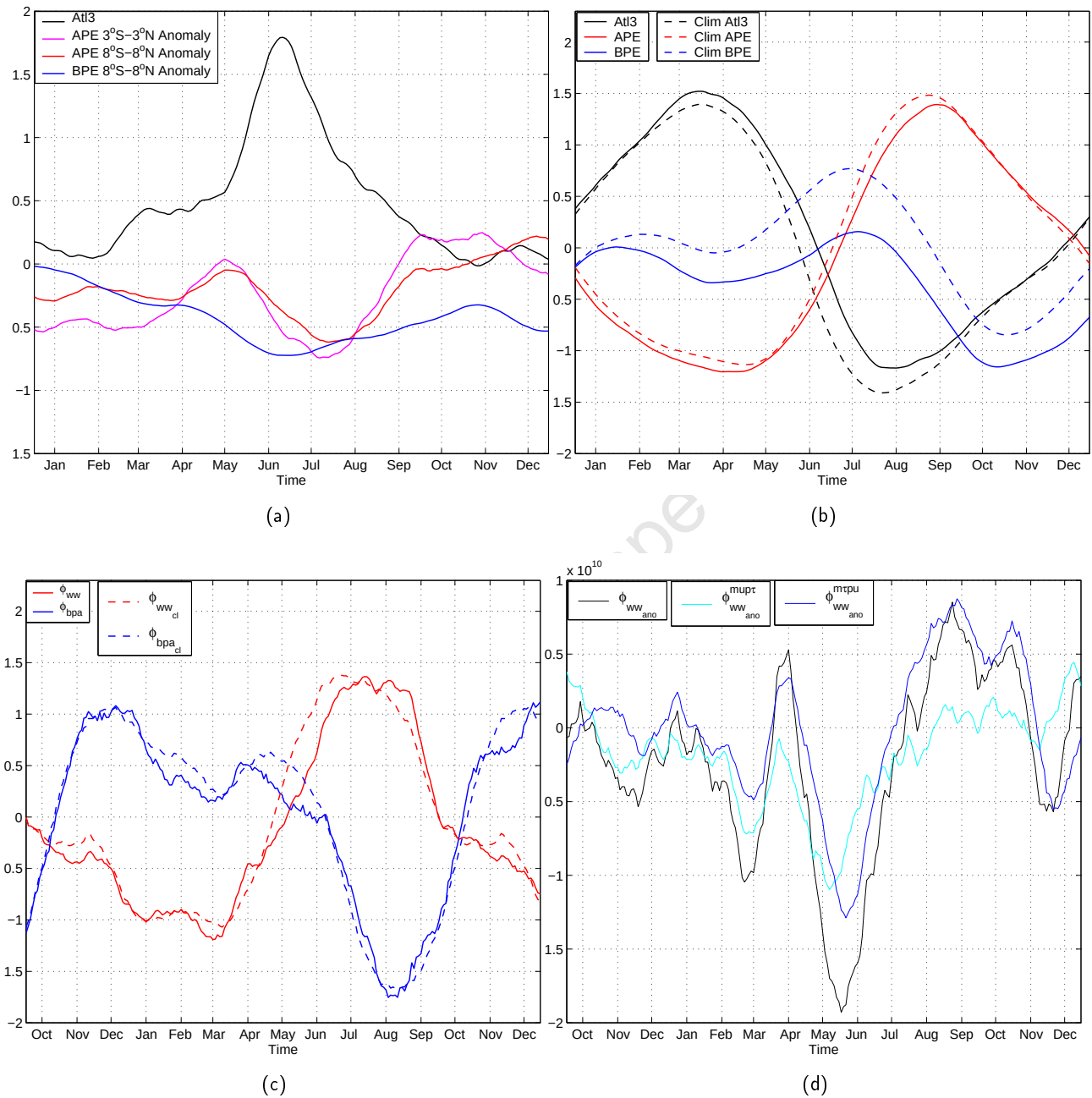


Figure 6.24: Warm year (1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999) composites of (a) anomalous Atl3 SST, equatorial Atlantic APE (3°S–3°N 60°W–15°E 0–400m), tropical Atlantic APE (8°S–8°N 60°W–15°E 0–400m) and tropical Atlantic BPE (8°S–8°N 60°W–15°E 0–400m); (b) total versus climatological values of Atl3 SST, tropical Atlantic APE and tropical Atlantic BPE; (c) total versus climatological values for wind work (Φ_{ww}) and advection of the density field (Φ_{bpa}); (d) the relative contribution of mean perturbation terms $\Phi_{ww_{ano}}^{mupr}$ and $\Phi_{ww_{ano}}^{mtpu}$ to anomalous wind power. In (a)–(c) values have been normalised while in (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability.

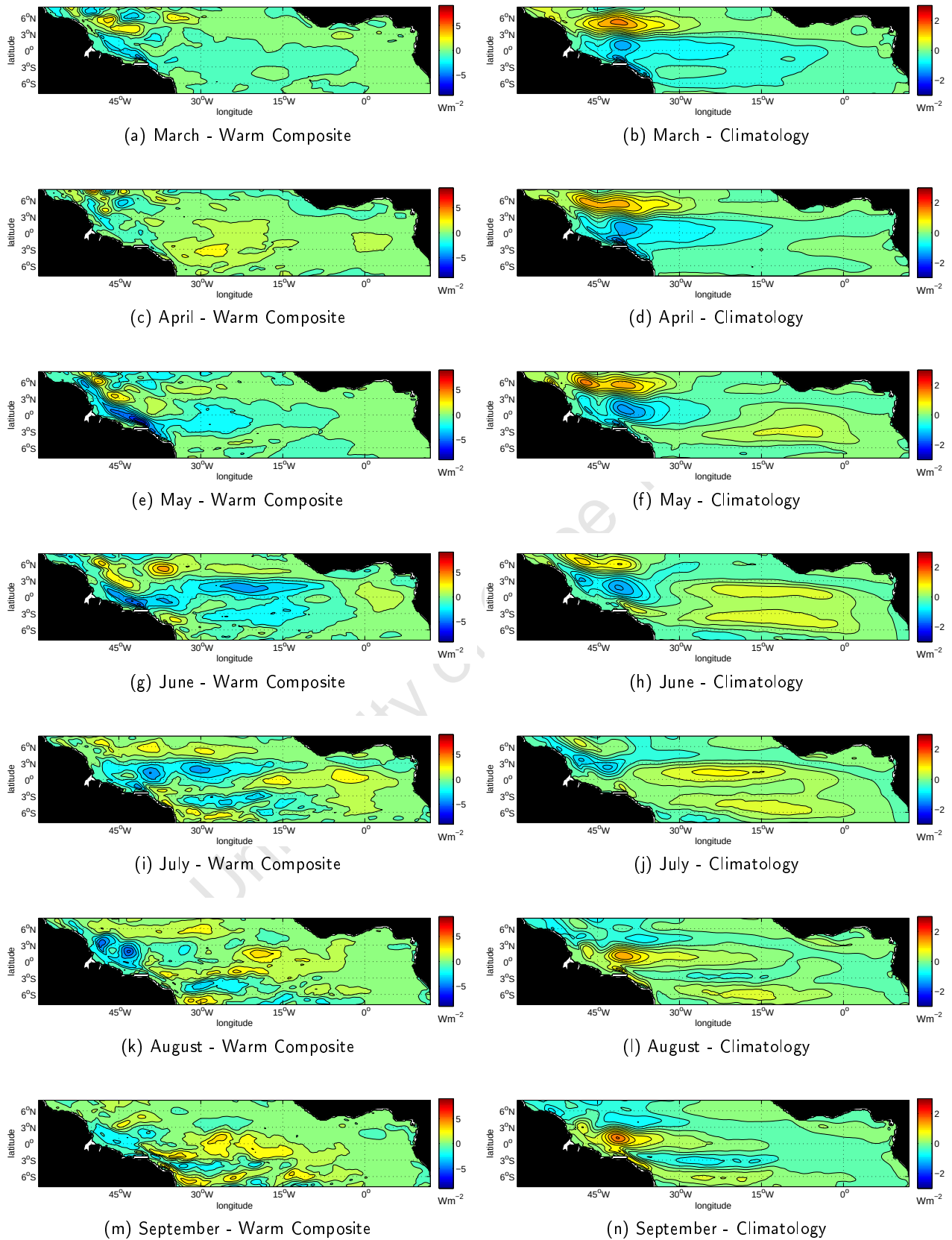


Figure 6.25: Warm year (1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999) composite monthly maps illustrating the spatial structure in wind power anomalies (Wm^{-2}) over the tropical Atlantic domain (8°S-8°N 60°W-15°E) between April and September are shown in the left column (a, c, e, g, i, k and m). These are compared with the spatial structure in climatological wind power anomalies from the annual mean in the right column (b, d, f, h, j, l and n).

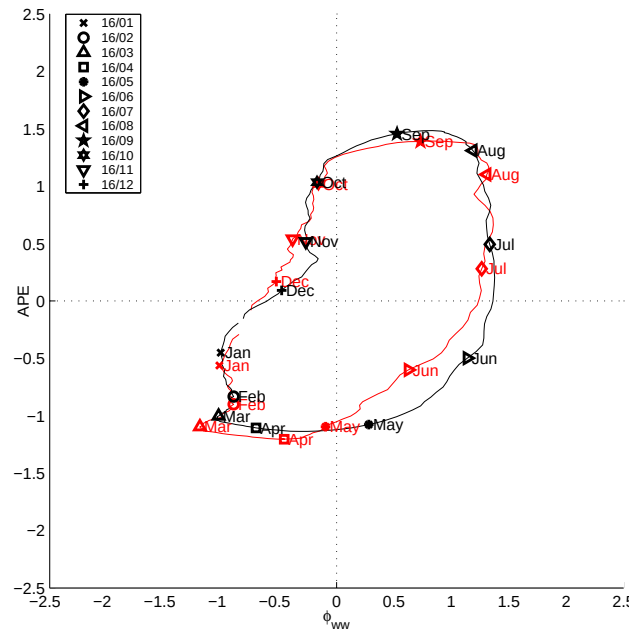


Figure 6.26: Warm year (1984, 1987, 1988, 1991, 1995, 1996, 1998 and 1999) composite of tropical Atlantic APE versus wind power phase diagram. The phase diagram based on composited total APE and wind power values is shown in red while the climatological phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability.

Figure 6.27 compares anomalous Atl3 SST, equatorial Atlantic APE, tropical Atlantic APE and tropical Atlantic BPE in 1988 against 1991, as well as the relative contribution of mean perturbation terms $\Phi_{wwano}^{mup\tau}$ and $\Phi_{wwano}^{m\tau pu}$ to anomalous 1988 and 1991 wind power. Figure 6.28 compares 1988 and 1991 total versus climatological values for Atl3 SST and equatorial Atlantic APE. Figure 6.29 compares the 1988 and 1991 tropical Atlantic APE versus wind power phase diagrams. Anomalous warm Atl3 SST and low APE conditions during the boreal summer of 1988 were damped between September and November (Figures 6.27a and 6.29a), whereas in 1991, anomalously warm Atl3 SST and low APE conditions between May and July reversed between August and December (Figures 6.27b and 6.29b). On comparing Figures 6.27c and 6.27d, it becomes apparent that in 1991 the delayed negative feedback associated with $\Phi_{wwano}^{m\tau pu}$ manages to reverse Atl3 SST and equatorial Atlantic APE anomalies, while in 1988 the anomalously weak winds acting on the climatological surface currents ($\Phi_{wwano}^{mup\tau}$) persist until August. In 1988, anomalously strong wind power due to $\Phi_{wwano}^{m\tau pu}$ between September and November only manages to damp and not reverse anomalous APE conditions. When viewing SST and APE anomalies from a total versus climatology perspective (Figure 6.28), it appears as though the 1991 event was predominantly the result of a shift in the phase of the seasonally excited thermocline mode while the 1988 event was the result of a decrease of the seasonally excited thermocline mode. In 1988, Atl3 SST reaches its minimum, and APE its maximum, in late July - early August, which is when these variables reach their climatological minimum (maximum). However, SST (APE) values are substantially higher (lower) than seasonal values (Figure 6.28). On the other hand, in 1991, Atl3 SST reaches its minimum, and APE its maximum, in late August- early September, and while APE values exceed their annual maximum, Atl3 SST values do not fall below their annual minimum (Figure 6.28). In Figure 6.28b it is evident that this shift in the timing of when Atl3 SST and equatorial APE reach their annual minimum/maximum results in anomalously warm SST prior to August and anomalously cool

SST subsequently.

6.4.2 Anomalous BPE

Figure 6.30 shows the relationship between inter-annual anomalies in equatorial and tropical Atlantic BPE. These two variables are strongly related with a maximum correlation coefficient of $r=0.91$ ($r^*=0.19$) found between them at zero lag. These BPE anomalies are primarily driven by fluctuations in the advection of the density field as shown by Figure 6.31. For the equatorial and tropical Atlantic domains, Figure 6.31 compares monthly BPE evolution anomalies, $\text{dBPE}_{\text{ano}}/\text{dt}$, with anomalous BPE evolution driven by fluctuations in the advection of the density field, $\Phi_{\text{bpa}_{\text{ano}}} = A_{\text{BPE}} - A_{\text{BPE}_{\text{cl}}}$ (Equation 5.19). Section 5.3.3 showed that seasonal BPE fluctuations are largely driven by seasonal fluctuations in the advection of the density field ($\Phi_{\text{bpa}_{\text{cl}}}$) that appear to be linked with seasonal changes in tropical Atlantic warm water escape. As seen in Figure 6.31, anomalous inter-annual equatorial and tropical Atlantic BPE evolution is primarily forced by inter-annual anomalies in the advection of the density field with a correlation coefficient of $r=0.94$ ($r^*=0.19$) between $\text{dBPE}_{\text{ano}}/\text{dt}$ and $\Phi_{\text{bpa}_{\text{ano}}}$ for the equatorial domain and $r=0.89$ ($r^*=0.19$) for the tropical domain. This result suggests that inter-annual BPE anomalies are primarily forced by the same process as seasonal BPE fluctuations, namely variability in the warm water escape process. Figure 6.18c suggests that anomalously high BPE typically associated with cold events (Figure 6.18a) is the result of anomalously high warm water escape between January and May, while Figure 6.24c suggests that the anomalous low BPE typically associated with warm events (Figure 6.24a) is the result of anomalously low warm water escape between January and June.

At the end of Section 5 we questioned whether anomalous BPE, particularly during April-May, could be one of the factors responsible for inter-annual modulations of the seasonally excited Bjerknes feedback. As revealed by the analysis conducted in Section 6.3, anomalous APE changes are not directly influenced by anomalous BPE changes as fluctuations in the wind power term are seen to be the dominant source of anomalous APE evolution. This finding does not however rule out the possibility that anomalously high (low) BPE during April-May could precondition the region and thus favour an enhanced (suppressed) seasonally excited Bjerknes feedback. The results of the composite analysis shown in Figures 6.18 and 6.24 support this hypothesis. BPE composite anomalies (Figures 6.18a and 6.24a) suggest that cold (warm) zonal mode events are typically associated with anomalously high (low) BPE during April-May. Figures 6.18a and 6.18b suggest that anomalous high BPE is generally associated with the occurrence of cold events, while Figures 6.24a and 6.24b suggest that anomalous low BPE is generally associated with the occurrence of warm events.

However Figure 6.32, which examines the relationship between equatorial Atlantic APE and BPE anomalies, shows an insignificant correlation of $r=0.27$ ($r^*=0.29$) between these variables at zero lag (Figure 6.32a). The maximum relationship occurs between equatorial Atlantic APE and BPE anomalies when BPE leads by one month ($r=0.34$, Figure 6.32b). Large BPE changes occur as rapidly as APE changes but less frequently (Figure 6.32a). When considering the seasonal dependence within the relationship between anomalous equatorial Atlantic APE and BPE it is evident that the strongest relationship between these variables occurs between August and December (Figure 6.32c, correlations are significant during this period). Correlation values seen in Figure 6.32 suggest that low (high) BPE anomalies generally lead low (high) APE anomalies by several months between May and October. However, between May and August correlation values are

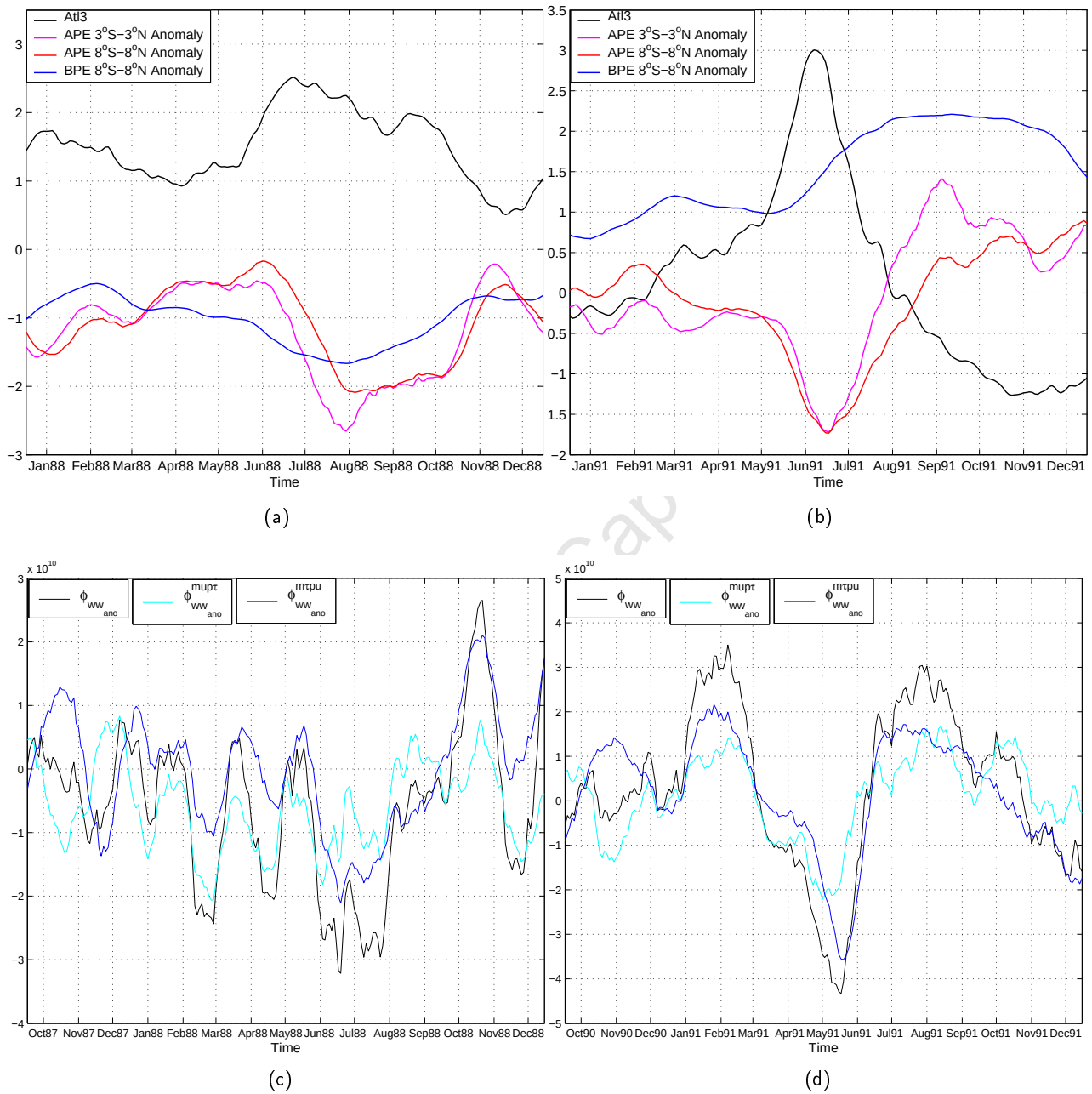


Figure 6.27: (a) 1988 and (b) 1991 anomalous Atl3 SST, equatorial Atlantic APE (3°S–3°N 60°W–15°E 0–400m), tropical Atlantic APE (8°S–8°N 60°W–15°E 0–400m) and tropical Atlantic BPE (8°S–8°N 60°W–15°E 0–400m). (c) The 1988 and (b) the 1991 relative contribution of mean perturbation terms $\Phi_{ww\,ano}^{mupr}$ and $\Phi_{ww\,ano}^{mtpu}$ to anomalous wind power. In (a) and (b) values have been normalised while in (c) and (d) values are given in Js^{-1} . A 14 day running mean has been applied to each time series to smooth out the high frequency variability.

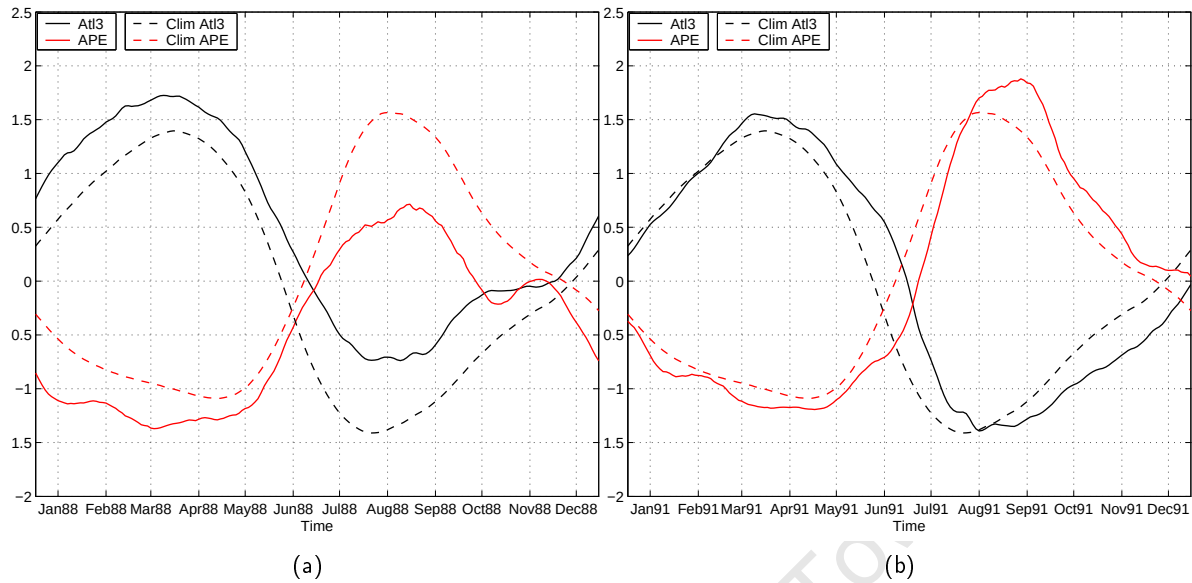


Figure 6.28: (a) 1988 and, (b) 1991, total versus climatological values for AtI3 SST and equatorial Atlantic APE. Values have been normalised. A 14 day running mean has been applied to each time series to smooth out the high frequency variability.

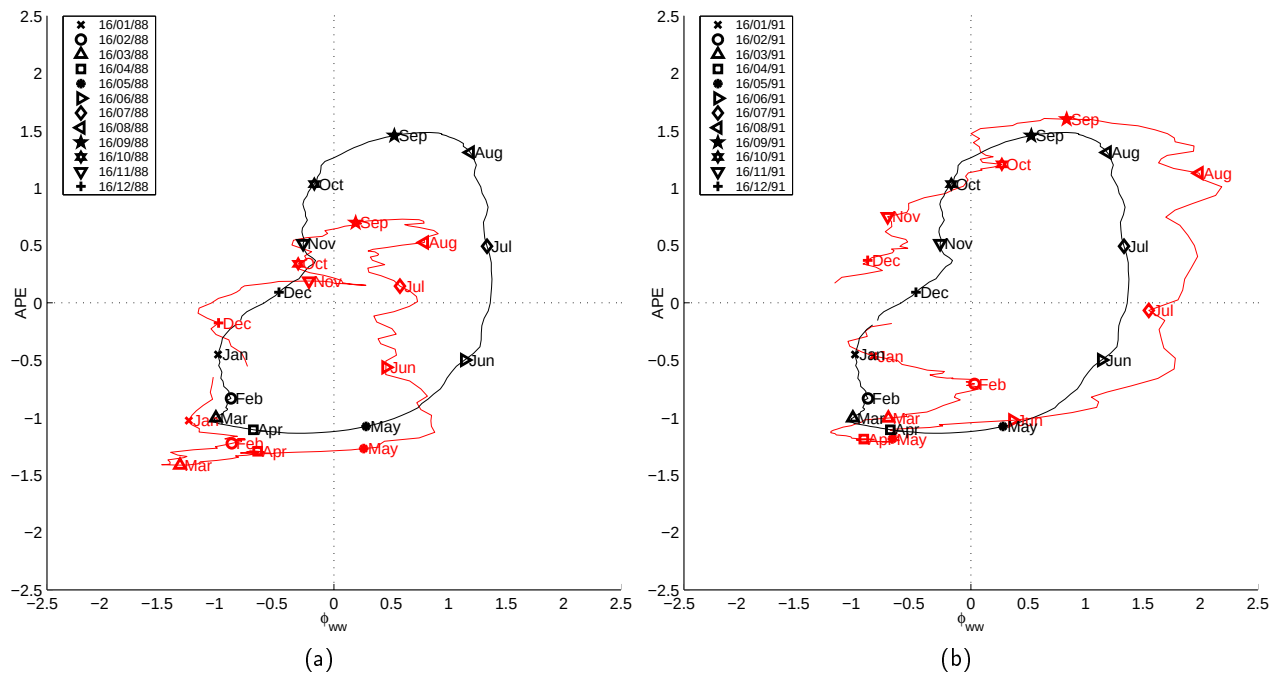


Figure 6.29: Tropical Atlantic APE versus wind power phase diagram for (a) 1988 and (b) 1991. The phase diagram based on total APE and wind power values is shown in red while the climatological phase diagram as in Figure 5.27 is shown in black. Values have been normalised and a 14 day running mean has been applied to the time series to smooth out the high frequency variability.

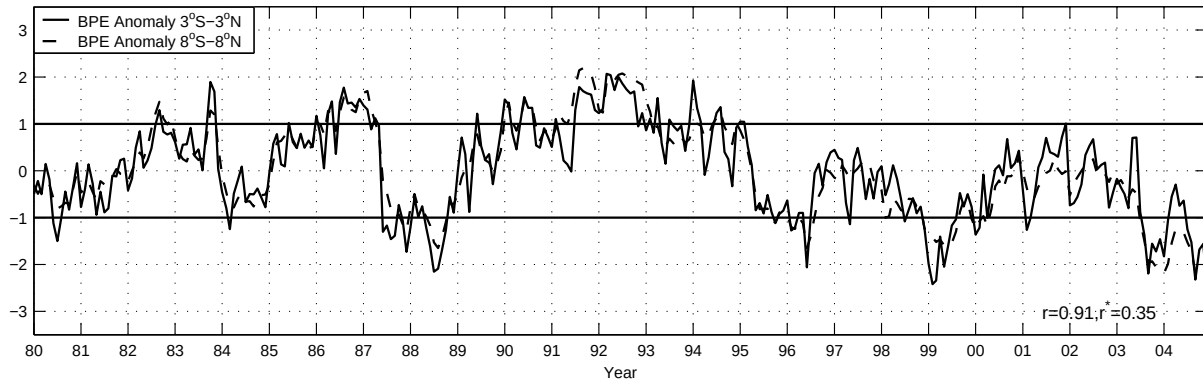
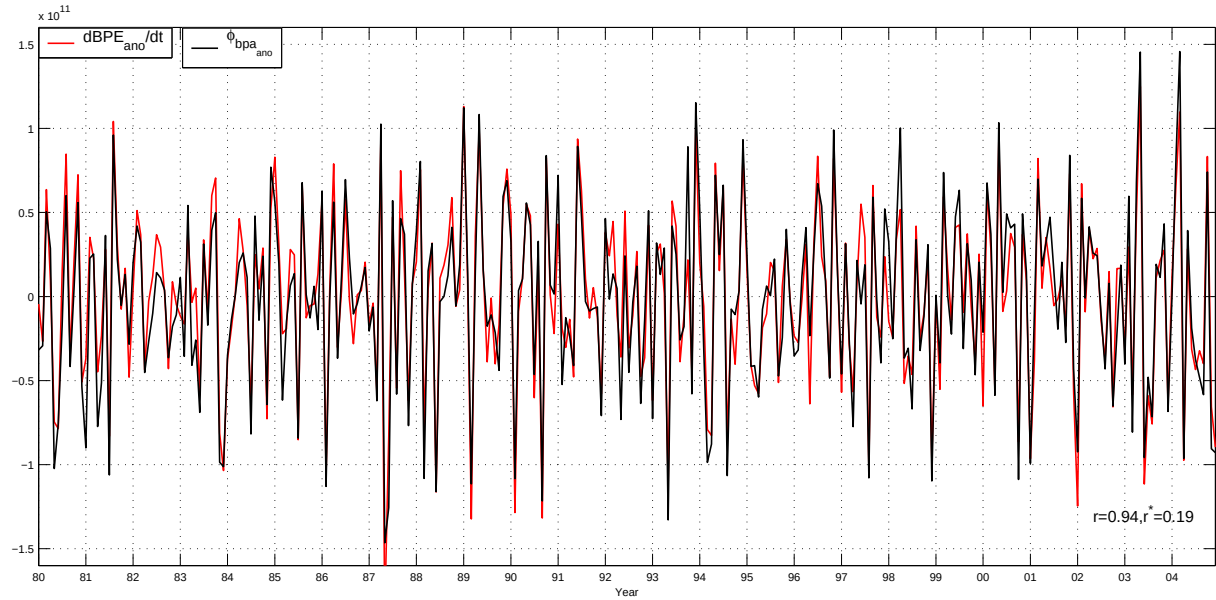


Figure 6.30: A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) BPE anomalies and tropical Atlantic (8°S - 8°N 60°W - 15°E 0-400m) BPE anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). BPE anomalies were calculated based on ROMS-TAtI output following Equation 3.2 and have been normalised.

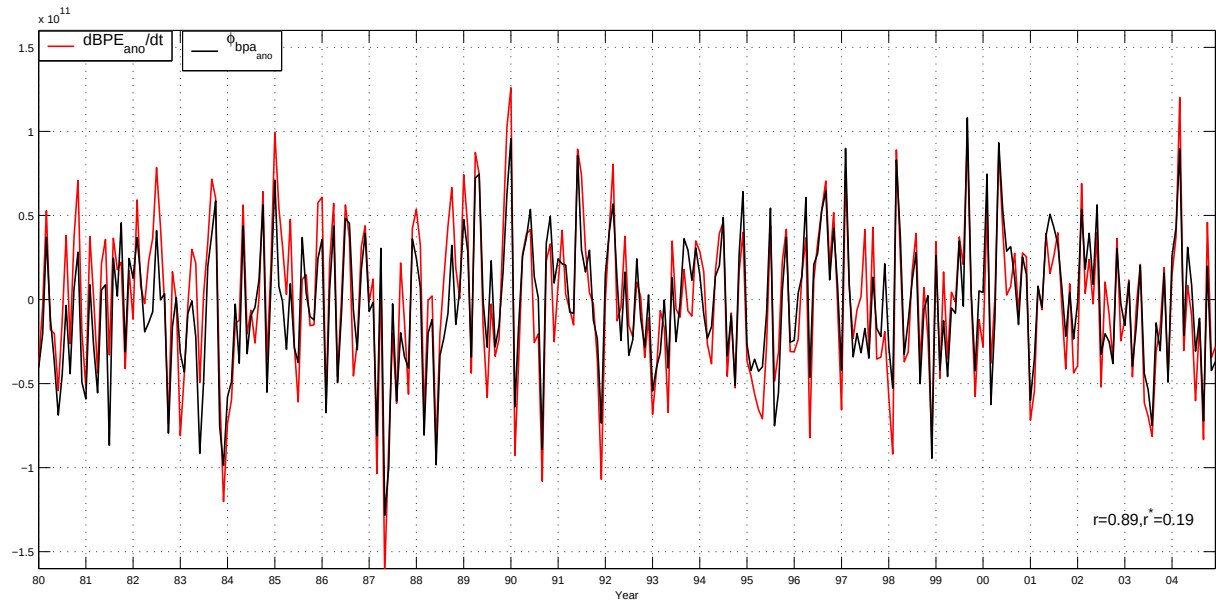
weak with only patches of significant correlations seen.

Values greater/less than one in Figure 6.30 indicate when monthly anomalies in equatorial and tropical Atlantic BPE exceed/fall below one standard deviation as values have been normalised. While the 1984, 1987, 1988, 1995, 1996, 1998 and 1999 warm events all coincided with periods of anomalously low BPE, with anomalous BPE values either approaching or exceeding one standard deviation, the 1991 event did not. In fact, BPE values were exceptionally high in 1991 (Figures 6.27b and 6.30). Likewise when one considers cold events, the 1983, 1986, 1992 and 1994 events all corresponded with periods of high BPE while the 2004 cold event occurred amidst a period when the BPE of the tropical Atlantic was exceptionally low (Figures 6.21b and 6.30). This analysis suggests that while anomalous BPE may be one factor contributing to the inter-annual modulations of the seasonally excited Bjerknes feedback it is not the dominant factor: the anomalous evolution of the seasonally excited Bjerknes feedback is not consistently associated with anomalous conditions in the background state of the upper tropical Atlantic ocean. Furthermore, it suggests that variability in the atmospheric component of the coupled system (remote atmospheric forcing or stochasticity of the atmosphere) rather than the BPE of the oceanic component may be the more dominant mechanism behind inter-annual modulations in the Bjerknes feedback which is seasonally excited in the Atlantic.

It is also worth noting the direct relationship seen between BPE and Atl3 SST anomalies. A significant anti-correlation of $r=-0.49$ ($r^*=0.29$) exists between equatorial Atlantic BPE and Atl3 SST anomalies at zero lag as shown in Figure 6.33a. The maximum anti-correlation ($r=-0.5$) is found between equatorial Atlantic BPE and Atl3 SST anomalies when the former leads by one month (Figure 6.33b). Figure 6.33c explores whether there is any seasonal dependence in the lag-lead correlation between monthly Atl3 SST and equatorial Atlantic BPE anomalies. The relationship between anomalous BPE and SST is strongest between July and October with BPE anomalies lead by 1-2 months. This result suggests that Atl3 SST variability, particularly between July and October exhibits a dependence on BPE anomalies during prior months. Seasonally this period coincides with the months of the year when advection of the density field (Φ_{bpa}) acts to decrease BPE (Figure 5.28a) as warm water escape to the north is inhibited (Section 5.3.3).



(a)



(b)

Figure 6.31: A comparison between monthly inter-annual anomalies in BPE evolution ($\text{dBPE}_{\text{ano}}/\text{dt}$ in Js^{-1}) and BPE evolution driven by fluctuations in the advection of the density field ($\Phi_{\text{bpa}_{\text{ano}}}$ in Js^{-1}), assessed over (a) the equatorial Atlantic domain ($3^{\circ}\text{S}-3^{\circ}\text{N}$ $60^{\circ}\text{W}-15^{\circ}\text{E}$ $0-400\text{m}$) and (b) the tropical Atlantic domain ($8^{\circ}\text{S}-8^{\circ}\text{N}$ $60^{\circ}\text{W}-15^{\circ}\text{E}$ $0-400\text{m}$). The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*).

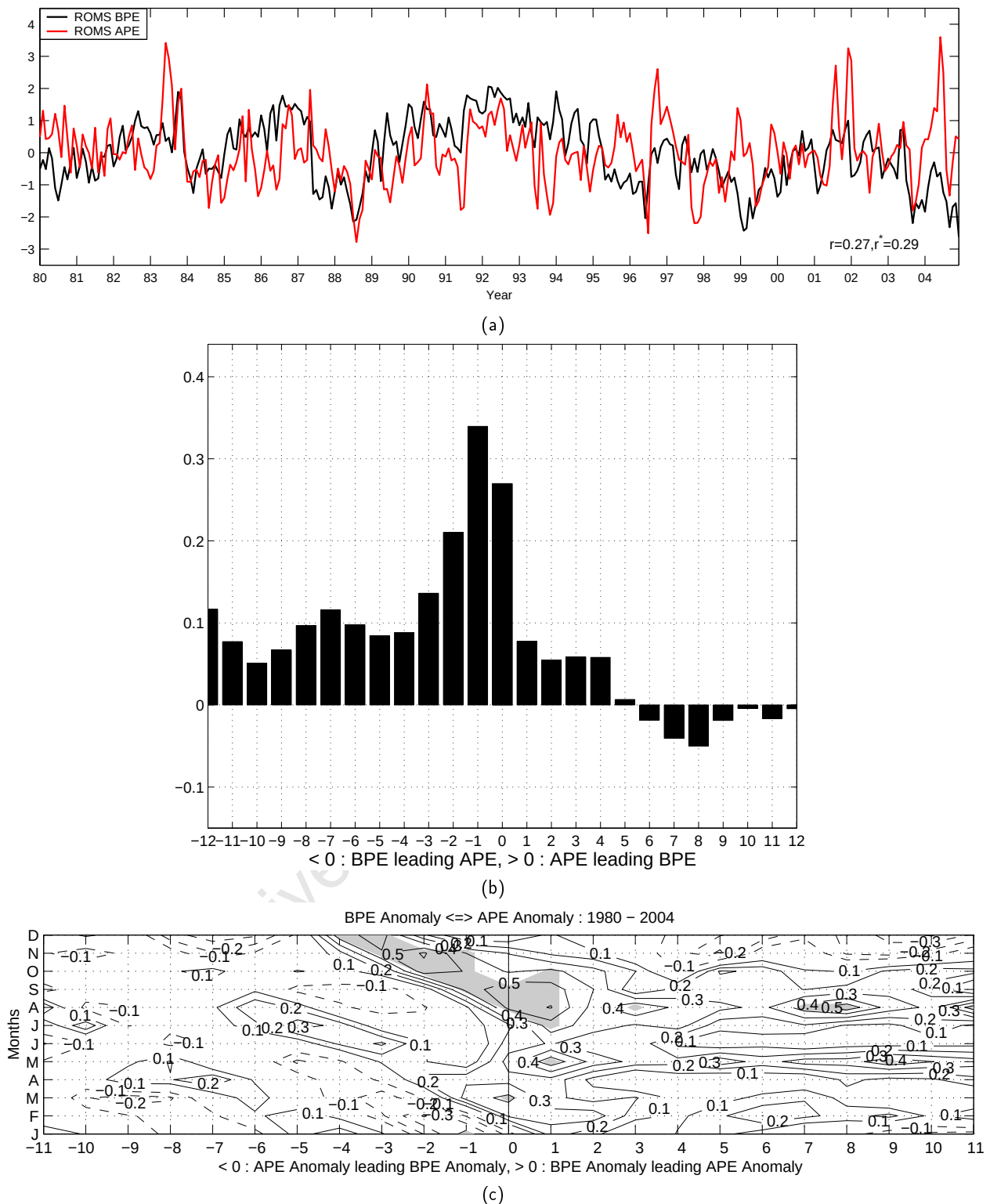


Figure 6.32: (a) A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE and BPE anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). APE and BPE values have been normalised. (b) Time lag correlations between equatorial APE and BPE anomalies. (c) The seasonal dependence in correlations between equatorial Atlantic APE and BPE anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. APE and BPE anomalies were calculated based on ROMS-TAtl output following Equation 3.2.

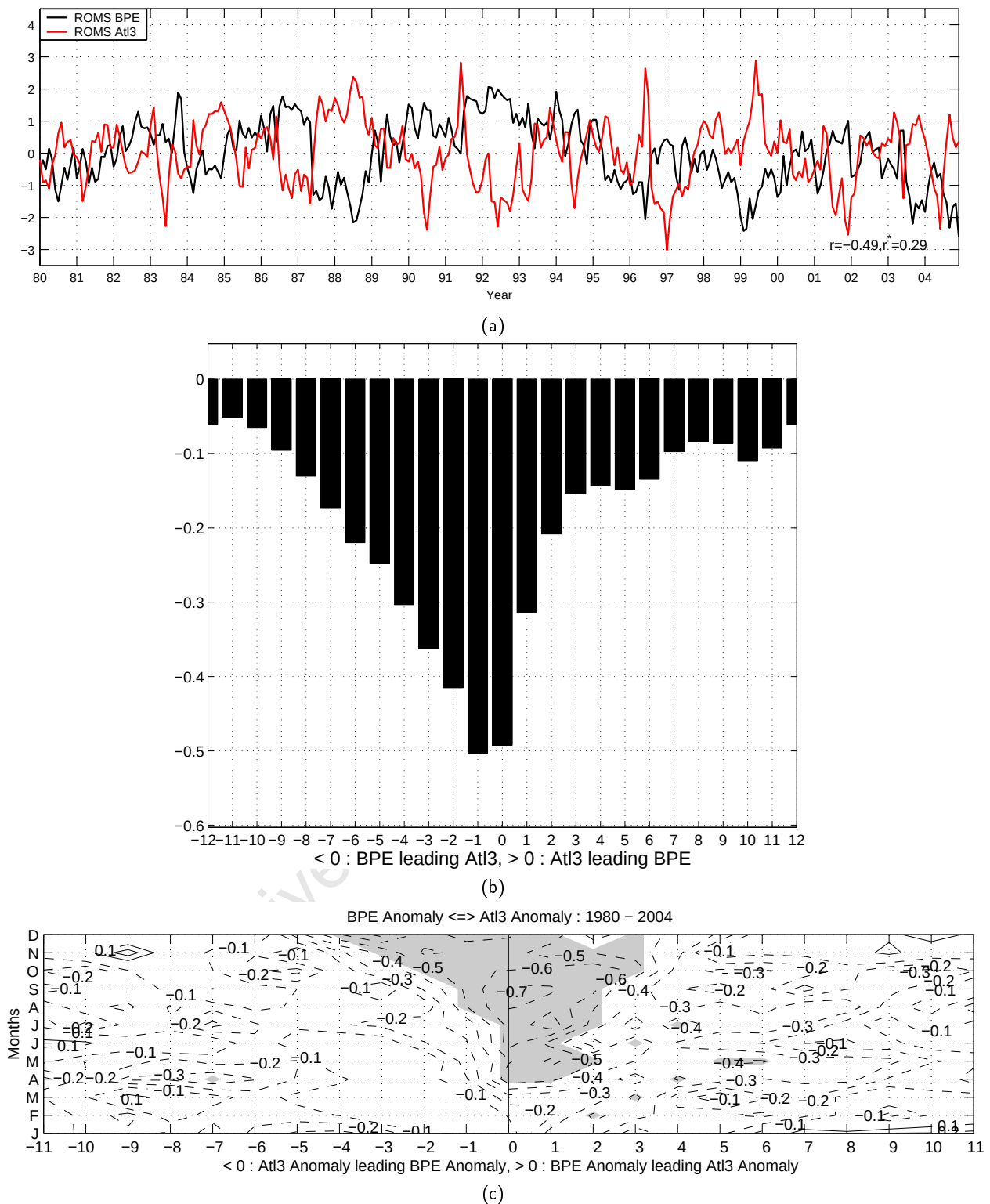


Figure 6.33: (a) A comparison between inter-annual equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) BPE and Atl3 SST anomalies. The correlation between anomalies (r) is given, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and BPE values have been normalised. (b) Time lag correlations between equatorial Atlantic BPE and Atl3 SST anomalies. (c) The seasonal dependence in correlations between equatorial Atlantic BPE and Atl3 SST anomalies. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey. BPE anomalies were calculated based on ROMS-TAtl output following Equation 3.2.

6.5 Conclusions

In this chapter the role of ocean dynamics within the zonal mode of the equatorial Atlantic has been assessed from an energetics perspective using the ROMS-TAtl simulation. The evidence offered here suggests that inter-annual variability within the equatorial Atlantic may largely be regarded as a modulation of seasonally active coupled processes.

The greatest inter-annual variability in Atl3 SST, and equatorial and tropical Atlantic APE occurs during boreal summer. Variability in the timing and intensity of seasonally stimulated subsurface cooling associated with remotely forced thermocline depth variability and the development of the cold tongue during boreal summer appears to be the primary cause of SST anomalies associated with the zonal mode. Anomalous APE during these months is strongly correlated with anomalous Atl3 SST and largely forced by anomalous wind power over the tropical Atlantic domain. The anomalous relationship between wind power, APE and Atl3 SST between April and July suggests that inter-annual variability during these months is largely driven by the anomalous evolution of the Bjerknes feedback which is seasonally excited during these months.

Analysing inter-annual thermocline depth variability associated with the zonal mode from an energetics perspective, as well as within the context of seasonal variability, has provided insight into the role of the delayed, negative, ocean memory feedback mechanism. This feedback mechanism operates on much shorter time scales than in the Pacific and is seen as largely being associated with inter-annual modulations of the seasonally excited thermocline mode's delayed, negative feedback response between August and October.

The analysis conducted within this chapter therefore suggests that inter-annual variability in Atl3 SST correlated with anomalous APE can largely be regarded as inter-annual modulations of the seasonally excited Bjerknes feedback and its associated thermocline mode. While an amplified (suppressed) seasonally excited Bjerknes feedback between April-July leads to the growth of anomalous conditions, the seasonally active decay mechanism associated with the delayed deceleration of surface currents between August-October is also amplified (suppressed), thereby acting to damp the anomalous conditions.

Inter-annual variability in the seasonally excited thermocline mode appears to be the result of inter-annual anomalies in either the timing or intensity of the Bjerknes feedback which is seasonally excited between April and July. Inter-annual variability in the BPE of the tropical Atlantic domain, driven by inter-annual variability in the tropical Atlantic's warm water escape process, represents a mechanism potentially responsible for inter-annual modulations in the seasonally excited Bjerknes feedback. The results obtained in this chapter suggest however that the relationship between anomalous BPE and zonal mode events is inconsistent suggesting that remote atmospheric forcing may play a more dominant role in driving inter-annual fluctuations in the timing and intensity of the seasonally excited Bjerknes feedback and hence in the thermocline mode.

Chapter 7

Summary of Conclusions and Outlook

In this thesis an assessment of the energetics of equatorial Atlantic variability, and the physical processes driving SST variability, has offered insight into the role of the ocean within local equatorial Atlantic ocean-atmosphere interactions at both inter-annual and seasonal timescales.

A number of key questions were raised in the introductory chapter of this thesis (Chapter 1):

Firstly - What distinct processes occurring in the Atlantic may account for the observed differences in the seasonal central-eastern basin SST signature between the Pacific and the Atlantic? This question is addressed by results obtained in both Chapters 3 and 5. Distinctions between the seasonal SST signature in the central-eastern Pacific and the central-eastern Atlantic are ascribed to the presence, in the Atlantic, of a seasonally excited Bjerknes feedback as well as a delayed negative feedback mechanism which are absent from the Pacific. The seasonally active delayed negative feedback mechanism takes the form of ocean memory similar to that observed in the Pacific on inter-annual timescales. The sharper, enhanced seasonal decline of central-eastern basin SST observed in the Atlantic in comparison to the Pacific is attributed to the fact that the seasonally excited growth of asymmetric conditions about the equator is accompanied by a seasonally excited Bjerknes feedback as the thermocline shoals seasonally in the Atlantic. Furthermore, the fact that cool conditions do not persist as long as they do in the Pacific is ascribed to seasonal warming that is in-phase with a deepening of the thermocline in the east from August. This seasonal decay in the seasonally excited Bjerknes feedback and deepening of the thermocline in the east is seen to be the result of a delayed negative feedback mechanism similar to the negative feedback mechanism operating inter-annually in the Pacific. Between April-September, a circular relationship between APE and the work done on the ocean by the wind suggests that a seasonally excited thermocline mode of coupled variability plays an active role in the seasonal cycle in the tropical Atlantic.

Secondly - Why do inter-annual SST fluctuations dominate over seasonal fluctuations in the central-eastern Pacific while the opposite is true in the Atlantic? Is it because, unlike in the Pacific, the Bjerknes mechanism acts primarily on seasonal time scales in the Atlantic? Like in the Pacific, the dynamic process to which the development of the equatorial cold tongue is attributed is anomalous vertical advection due to anomalous Ekman pumping and its associated Ekman feedback mechanism (Mitchell and Wallace, 1992; Chang and Philander, 1994). An in-phase relationship between seasonal APE and Atl3 SST fluctuations suggests however that the large annual increase in the intensity of cooling due to vertical advection is not purely a locally forced phenomena, but is largely due to the horizontal redistribution of warm surface waters in

response to large scale changes in wind forcing. The results obtained in Chapters 3 and 5 point to the fact that the Bjerknes feedback mechanism, which only operates on inter-annual timescales in the Pacific, is in fact seasonally active in the Atlantic. Unlike in the Pacific where from an oceanic perspective a new mechanism is introduced at inter-annual timescales in the form of basin wide thermocline depth adjustment, this is not the case in the Atlantic as the largest APE changes occur seasonally. The difference in the time-scale of the largest SST fluctuations between the Pacific and Atlantic basins is therefore attributed to the fact in the Pacific the Bjerknes feedback and its associated delayed negative feedback mechanism operates on inter-annual time-scales whereas in the Atlantic it operates seasonally.

Thirdly - Can inter-annual SST variability in the central-eastern equatorial Atlantic be explained in terms of a modulation of seasonally active coupled variability and, if so, what factors are responsible for the modulation of seasonally active processes? This question is addressed in Chapters 3 and 6 where the role of ocean dynamics within the zonal mode has been assessed from an energetics perspective. Inter-annual SST variability in the central-eastern Atlantic correlated with anomalous thermocline depth variability appears to be largely due to a modification of the dynamic ocean processes which regulate eastern-central basin SST seasonally. In this regard the zonal mode is unlike ENSO where this process operates on inter-annual timescales and is dynamically distinct from the dominant process operating on seasonal timescales. Seasonal modulations in the timing and intensity of the seasonally excited Bjerknes feedback between April and July are seen as being the primary cause of anomalous SST variability associated with the zonal mode. This perspective on variability associated with the zonal mode accounts for differences between ENSO and the zonal mode mentioned in Chapter 2.

- Firstly, the weaker, less consistent, anomalous wind-SST relationship observed in the Atlantic in comparison to the Pacific. As an inter-annual modulation of the seasonally excited Bjerknes feedback, active in the Atlantic between the months of April and July, Atl3 SST and western basin zonal wind stress anomalies are only strongly related during these months. Atl3 SST anomalies which occur outside of this season are more likely to be the result of remote atmospheric forcing or a different local air-sea interaction process.
- Secondly, the fact that the Atlantic has more high frequency variability (shorter events) with larger contributions at seasonal time scales. This difference is attributed to the fact that coupled variability associated with the zonal mode is due to a modulation of seasonally active variability while that associated with ENSO is due to a natural mode of variability operating on inter-annual time scales.
- Thirdly, the zonal mode and ENSO exhibit very different relationships with the seasonal cycle, SST anomalies associated with ENSO events peak in boreal winter, while those associated with the zonal mode peak in boreal summer. Interpreting the zonal mode as an inter-annual modulation of the seasonally active thermocline mode which operates in the Atlantic during boreal summer months accounts for this difference.

Anomalous tropical Atlantic BPE appears to be a possible factor promoting the anomalous evolution of the Bjerknes feedback which is seasonally excited. However, the relationship between APE and BPE anomalies is inconsistent implying that atmospheric variability associated with remote atmospheric forcing or stochasticity of the atmosphere may play a more important role in modulating the timing and intensity of the seasonally excited Bjerknes feedback.

Finally - What is the role of ocean memory within central-eastern equatorial Atlantic SST variability? In Chapters 5 and 6, an investigation into the energetics of seasonal and inter-annual, equatorial and tropical Atlantic variability has shed light on the role of the delayed, negative, ocean memory feedback mechanism within equatorial Atlantic coupled variability at both seasonal and inter-annual timescales. As with the Bjerknes feedback, a delayed, negative, ocean memory feedback mechanism, similar to the negative feedback mechanism operating inter-annually in the Pacific, is seen to act primarily on seasonal timescales in the Atlantic. Transient induced changes in surface currents, associated with the delayed response of the tropical Atlantic to seasonal changes in wind stress forcing, affect the ability of the wind to do work on the ocean between August-October. This delayed, negative, ocean memory feedback mechanism is associated with the decay of the Bjerknes feedback that is seasonally excited between April and July as part of a seasonally active thermocline mode. From the APE-wind power perspective this negative feedback mechanism is seen in the form of transient induced changes in surface currents, associated with the delayed response of the tropical Atlantic to seasonal changes in wind stress forcing, which affect the ability of the wind to do work on the ocean. As seen inter-annually in the Pacific, this seasonally active delayed negative feedback is also associated with a recharge of equatorial zonal mean heat content (a decrease in BPE) in response to the intensification of the easterly trades due to the redistribution of mass between certain regions off the equator and the equatorial band. While the APE-wind power perspective does not conflict with the recharge/discharge of equatorial zonal mean heat content perspective it points to the importance of the the ocean wave propagation process and the associated changes in surface currents in the phase-transition mechanism.

In terms of inter-annual variability, the delayed negative feedback mechanism is seen as largely being associated with inter-annual modulations of the seasonally active delayed, negative feedback response. The ocean memory mechanism therefore appears to operate on much shorter time scales than in the Pacific. The analysis conducted in chapter 6 suggests that inter-annual variability in SST associated with the zonal mode can largely be regarded as inter-annual modulations of the seasonally excited Bjerknes feedback and its associated thermocline mode. While an amplified (suppressed) seasonally excited Bjerknes feedback between April-July leads to the growth of anomalous conditions, the seasonally active decay mechanism associated with the delayed deceleration of surface currents between August-October is also amplified (suppressed) thereby acting to damp the anomalies.

The analysis undertaken within this thesis has focused on the oceanic component of tropical Atlantic coupled ocean-atmosphere variability. As mentioned in the introductory chapter (Chapter 1), this oceanic focus is motivated by the fact that it is ocean dynamics (as the slow time scale and source of oscillatory behaviour) that usually provides predictability in the tropical coupled system. Although the atmospheric influence has been assessed within this thesis in terms of surface wind changes that are correlated with SST and APE changes, this limited approach is seen as one of the caveats of this study as the full atmospheric response to seasonal and inter-annual fluctuations in tropical Atlantic SST forcing has not been explicitly diagnosed. In terms of the seasonal response of the atmosphere to seasonal tropical Atlantic SST changes we rely largely on invoking the results obtained by Okumura and Xie (2004). As mentioned in Chapter 3, the sensitivity of the seasonal development of zonal winds in the western-central equatorial Atlantic to zonal SST gradients which develop as a result of the seasonal development of the cold tongue has been demonstrated by Okumura and Xie (2004). Using an atmospheric general circulation model Okumura and Xie (2004) have demonstrated the sensitivity of the atmosphere to the rapid cooling of SSTs in the central-

eastern equatorial Atlantic. During June-July-August, equatorial easterlies in the western-central Atlantic are greater in the control simulation forced by the full seasonal cycle in global SST than in the simulation where equatorial Atlantic SST is held constant in time from the 15 of April onwards. The response of the atmosphere to inter-annual SST anomalies has been assessed purely in terms of lag-lead zonal wind stress versus SST correlations and the results established are consistent with those obtained by Keenlyside and Latif (2007) based on observational data. Furthermore, studies employing atmospheric general circulations models indicate that the atmosphere, with a response in line with the Bjerknes feedback, is sensitive to equatorial Atlantic SST anomalies (Chang et al., 2000; Sutton et al., 2000). The second caveat that should be noted is that the APE evolution analysis conducted within this thesis is based solely on the ROMS-TAtl simulation.

As seen in Chapters 3 and 5, the seasonal cycle in the tropical Atlantic is associated with far larger changes in the climatological background state than in the Pacific. The seasonal cycle in the Atlantic is so dominant that it is able to strongly influence the evolution of its inter-annual variability. In late boreal spring/early boreal summer, the sudden growth of asymmetric conditions about the equator acts to promote the seasonal excitation of a Bjerknes feedback and its associated thermocline mode. However, by late austral summer, seasonal forcing has managed to damp this thermocline mode and re-establish more equatorially symmetric conditions. The results obtained within Chapter 6 suggest that this large seasonal change in the background conditions of the coupled system acts to suppress the influence of remotely forced ocean dynamics within coupled equatorially focused variability between January and April. During this period, weak surface winds ensure the decoupling of subsurface and surface ocean layers and increase the sensitivity to surface heat flux anomalies. As a result of these large seasonal changes, inter-annual coupled variability involving remotely forced thermocline depth variability does not appear to support a free oscillatory mode acting on inter-annual time scales as in the Pacific. Instead, inter-annual variability involving remotely forced thermocline depth variability appears to take the form of inter-annual modulations of a seasonally excited Bjerknes feedback and its associated thermocline mode.

The results obtained within this thesis suggest that the ocean memory mechanism operates on much shorter time scales in the Atlantic than in the Pacific: largely as inter-annual modulations of the seasonally active, delayed, negative feedback response associated with the seasonally excited thermocline mode. This finding has important implications for predictability. We speculate that this difference between the two basins is due to: firstly, the smaller Atlantic basin size which allows the thermocline to adjust to seasonal changes in wind forcing supporting the presence of a seasonally excited thermocline mode - a phenomenon which is seasonally absent from the Pacific; secondly, the fact that the seasonal cycle in the tropical Atlantic is associated with far larger changes in the climatological background state than in the Pacific. This view of the zonal mode, as a modulation of the seasonally active thermocline mode, is consistent with studies suggesting the zonal mode is damped and noise driven (Latif and Barnett, 1995; Nobre et al., 2003; Illig and Dewitte, 2006; Kushnir et al., 2006; Keenlyside and Latif, 2007) as ocean memory is not seen to play a role in sustaining coupled climate oscillations on inter-annual timescales.

This distinct equatorial Atlantic behaviour has implications for the predictability of inter-annual fluctuations associated with the zonal mode. Results obtained in this thesis point to the need for future investigation of the processes behind the anomalous evolution of the seasonally excited Bjerknes feedback in the equatorial Atlantic between April and July.

The recent work of Lübbecke et al. (2010) suggests that inter-annual variability in the strength of the South Atlantic subtropical anticyclone plays a role in the development of anomalous central-eastern basin equatorial Atlantic SSTs during boreal summer months. As mentioned in Chapter 2 the occurrence of a boreal summer zonal mode event is often linked with the occurrence of a Benguela Niño/Niña in the preceding boreal spring months (March-April, Reason et al., 2006; Lübbecke et al., 2010). Lübbecke et al. (2010) refer to combined zonal mode-Benguela Niño warm events as eastern tropical Atlantic Niños which start with a weakening of the south-east trades linked to fluctuations in the strength of the South Atlantic subtropical anticyclone. They suggest that the relaxed trades excite eastward equatorial Kelvin waves which first influence SSTs in the ABA region where the thermocline outcrops in March/April and only later in June/July does the seasonal shallowing of the thermocline allow subsurface anomalies to be reflected in the Atl3 SST field. In view of the results obtained in this thesis, the results obtained by Lübbecke et al. (2010) suggest that variability in the strength of the South Atlantic subtropical anticyclone is potentially an important source of anomalous tropical Atlantic wind power in austral summer and boreal spring. This anomalous wind power could in turn lead to anomalous APE values and hence the anomalous evolution of the seasonally excited Bjerknes feedback as the cold tongue develops during boreal summer months. The relationship between variability in the strength of the South Atlantic subtropical anticyclone and wind work over the tropical Atlantic remains an important topic for future research.

The role of stochastic wind forcing in the anomalous evolution of the seasonally excited Bjerknes feedback and hence zonal mode events is another important factor which deserves further investigation. The results obtained by Marin et al. (2009) (who have compared the development of the cold tongue in the contrasting years of 2005 and 2006) suggest that wind bursts are the cause of inter-annual variability in the timing of the development of the cold tongue. Differences between the two years were found by Marin et al. (2009) to be predominantly the result of a shift in the timing of the development of the cold tongue. The trade winds in the western Atlantic were stronger in April-May of 2005 than April-May 2006 and furthermore an anomalous intra-seasonal intensification of the southeastern trades in mid-May 2005 resulted in the rapid and early cooling of SSTs over a broad region extending from 20°W to the African Coast and from 6°S to the Equator. This study points to the ability of local intra-seasonal wind stress variability in the Gulf of Guinea to initiate the cold tongue season (Marin et al., 2009). Furthermore, Foltz et al. (2003) have shown a link to exist between intra-seasonal wind bursts in the tropical Atlantic (10°N-10°S) and the Madden-Julian Oscillation.

The role of remote ENSO forcing in the development of zonal mode events is yet another key area for future research. As discussed in Chapter 2, the reason for the appearance of anomalously strong trades in the western Atlantic in response to some El Niño events and not others is unclear. Chang et al. (2006b) suggest that the development of anomalously strong trades in response to El Niño events may depend on the nature of SST conditions prior to the El Niño event, stochasticity of the atmosphere and/or variability in the structure and duration of the atmospheric heating associated with individual El Niño events. Due to the predictability associated with ENSO, an improved understanding of the factors which lead to the appearance of anomalous trades in the western equatorial Atlantic in response to some ENSO events will potentially enhance the predictability of zonal mode events.

Finally, the decay of meridional mode events in April coincides with the onset period of zonal mode events. Servain et al. (1999, 2000) suggest that the zonal and meridional modes are related by the fact that they are both a result of anomalous fluctuations in the trade wind system associated with latitudinal displacements

of the ITCZ. The relationship between these two modes remains a important area for future research as an understanding of how conditions associated with the decay of the meridional mode in boreal spring will impact on the development of zonal mode events in boreal summer represents a potential source of predictability.

References

- Arakawa, A. and V. Lamb. 1977. "Computational design of the basic dynamical processes of the UCLA general circulation model." *Method Comput. Phys.* 17:173–265.
- Barreiro, M., P. Chang, L. Ji, R. Saravanan, and A. Giannini. 2005. "Dynamical elements of predicting boreal spring tropical Atlantic sea-surface temperatures." *Dyn. Atmos. Oceans.* 39:61–85.
- Battisti, D. S. 1988. "Dynamics and thermodynamics of a warming event in a coupled tropical atmosphere-ocean model." *J. Atmos. Sci.* 45:2889–2919.
- Battisti, D. S. and A. C. Hirst. 1989. "Interannual variability in a tropical atmosphere-ocean model: Influence of the basic state, ocean geometry and nonlinearity." *J. Atmos. Sci.* 46:1687–1712.
- Bjerknes, J. 1969. "Atmospheric teleconnections from the equatorial Pacific." *Mon. Weather Rev.* 97:163–172.
- Boccaletti, G., R. C. Pacanowski, S. G. H. Philander, and A. V. Fedorov. 2004. "The Thermal Structure of the Upper Ocean." *J. Phys. Oceanogr.* 34:888–902.
- Boyer, D. C., H. J. Boyer, I. Fossen, and A. Kreiner. 2001. "Changes in abundance of the northern Benguela sardine stock during the decade 1990-2000, with comments on the relative importance of fishing and the environment." *S. Afr. J. Mar. Sci.* 23:67–84.
- Brainerd, K. E. and M. C. Gregg. 1995. "Surface mixed and mixing layer depths." *Deep-Sea Res.* 42A(9):1521–1543.
- Brandt, P., G. Caniaux, B. Bourlès, A. Lazar, M. Dengler, A. Funk, V. Hormann, H. Giordani, and F. Marin. 2010. "Equatorial upper-ocean dynamics and their interaction with the West African monsoon." *Atmospheric Science Letters*. In press.
- Brown, J. N. and A. V. Fedorov. 2010. "How much energy is transferred from the winds to the thermocline on ENSO timescales?" *J. Climate.* 23(6):1563–1580.
- Bunge, L. and A. J. Clarke. 2009. "Seasonal Propagation of Sea Level along the Equator in the Atlantic." *J. Phys. Oceanogr.* 39:1069–1073.
- Byrne, D. A., D. L. Witter, D. R. Watts, N. R. Pettigrew, C. M. Duncombe Rae, and S. BakerYeboah. 2006. "Interocean heat and salt transports from the Agulhas Leakage: First results from ASTTEX." In *Eos Trans. AGU, 87(36), Ocean Sci. Meet. Suppl., Abstract OS22C-03*.

- Cane, M. A. 1979. "The response of an equatorial ocean to simple wind-stress patterns, I, Model formulations and analytic results." *J. Marine. Res.* 37:233–252.
- Cane, M. A. and D. W. Moore. 1981. "A Note on Low-Frequency Equatorial Basin Modes." *J. Phys. Oceanogr.* 11:1578–1584.
- Carton, J. A., G. A. Chepurin, X. Cao, and B. S. Giese. 2000. "A Simple Ocean Data Assimilation analysis of the global upper ocean 1950-1995 part I: Methodology." *J. Phys. Oceanogr.* 30:294–309.
- Carton, J. A. and B. S. Giese. 2008. "A Reanalysis of Ocean Climate Using Simple Ocean Data Assimilation (SODA)." *Mon. Weather Rev.* 136:2999–3017.
- Carton, J. A. and B. Huang. 1994. "Warm events in the tropical Atlantic." *J. Phys. Oceanogr.* 24:888–903.
- Carton, J. A. and Z. X. Zhou. 1997. "Annual cycle of sea surface temperature in the tropical Atlantic Ocean." *J. Geophys. Res.* 102:27813–27824.
- Chang, P. 1993. "Seasonal cycle of sea surface temperature and mixed layer heat budget in the tropical Pacific Ocean." *Geophys. Res. Letters.* 20:2079–2082.
- Chang, P. 1994. "A study of seasonal cycle of sea surface temperature in the tropical Pacific Ocean using reduced gravity models." *J. Geophys. Res.* 99:7725–7741.
- Chang, P., Y. Fang, R. Saravanan, L. Ji, and H. Seidel. 2006b. "The cause of the fragile relationship between the Pacific El Niño and the Atlantic Niño." *Nature.* 443:324–328.
- Chang, P. and G. Philander. 1994. "A coupled ocean-atmosphere instability of relevance to the seasonal cycle." *J. Atmos. Sci.* 51:3627–3648.
- Chang, P., R. Saravanan, L. Ji, and G. C. Hegerl. 2000. "The Effect of Local Sea Surface Temperatures on Atmospheric Circulation over the Tropical Atlantic Sector." *J. Climate.* 13:2195–2216.
- Chang, P., T. Yamagata, P. Schopf, S. K. Behera, J. Carton, W. S. Kessler, G. Meyers, T. Qu, F. Schott, S. Shetye, and S.-P. Xie. 2006a. "Climate Fluctuations of Tropical Coupled Systems - The Role of Ocean Dynamics." *J. Climate.* 19:5122–5174.
- Chiang, J. C. H. and B. R. Lintner. 2005. "Mechanisms of remote tropical surface warming during El Niño." *J. Climate.* 18:4130–4149.
- Chiang, J. C. H. and A. H. Sobel. 2002. "Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate." *J. Climate.* 15:2616–2631.
- Colberg, F., C. J. C. Reason, and K. Rodgers. 2004. "South Atlantic response to El Niño-Southern Oscillation induced climate variability in an ocean general circulation model." *J. Geophys. Res.* 109(C12), C12015, doi:10.1029/2004JC002301.
- DaSilva, A. M., C. C. Young, and S. Levitus. 1994. "Atlas of surface marine data 1994, Vol. 1, Algorithms and Procedures." Tech. rep., NOAA Atlas NESDIS 6, U.S. Department of Commerce, 74pp.
- Delecluse, P., J. Servain, C. Levy, K. Arpe, and L. Bengtsson. 1994. "On the connection between the 1984 Atlantic warm event and the 1982-1983 ENSO." *Tellus.* 46A:448–464.

- Ding, H., N. S. Keenlyside, and M. Latif. 2009. "Seasonal cycle in the upper equatorial Atlantic Ocean." *J. Geophys. Res.* 114(C9), C09016, doi:10.1029/2009JC005418.
- Enfield, D. B. and D. A. Mayer. 1997. "Tropical Atlantic SST variability and its relation to El Niño-Southern Oscillation." *J. Geophys. Res.* 102:929–945.
- Fedorov, A. V. 2007. "Net Energy Dissipation Rates in the Tropical Ocean and ENSO Dynamics." *J. Climate.* 20:1108–1117.
- Fedorov, A. V., S. L. Harper, S. G. H. Philander, B. Winter, and A. Wittenberg. 2003. "How predictable is El Niño?" *Bull. Amer. Meteor. Soc.* 84:911–919.
- Florenchie, P., J. R. E. Lutjeharms, C. J. C. Reason, S. Masson, and M. Rouault. 2003. "The source of Benguela Niños in the South Atlantic Ocean." *Geophys. Res. Letters.* 30(10), 1505, doi:10.1029/2003GL017172.
- Florenchie, P., C. J. C. Reason, J. R. E. Lutjeharms, M. Rouault, C. Roy, and S. Masson. 2004. "Evolution of Interannual Warm and Cold Events in the Southeast Atlantic Ocean." *J. Climate.* 17:2318–2334.
- Foltz, G. R., S. A. Grodsky, J. A. Carton, and M. J. McPhaden. 2003. "Seasonal mixed layer heat budget of the tropical Atlantic Ocean." *J. Geophys. Res.* 108(C5), 3146, doi:10.1029/2002JC001584.
- Frankignoul, C. and K. Hasselmann. 1977. "Stochastic climate models. Part II: Application to SST anomalies and thermocline variability." *Tellus.* 29:289–305.
- Fratantoni, D. M., W. E. Johns, T. L. Townsend, and H. E. Hurlburt. 2000. "Low-latitude circulation and mass transport pathways in a model of the tropical Atlantic ocean." *J. Phys. Oceanogr.* 30:1944–1966.
- Gill, A. E. 1982. *Atmosphere-Ocean Dynamics*. New York: Academic Press.
- Goddard, L. and S. G. H. Philander. 2000. "The energetics of El Niño and La Niña." *J. Climate.* 13:1496–1516.
- Hayes, S. P., P. Chang, and M. J. McPhaden. 1991. "Variability of the sea surface temperature in the eastern equatorial Pacific during 1986-1988." *J. Geophys. Res.* 96:10553–10566.
- Hirst, A. C. 1986. "Unstable and damped equatorial modes in simple coupled ocean-atmosphere models." *J. Atmos. Sci.* 43:606–632.
- Hisard, P. 1980. "Observation de réponses de type "El Niño" dans l'Atlantique tropical oriental, Golfe de Guinée." *Oceanol. Acta.* 3:69–78.
- Holliday, D. and M. E. McIntyre. 1981. "On potential energy density in an incompressible stratified fluid." *J. Fluid. Mech.* 107:221–225.
- Huang, R. X. 1998. "Mixing and available potential energy in a Boussinesq ocean." *J. Phys. Oceanogr.* 28:669–678.
- Illig, S. and B. Dewitte. 2006. "Local Coupled Equatorial Variability versus Remote ENSO Forcing in an Intermediate Coupled Model of the Tropical Atlantic." *J. Climate.* 19:5227–5252.

- Jackett, D. R. and T. J. McDougall. 1995. "Minimal Adjustment of Hydrostatic Profiles to Achieve Static Stability." *J. Atmos. Ocean. Tech.* 12:381–389.
- Jin, F. 1997. "An Equatorial Ocean Recharge Paradigm for ENSO. Part I: Conceptual Model." *J. Atmos. Sci.* 54:811–829.
- Johns, W. E., T. N. Lee, R. C. Beardsley, J. Candela, R. Limeburner, and B. Castro. 1998. "Annual Cycle and Variability of the North Brazil Current." *J. Phys. Oceanogr.* 28:103–128.
- Kanamitsu, M., W. Ebisuzaki, J. Woollen, S.-K. Yang, J. J. Hnilo, M. Fiorino, and G. L. Potter. 2002. "NCEP-DOE AMIP-II Reanalysis (R-2)." *Bull. Amer. Meteor. Soc.* 83:1631–1643.
- Katz, E. J. 1981. "Dynamic topography of the sea surface in the equatorial Atlantic." *J. Marine. Res.* 39:53–63.
- Katz, E. L. 1987. "Seasonal response of the sea surface to the wind in the equatorial Atlantic." *J. Geophys. Res.* 92:1885–1893.
- Keenlyside, N. S. and M. Latif. 2007. "Understanding Equatorial Atlantic Interannual Variability." *J. Climate.* 20:131–142.
- Klein, S.A. and D.L. Hartmann. 1993. "The seasonal cycle of low stratiform clouds." *J. Climate.* 6(8):1587–1606.
- Kondo, J. 1975. "Air-sea bulk transfer coefficients in diabatic conditions." *Boundary-Layer Meteorology* 9(1):91–112.
- Kraus, E. B. 1977. *Modelling and Prediction of the Upper Layers of the Ocean*. Oxford: Pergamon Press.
- Kushnir, Y., W. A. Robinson, P. Chang, and A. W. Robertson. 2006. "The Physical Basis for Predicting Atlantic Sector Seasonal-to-Interannual Climate Variability." *J. Climate.* 19:5949–5970.
- Large, W. G., J. C. McWilliams, and S. C. Doney. 1994. "Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization." *Rev. of Geophys.* 32:363–403.
- Latif, M. and T. P. Barnett. 1995. "Interactions of the tropical oceans." *J. Climate.* 8:952–964.
- Latif, M. and A. Grötzner. 2000. "The equatorial Atlantic oscillation and its response to ENSO." *Climate Dynamics.* 16:213–218.
- Lee, S.-K. and G. T. Csanady. 1999. "Warm water formation and escape in the upper tropical Atlantic Ocean 2. A numerical model study." *J. Geophys. Res.* 104:29573–29590.
- Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, and H. E. Garcia. 2006. "World Ocean Atlas 2005, Volume 1: Temperature." Tech. rep., S. Levitus, Ed. NOAA Atlas NESDIS 61, U.S. Gov. Printing Office, Washington, D.C., 182 pp.
URL <http://www.nodc.noaa.gov/OC5/indprod.html>.
- Lorbacher, K., D. Dommenges, P.P. Niiler, and A. Kohl. 2006. "Ocean mixed layer depth: A subsurface proxy of ocean-atmosphere variability." *J. Geophys. Res.* 111(C7), C07010, doi:10.1029/2003JC002157.

- Lorenz, E. N. 1955. "Available potential energy and the maintenance of the general circulation." *Tellus*. 7:157–167.
- Lübbecke, J. F., C. W. Böning, N. S. Keenlyside, and S. P. Xie. 2010. "On the connection between Benguela and Equatorial Atlantic Niños and the role of the South Atlantic Anticyclone." *J. Geophys. Res.* in press.
- Lumpkin, R. and Z. Garraffo. 2005. "Evaluating the Decomposition of Tropical Atlantic Drifter Observations." *J. Atmos. Ocean. Tech.* 22:1403–1415.
- Lumpkin, R. and S. L. Garzoli. 2005. "Near-surface circulation in the Tropical Ocean." *Deep-Sea Res.* 52:495–518.
- Marchesiello, P., L. Debreu, and X. Couvelard. 2009. "Spurious diapycnal mixing in terrain-following coordinate models: The problem and a solution." *Ocean. Mod.* 26:156–169.
- Marchesiello, P., J. C. McWilliams, and A. Shchepetkin. 2001. "Open boundary conditions for long-term integration of regional oceanic models." *Ocean. Mod.* 3:1–20.
- Marin, F., G. Caniaux, H. Giordani, B. Bourlés, Y. Gouriou, and E. Key. 2009. "Why Were Sea Surface Temperatures so Different in the Eastern Equatorial Atlantic in June 2005 and 2006?" *J. Phys. Oceanogr.* 39:1416–1431.
- McCreary, J. P. Jr. and P. Lu. 1994. "Interaction between the subtropical and the equatorial oceans: The subtropical cell." *J. Phys. Oceanogr.* 24:466–497.
- McPhaden, M. J., A. J. Busalacchi, R. Cheney, J. Donguy, K. S. Gage, D. Halpern, M. Ji, P. Julian, G. Meyers, G. T. Mitchum, P. P. Niiler, J. Picaut, R. W. Reynolds, N. Smith, and K. Takeuchi. 1998. "The Tropical Ocean-Global Atmosphere observing system: A decade of progress." *J. Geophys. Res.* 103:14169–14240.
- Menkes, C.E.R., J.G. Vialard, S.G. Kennan, J.P. Boulanger, and G.V. Madec. 2006. "A modeling study of the impact of tropical instability waves on the heat budget of the eastern equatorial Pacific." *J. Phys. Oceanogr.* 36:847–865.
- Merle, J. 1980a. "Seasonal Heat Budget in the Equatorial Atlantic Ocean." *J. Phys. Oceanogr.* 10:464–469.
- Merle, J. 1980b. "Variabilité thermique annuelle et interannuelle de l'océan Atlantique équatorial Est. L'hypothèse d'un "El Niño" Atlantique." *Oceanol. Acta.* 3:209–220.
- Merle, J. and S. Arnault. 1985. "Seasonal variability of the dynamic topography in the tropical Atlantic Ocean." *J. Marine. Res.* 43:267–288.
- Mitchell, T. P. and J. M. Wallace. 1992. "The annual cycle in equatorial convection and sea surface temperature." *J. Climate.* 5:1140–1156.
- Molemaker, M. J. and J. C. McWilliams. 2010. "Local balance and cross-scale flux of available potential energy." *J. Fluid. Mech.* 645:295–314.

- Molines, J. M., B. Barnier, T. Penduff, L. Brodeau, A.M. Treguier, S. Theetten, and G. Madec. 2006. "Definition of the interannual experiment ORCA025-G70, 1958-2004." Tech. Rep. Report LEGI-DRA-2-11-2006, LEGI, 34pp.
URL <http://www.ifremer.fr/lpo/drakkar>
- Murtugudde, R. G., J. Ballabrera-Poy, J. Beauchamp, and A. J. Busalacchi. 2001. "Relationship between zonal and meridional modes in the tropical Atlantic." *Geophys. Res. Letters*. 23(23):4463–4466, doi:10.1029/2001GL013407.
- Neelin, J. D. 1991. "The slow sea surface temperature mode and the fast-wave limit: Analytic theory for tropical interannual oscillations and experiments in a hybrid coupled model." *J. Atmos. Sci.* 48:584–606.
- Neelin, J. D., D. S. Battisti, A. C. Hirst, F. Jin, Y. Wakata, T. Yamagata, and S. E. Zebiak. 1998. "ENSO theory." *J. Geophys. Res.* 103(C7):14261–14290.
- Nobre, P., S. E. Zebiak, and B. P. Kirtman. 2003. "Local and remote sources of tropical atlantic variability as inferred from the results of a hybrid ocean-atmosphere coupled model." *Geophys. Res. Letters*. 30(5), 8008, doi:10.1029/2002GL015785.
- Okumura, Y. and S.-P. Xie. 2004. "Interaction of the Atlantic Equatorial Cold Tongue and the African Monsoon." *J. Climate*. 17:3589–3602.
- Okumura, Y. and S.-P. Xie. 2006. "Some overlooked features of tropical Atlantic climate leading to a new Niño-like phenomenon." *J. Climate*. 19:5859–5874.
- Okumura, Y., S.-P. Xie, A. Numaguti, and Y. Tanimoto. 2001. "Tropical Atlantic air-sea interaction and its influence on the NAO." *Geophys. Res. Letters*. 28:1507–1510.
- Oort, A. H., S. C. Ascher, S. Levitus, and J. P. Peixoto. 1989. "New estimates of the available potential energy in the world ocean." *J. Geophys. Res.* 94:3187–3200.
- du Penhoat, Y. and A. M. Treguier. 1985. "The seasonal linear response of the tropical Atlantic ocean." *J. Phys. Oceanogr.* 15:316–329.
- Penven, P., P. Marchesiello, L. Debreu, and J. Lefèvre. 2008. "Software tools for pre- and post-processing of oceanic regional simulations." *Environmental Modelling & Software*. 23:660–662.
- Peter, A., M. Le Henaff, Y. Penhoat Y. du Penhoat, C.E. Menkes, F. Marin, J. Vialard, G. Caniaux, and A. Lazar. 2006. "A model study of the seasonal mixed layer heat budget in the equatorial Atlantic." *J. Geophys. Res.* 111, C06014, doi:10.1029/2005JC003157.
- Philander, S. G. H. 1990. *El Niño, La Niña, and the Southern Oscillation*. San Diego: Academic Press Inc.
- Philander, S. G. H., D. Gu, D Halpern, G Lambert, N.-C. Lau, T. Li, and R. C. Pacanowski. 1996. "Why the ITCZ Is Mostly North of the Equator?" *J. Climate*. 9:2958–2972.
- Philander, S. G. H. and R. C. Pacanowski. 1981. "Response of equatorial oceans to periodic forcing." *J. Geophys. Res.* 86:1903–1916.

- Philander, S. G. H. and R. C. Pacanowski. 1986a. "A model of the seasonal cycle in the tropical Atlantic Ocean." *J. Geophys. Res.* 91:14192–14206.
- Philander, S. G. H. and R. C. Pacanowski. 1986b. "The mass and heat budgets in a model of the tropical Atlantic." *J. Geophys. Res.* 91:14212–14220.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan. 2003. "Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century." *J. Geophys. Res.* 108(D14), 4407, doi:10.1029/2002JD002670.
- Reason, C.J.C., P. Florenchie, M. Rouault, and J. Veitch. 2006. "Influences of Large Scale Climate Modes and Agulhas System Variability on the BCLME Region." In *Benguela - Predicting a Large Marine Ecosystem*, Vere Shannon, Gotthilf Hempel, Paola Malanotte-Rizzoli, Coleen Moloney, and John Woods, eds., vol. 14 of *Large Marine Ecosystems*, pages 223 – 238, Elsevier.
- Reid, R., B. Elliot, and D. Olson. 1981. "Available potential energy: A clarification." *J. Phys. Oceanogr.* 11:15–29.
- Rio, H-M., P. Schaeffer, G. Moreaux, J-M. Lemoine, and E. Bronner. 2009. "A new Mean Dynamic Topography computed over the global ocean from GRACE data, altimetry and in-situ measurements.", poster communication at OceanObs09 symposium 21-25 September 2009 Venice.
URL http://www.aviso.oceanobs.com/fileadmin/documents/data/products/auxiliary/MDT_Cnes-CLS09_poster_Oceanobs09.pdf
- Rouault, M., P. Florenchie, N. Fauchereau, and C. J. C. Reason. 2003. "South East tropical Atlantic warm events and southern African rainfall." *Geophys. Res. Letters*. 30(5), 8009, doi:10.1029/2002GL014840.
- Rouault, M., S. Illig, C. Bartholomae, C. J. C. Reason, and A. Bentamy. 2007. "Propagation and origin of warm anomalies in the Angola Benguela upwelling system in 2001." *J. Marine. Sys.* 68:473–488.
- Rouault, M., J. Servain, C. J. C. Reason, B. Boulès, M. J. Rouault, and N. Fauchereau. 2009. "Extension of PIRATA in the tropical South-East Atlantic: an initial one-year experiment." *Afr. J. Mar. Sci.* 31:63–71.
- Rowell, D. P. 1998. "Assessing Potential Seasonal Predictability with an Ensemble of Multidecadal GCM Simulations." *J. Climate*. 11:109–120.
- Ruiz-Barradas, A., J. A. Carton, and S. Nigam. 2000. "Structure of interannual-to-decadal climate variability in the tropical Atlantic sector." *J. Climate*. 13:3285–3297.
- Schouten, M. W., R. P. Matano, and T. P. Strub. 2005. "A description of the seasonal cycle of the equatorial Atlantic from altimeter data." *Deep-Sea Res.* 52:477–493.
- Seager, R., S.E. Zebiak, and M.A. Cane. 1988. "A model of the tropical Pacific sea surface temperature climatology." *J. Geophys. Res.* 93(C2):1265–1280.
- Servain, J., H. L. Ayina, and H. Roquet. 2000. "The relationship between the simulated climatic variability modes of the tropical Atlantic." *Int. J. Climatol.* 20:939–953.
- Servain, J., A. Busalacchi, M. McPhaden, A. Moura, G. Reverdin, M. Vianna, and S. Zebiak. 1998. "A Pilot Research Moored Array in the Tropical Atlantic (PIRATA)." *Bull. Amer. Meteor. Soc.* 79:2019–2031.

- Servain, J., J. Picaut, and J. Merle. 1982. "Evidence of remote forcing in the equatorial Atlantic Ocean." *J. Phys. Oceanogr.* 12:457–463.
- Servain, J., I. Wainer, J. P. McCreary, and A. Dessier. 1999. "Relationship between the equatorial and meridional modes of climatic variability in the tropical Atlantic." *Geophys. Res. Letters*. 26(4):485–488, doi:10.1029/1999GL900014.
- Shannon, L. V., A. J. Boyd, G. B. Brundrit, and J. Taunton-Clark. 1986. "On the existence of an El Niño-type phenomenon in the Benguela system." *J. Marine. Sys.* 44:495–520.
- Shchepetkin, A. and J. McWilliams. 1998. "Quasi-Monotone Advection Schemes Based on Explicit Locally Adaptive Dissipation." *Mon. Weather Rev.* 126:1541–1580.
- Shchepetkin, A. and J. McWilliams. 2003. "A method for computing horizontal pressure-gradient force in an oceanic model with a nonaligned vertical coordinate." *J. Geophys. Res.* 108(C3), 3090, doi:10.1029/2001JC001047.
- Shchepetkin, A. and J. McWilliams. 2005. "The regional oceanic modeling system (ROMS): a split-explicit, free-surface, topography-following-coordinate oceanic model." *Ocean. Mod.* 9:347–404.
- Song, Y. and D. Haidvogel. 1994. "A semi-implicit ocean circulation model using a generalized topography following coordinate system." *J. Comput. Phys.* 115:228–244.
- Sterl, A. and W. Hazeleger. 2003. "Coupled variability and air-sea interaction in the South Atlantic Ocean." *Climate Dynamics*. 21:559–571.
- Suarez, M. J. and P. S. Schopf. 1988. "A delayed action oscillator for ENSO ." *J. Atmos. Sci.* 45:3283–3287.
- Sutton, R. T., S. P. Jewson, and D. P. Rowell. 2000. "The elements of climate variability in the tropical Atlantic region." *J. Climate*. 13:3261–3284.
- Swenson, M. S. and D. V. Hansen. 1999. "Tropical Pacific Ocean Mixed Layer Heat Budget: The Pacific Cold Tongue." *J. Phys. Oceanogr.* 29:69–81.
- Tanimoto, Y. and S.-P. Xie. 2002. "Inter-hemispheric decadal variations in SST, surface wind, heat flux and cloud cover over the Atlantic Ocean." *J. Meteor. Soc. Japan*. 80:1199–1219.
- Trenberth, K. E., G. W. Branstator, D. Karoly, A. Kumar, N.-C. Lau, and C. Ropelewski. 1998. "Progress during TOGA in understanding and modeling global teleconnections associated with tropical sea surface temperatures." *J. Geophys. Res.* 103(C7):14291–14324, doi:10.1029/97JC01444.
- Vauclair, F. and Y. du Penhoat. 2001. "Interannual variability of the upper layer of the tropical Atlantic Ocean from in situ data between 1979 and 1999." *Climate Dynamics*. 17:527–546.
- Vauclair, F., Y. du Penhoat, and G. Reverdin. 2004. "Heat and Mass Budgets of the Warm Upper Layer of the Tropical Atlantic Ocean in 1979–99." *J. Phys. Oceanogr.* 34:903–919.
- Vialard, J. and P. Delecluse. 1998a. "An OGCM Study for the TOGA Decade. Part I: Role of Salinity in the Physics of the Western Pacific Fresh Pool." *J. Phys. Oceanogr.* 28:1071–1088.

- Vialard, J. and P. Delecluse. 1998b. "An OGCM study for the TOGA decade. Part II: Barrier-layer formation and variability." *J. Phys. Oceanogr.* 28:1089–1106.
- Wallace, J. M., E. M. Rasmusson, T. P. Mitchell, V. E. Kousky, E. S. Sarachik, and H. von Storch. 1998. "On the structure and evolution of ENSO-related climate variability in the tropical Pacific: Lessons from TOGA." *J. Geophys. Res.* 103(C7):14241–14259, doi:10.1029/97JC02905.
- Wang, C. and D. B. Enfield. 2001. "The tropical Western Hemisphere warm pool." *Geophys. Res. Letters.* 28(8):1635–1638, doi:10.1029/2000GL011763.
- Wang, W. and M. J. McPhaden. 1999. "The Surface-Layer Heat Balance in the Equatorial Pacific Ocean. Part I: Mean Seasonal Cycle." *J. Phys. Oceanogr.* 29:1812–1931.
- Weijer, W., W. P. M. De Ruijter, A. Sterl, and S. S. Drijfhout. 2002. "Response of the Atlantic overturning circulation to South Atlantic sources of buoyancy." *Global and Planetary Change.* 34(3-4):293–311.
- Weingartner, T. J. and R. H. Weisberg. 1991. "A Description of the Annual Cycle in Sea Surface Temperature and Upper Ocean Heat in the Equatorial Atlantic." *J. Phys. Oceanogr.* 21:83–96.
- Winters, K. B., P. N. Lombard, J. J. Riley, and E. A. D'Asaro. 1995. "Available potential energy and mixing in density-stratified fluids." *J. Fluid. Mech.* 289:115–128.
- Xie, S.-P. and J. A. Carton. 2004. "Tropical Atlantic variability: Patterns, mechanisms, and impacts." *Earth Climate: The Ocean-Atmosphere Interaction, Geophys. Monogr.* 147:121–142.
- Xie, S.-P. and S. G. H. Philander. 1994. "A coupled ocean-atmosphere model of relevance to the ITCZ in the eastern Pacific." *Tellus.* 46A:340–350.
- Yu, L., X. Jin, and R. A. Weller. 2006. "Role of net surface heat flux in seasonal variations of sea surface temperature in the tropical Atlantic Ocean." *J. Climate.* 19:6153–6169.
- Yu, X. and M. J. McPhaden. 1999. "Seasonal Variability in the Equatorial Pacific." *J. Phys. Oceanogr.* 29:925–947.
- Zebiak, S. E. 1993. "Air-sea interaction in the equatorial Atlantic region." *J. Climate.* 6:1567–1586.
- Zebiak, S. E. and M. Cane. 1987. "A model El Niño-Southern Oscillation." *Mon. Weather Rev.* 115:2262–2278.

Appendix A

ORCA Figures

University of Cape Town

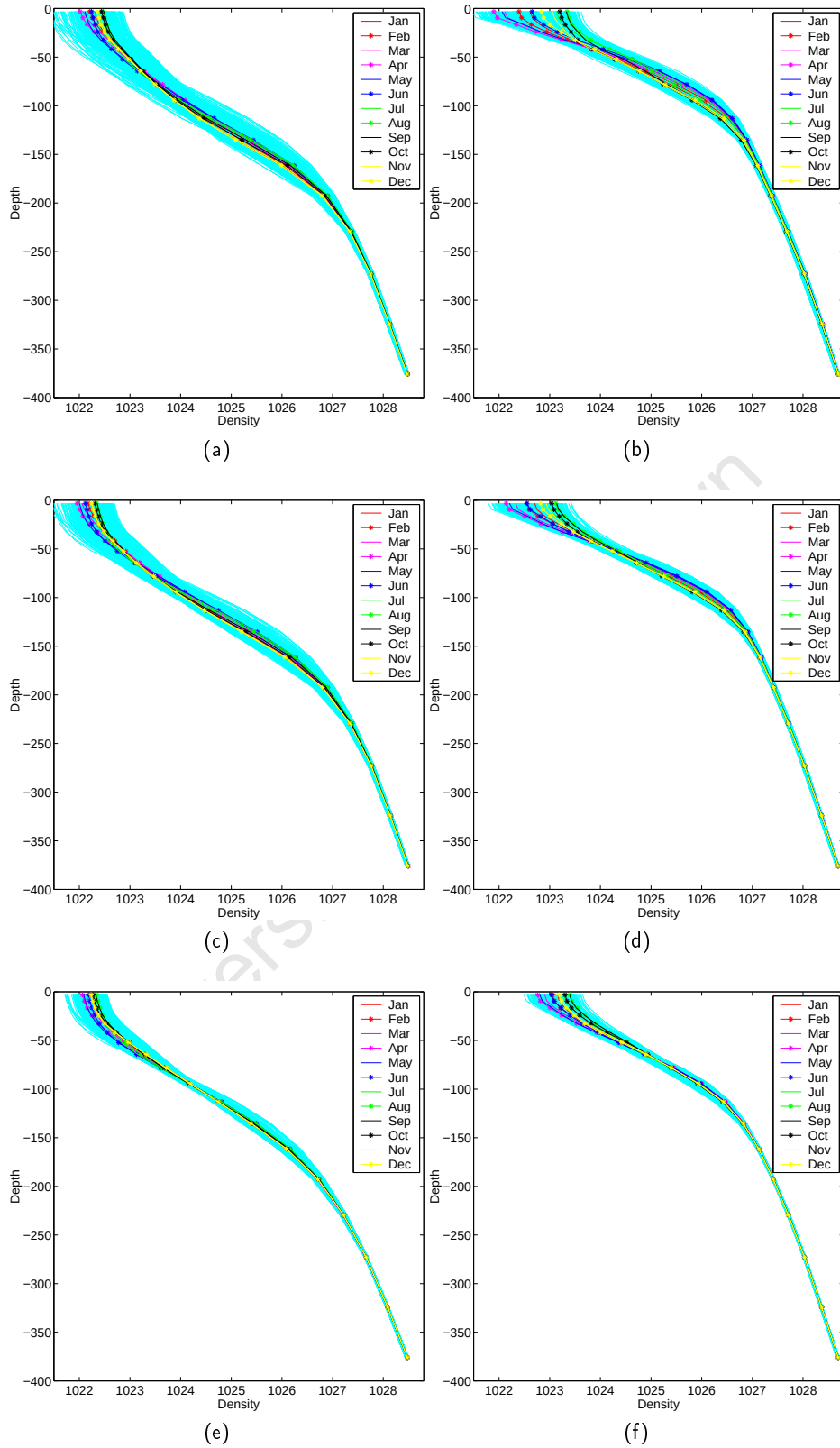


Figure A.1: Variability in the vertical density profile estimated by horizontally averaging the monthly mean density field within the chosen domain: (a) the equatorial Pacific (5°N-5°S 130°E-85°W), (b) the equatorial Atlantic (3°S-3°N 60°W-15°E), (c) the tropical Pacific (8°N-8°S 130°E-85°W), (d) the tropical Atlantic (8°S-8°N 60°W-15°E), (e) the extended tropical Pacific (15°N-15°S 130°E-85°W), and (f) the extended tropical Atlantic (15°S-15°N 60°W-15°E). Seasonal variability in the vertical density profile is overlayed on the total spread of vertical density profiles between 1958-2004 shown in cyan. Depth (m) is represented on the y-axis and density (kgm^{-3}) on the x-axis. Monthly density fields are derived from the monthly ORCA temperature and salinity fields.

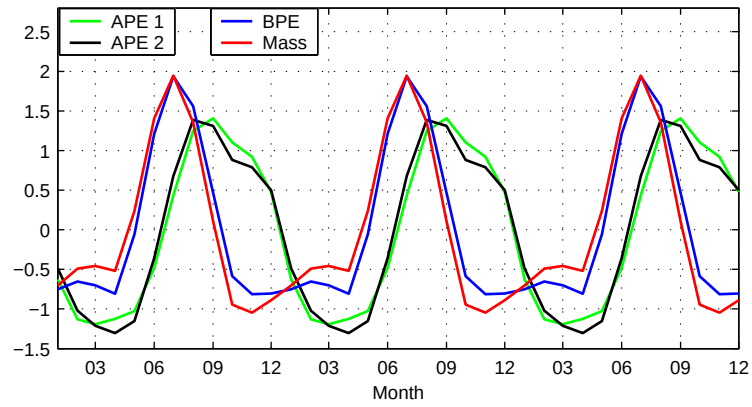


Figure A.2: Seasonal changes in equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE obtained using Equation 3.3, APE and BPE obtained using Equation 3.2, as well as seasonal changes in the Mass of this domain. Mass, BPE and APE values have been normalised. Values calculated using ORCA data.

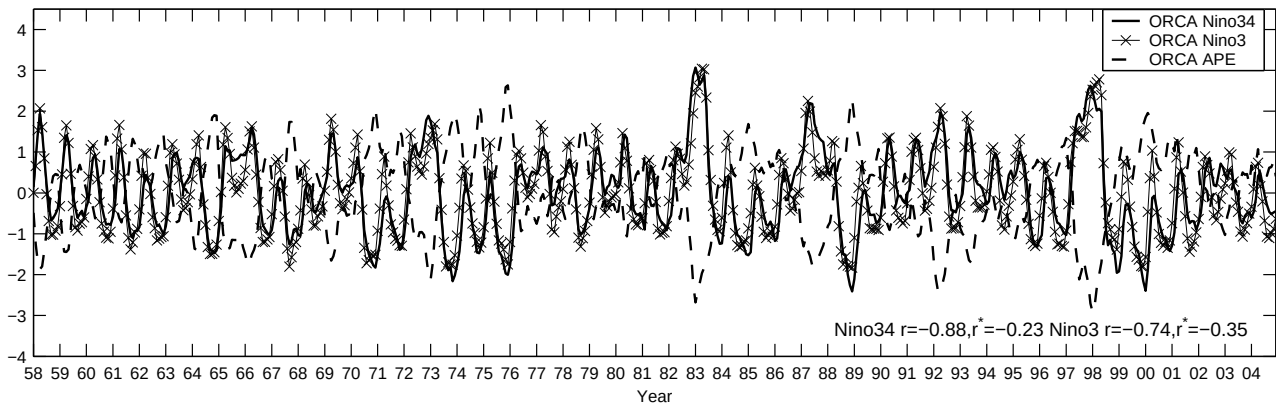


Figure A.3: A comparison between total Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE variability. SST and APE values were calculated using ORCA data. The correlation between each SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended.

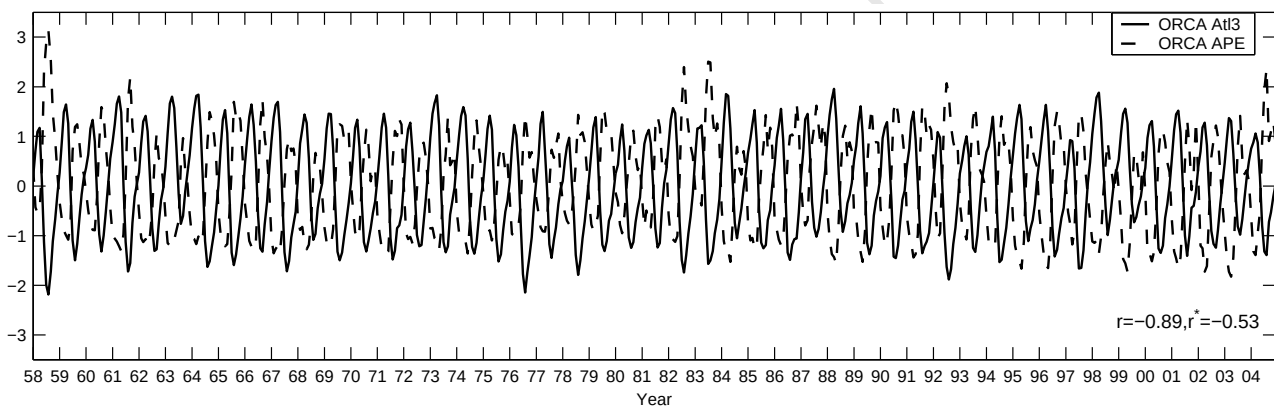


Figure A.4: A comparison between total Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE variability. SST and APE values were calculated using ORCA data. The correlation between the Atl3 SST index and APE values (r) is give on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised and detrended.

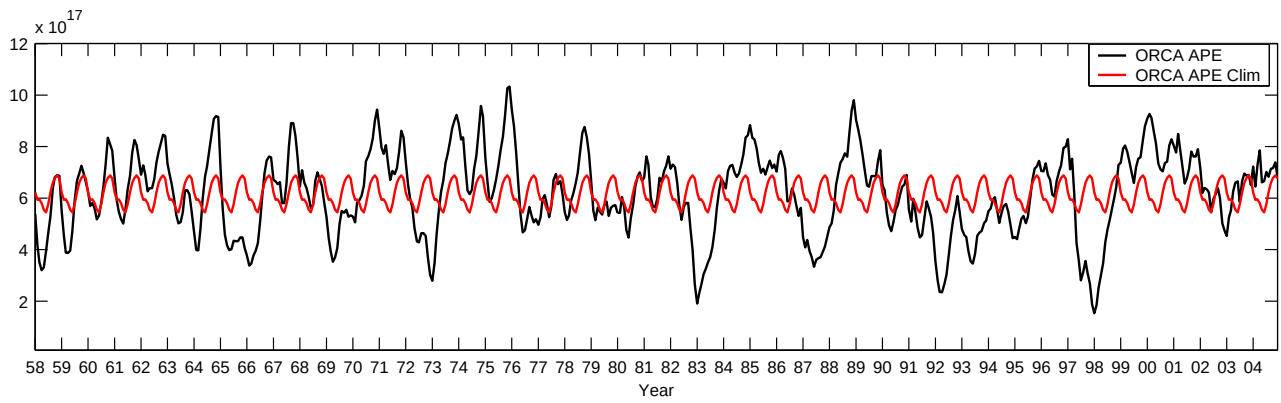


Figure A.5: A comparison between total and climatological equatorial Pacific (5°N - 5°S 130°E - 85°W 0-400m) APE values. APE values were calculated using ORCA data and have been detrended.

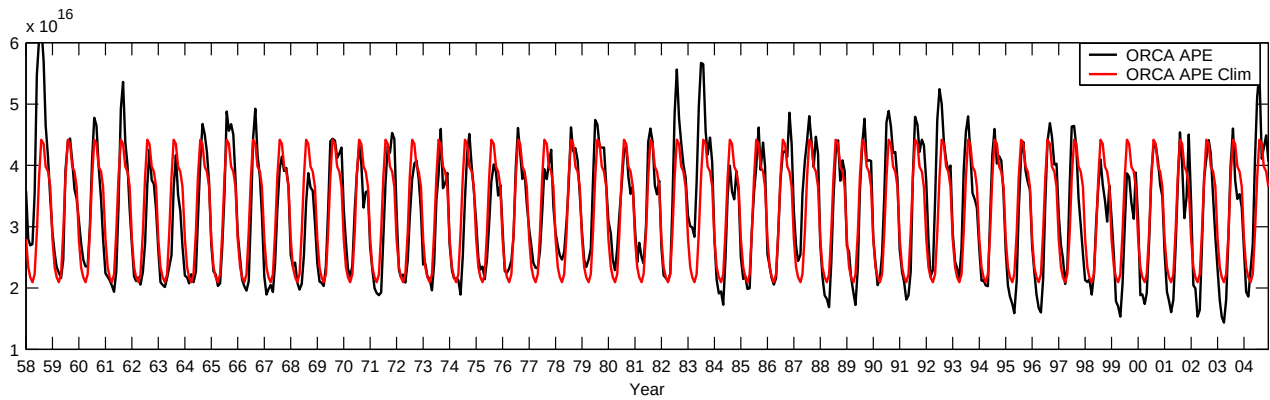
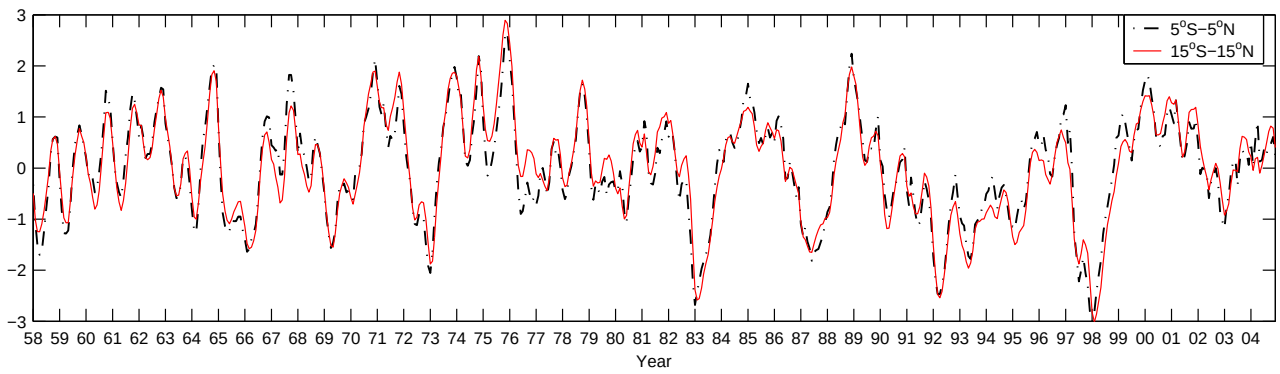
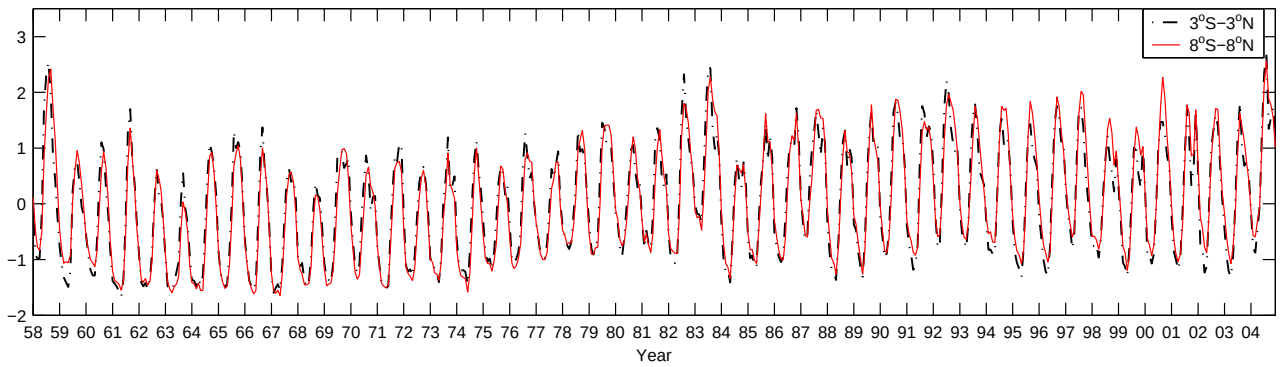


Figure A.6: A comparison between total and climatological equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE values. APE values were calculated using ORCA data and have been detrended.



(a)



(b)

Figure A.7: (a) Equatorial Pacific (5°N - 5°S 130°E - 85°W 0-400m) APE versus tropical Pacific (15°N - 15°S 130°E - 85°W 0-400m) APE. (b) Equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE versus tropical Atlantic (8°S - 8°N 60°W - 15°E 0-400m) APE. APE values were calculated using ORCA data.

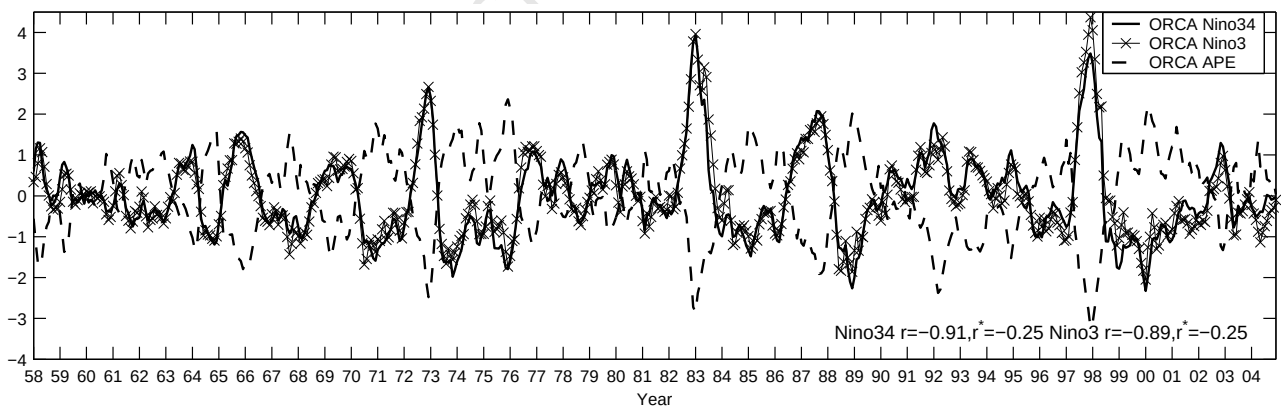


Figure A.8: A comparison between inter-annual Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N - 5°S 130°E - 85°W 0-400m) APE anomalies. SST and APE anomalies were calculated using ORCA data. The correlation between anomalies in each SST index and APE (r) is given on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended.

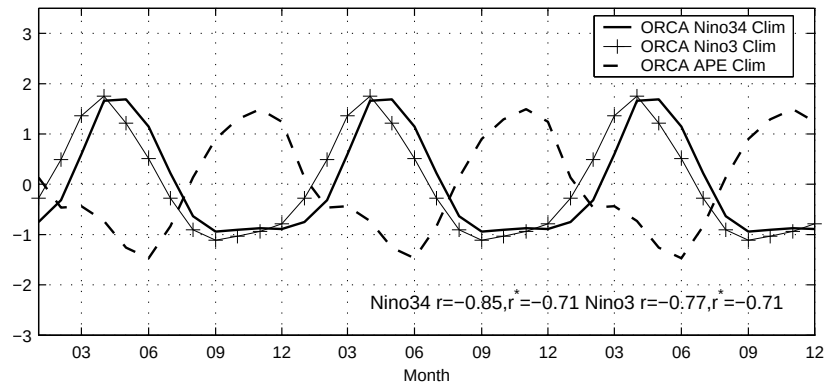


Figure A.9: A comparison between climatological Niño 3.4 SST, Niño 3 SST and equatorial Pacific (5°N-5°S 130°E-85°W 0-400m) APE. SST and APE values were calculated using ORCA data. The correlation between each SST index and APE values (r) is given on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised.

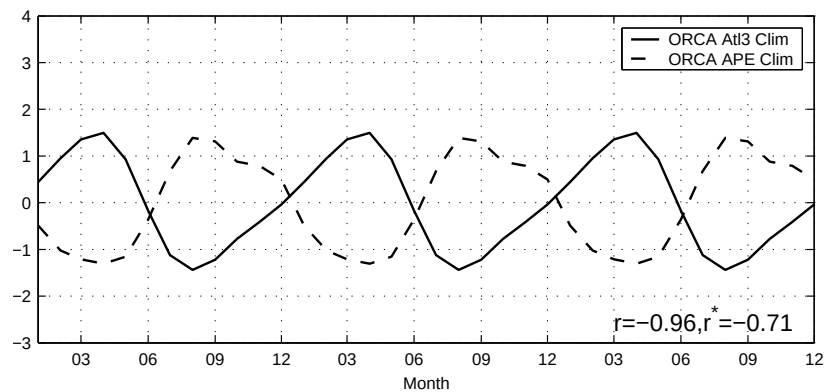


Figure A.10: A comparison between climatological Atl3 SST and equatorial Atlantic (3°S-3°N 60°W-15°E 0-400m) APE. SST and APE values were calculated using ORCA data. The correlation between the Atl3 SST index and equatorial Atlantic APE values (r) is given on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised.

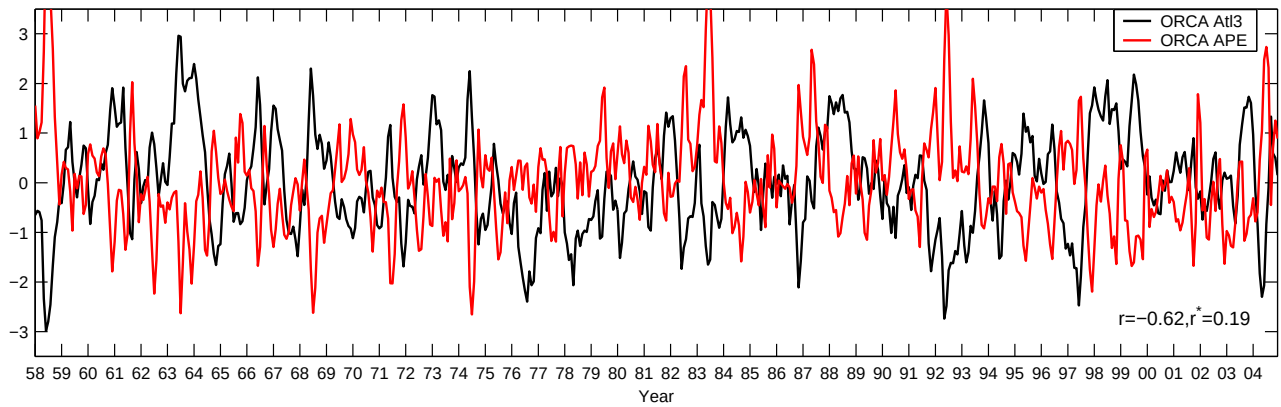


Figure A.11: A comparison between inter-annual Atl3 SST and equatorial Atlantic (3°S - 3°N 60°W - 15°E 0-400m) APE anomalies. SST and APE anomalies were calculated using ORCA data. The correlation between anomalies in each SST index and APE (r) is given on each figure, together with the required correlation value for significance at the 95% level taking into account the effective degrees of freedom (r^*). SST and APE values have been normalised as well as detrended.

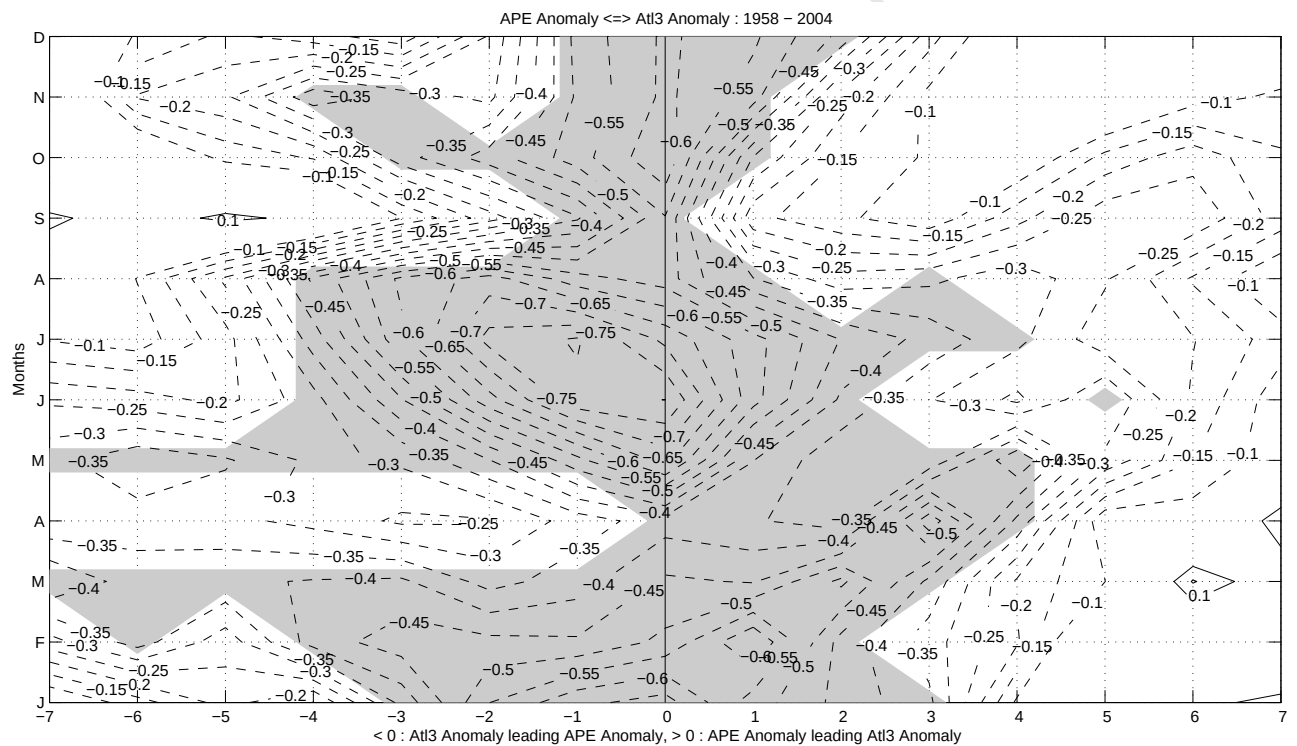


Figure A.12: The seasonal dependence in correlations between Atl3 SST and equatorial Atlantic APE anomalies. SST and APE anomalies were calculated using ORCA data. Positive correlations are contoured in solid lines and negative correlations in dashed lines. Correlations significant above the 95 percentile are shaded in grey.

Appendix B

Surface Layer Heat Budget

The mean temperature, $\bar{T}$, within a surface layer of depth $h(t,x,y)$ is defined as:

$$\begin{aligned}\bar{T} &= \frac{1}{h} \int_{-h}^0 T dz \\ \bar{T}h &= \int_{-h}^0 T dz\end{aligned}\tag{B.1}$$

Taking the time derivative of Equation B.1 gives us:

$$\begin{aligned}\frac{\partial \bar{T}h}{\partial t} &= \frac{\partial \int_{-h}^0 T dz}{\partial t} \\ h \frac{\partial \bar{T}}{\partial t} + \frac{\partial h}{\partial t} \bar{T} &= \int_{-h}^0 \frac{\partial T}{\partial t} dz + \frac{\partial h}{\partial t} T_{z=-h} \\ h \frac{\partial \bar{T}}{\partial t} &= \int_{-h}^0 \frac{\partial T}{\partial t} dz + \frac{\partial h}{\partial t} T_{z=-h} - \frac{\partial h}{\partial t} \bar{T} \\ \frac{\partial \bar{T}}{\partial t} &= \underbrace{\frac{1}{h} \int_{-h}^0 \frac{\partial T}{\partial t} dz}_{RHS1} - \underbrace{\frac{1}{h} \frac{\partial h}{\partial t} (\bar{T} - T_{z=-h})}_{RHS2}\end{aligned}\tag{B.2}$$

$$\frac{\partial \bar{T}}{\partial t} = \underbrace{\frac{1}{h} \int_{-h}^0 \frac{\partial T}{\partial t} dz}_{RHS1} - \underbrace{\frac{1}{h} \frac{\partial h}{\partial t} (\bar{T} - T_{z=-h})}_{RHS2}\tag{B.3}$$

$T_{z=-h}$ is the temperature at the base of the chosen surface layer. As expressed by Equation B.3 changes in the mean temperature of this layer are forced by the temperature evolution equation (Equation B.4) vertically integrated over this surface layer (Equation B.3, RHS1) and the entrainment of water into the surface layer from below (Equation B.3, RHS2). Where the temperature evolution equation (Equation B.4), is given by:

$$\begin{aligned}\frac{\partial T}{\partial t} &= -\vec{u} \cdot \nabla T + \nabla \cdot \left(K_{th} \frac{\partial T}{\partial x} + K_{th} \frac{\partial T}{\partial y} + K_{tv} \frac{\partial T}{\partial z} \right) + Q \\ &= -\vec{u} \cdot \nabla T + K_{th} \nabla^2 T^{horiz} + \frac{\partial}{\partial z} (K_{tv} \frac{\partial T}{\partial z}) + Q\end{aligned}\tag{B.4}$$

K_{th} represents the horizontal temperature diffusion coefficient, K_{tv} the vertical temperature diffusion coefficient and Q any additional source term, which in this case takes the form of subsurface heating due the penetration of the solar radiative flux $Q = \frac{q_s}{\rho_o C_p} \frac{df}{dz}$ (q_s the surface solar radiative flux and $f(z)$ the fraction of solar radiation reaching depth z).

Substituting Equation B.4 into B.3:

$$\begin{aligned} \frac{\partial \bar{T}}{\partial t} &= \frac{1}{h} \int_{-h}^0 \left\{ -\vec{u} \cdot \nabla T + K \nabla_{th}^2 T + \frac{\partial}{\partial z} (K_{tv} \frac{\partial T}{\partial z}) + \frac{q_s}{\rho_o C_p} \frac{df}{dz} \right\} dz - \frac{1}{h} \frac{\partial h}{\partial t} (\bar{T} - T_{z=-h}) \\ &= \underbrace{-\frac{1}{h} \int_{-h}^0 \vec{u} \cdot \nabla T dz}_A + \underbrace{\frac{1}{h} \int_{-h}^0 K_{th} \nabla^2 T dz}_{D_h} + \underbrace{\frac{1}{h} \int_{-h}^0 \frac{\partial}{\partial z} (K_{tv} \frac{\partial T}{\partial z}) dz}_{D_v} + \underbrace{\frac{1}{h} \int_{-h}^0 \frac{q_s}{\rho_o C_p} \frac{df}{dz} dz}_Q - \underbrace{\frac{1}{h} \frac{\partial h}{\partial t} (\bar{T} - T_{z=-h})}_E \end{aligned} \quad (B.5)$$

Term A represents the role of advective processes, and can be split into it's x , y and z components.

$$\underbrace{-\frac{1}{h} \int_{-h}^0 \vec{u} \cdot \nabla T dz}_A = \underbrace{-\frac{1}{h} \int_{-h}^0 u \frac{\partial T}{\partial x} dz}_{A_x} - \underbrace{\frac{1}{h} \int_{-h}^0 v \frac{\partial T}{\partial y} dz}_{A_y} - \underbrace{\frac{1}{h} \int_{-h}^0 w \frac{\partial T}{\partial z} dz}_{A_z} \quad (B.6)$$

Term D_h represents diffusive processes in the horizontal and term D_v the diffusive processes in the vertical. To separate out the surface heat flux forcing according to the surface boundary condition imposed $(K_{tv} \frac{\partial T}{\partial z})_{z=0} = \frac{q_*}{\rho_o C_p}$, D_v may be decomposed further.

$$\begin{aligned} \underbrace{\frac{1}{h} \int_{-h}^0 \frac{\partial}{\partial z} (K_{tv} \frac{\partial T}{\partial z}) dz}_{D_v} &= \frac{1}{h} (K_{tv} \frac{\partial T}{\partial z})_{z=0} - \frac{1}{h} (K_{tv} \frac{\partial T}{\partial z})_{z=-h} \\ &= \underbrace{\frac{q_*}{h \rho_o C_p}}_{RHS1} - \underbrace{\frac{1}{h} (K_{tv} \frac{\partial T}{\partial z})_{z=-h}}_{RHS2} \end{aligned} \quad (B.7)$$

q_* is the non penetrative part of the surface heat flux consisting of the net long-wave radiative flux, the latent heat flux and the sensible heat flux components, ρ_o is the reference density and C_p the specific heat capacity. D_v can then be dissolved. Firstly by combining the surface heat flux forcing term (RHS1) with the forcing term Q which represents heating due to the penetration of the solar radiative flux:

$$\begin{aligned} \underbrace{\frac{1}{h} \int_{-h}^0 \frac{q_s}{\rho_o C_p} \frac{df}{dz} dz}_Q + \frac{q_*}{h \rho_o C_p} &= \frac{q_s f_{z=0}}{h \rho_o C_p} - \frac{q_s f_{z=-h}}{h \rho_o C_p} + \frac{q_*}{h \rho_o C_p} \\ &= \underbrace{\frac{q_* + q_s(1 - f_{z=-h})}{h \rho_o C_p}}_F \end{aligned} \quad (B.8)$$

Secondly by combining the term representing diffusion at the base of the layer (RHS2) with term E which represents the entrainment of water into the surface layer from below. Creating term B an umbrella term for exchanges with underlying water other then vertical advection (*Vialard and Delecluse, 1998a*).

$$\underbrace{-\frac{1}{h}\frac{\partial h}{\partial t}(\bar{T} - T_{-h}) + \frac{1}{h}(K_{tv}\frac{\partial T}{\partial z})_{z=-h}}_E = -\underbrace{\frac{1}{h}\left\{\frac{\partial h}{\partial t}(\bar{T} - T_{z=-h}) - (K_{tv}\frac{\partial T}{\partial z})_{z=-h}\right\}}_B \quad (\text{B.9})$$

The evolution equation may finally be written as:

$$\begin{aligned} \frac{\partial \bar{T}}{\partial t} = & -\underbrace{\frac{1}{h}\int_{-h}^0 u \frac{\partial T}{\partial x} dz}_{A_x} - \underbrace{\frac{1}{h}\int_{-h}^0 v \frac{\partial T}{\partial y} dz}_{A_y} - \underbrace{\frac{1}{h}\int_{-h}^0 w \frac{\partial T}{\partial z} dz}_{A_z} \\ & + \underbrace{\frac{1}{h}\int_{-h}^0 K_{th} \nabla^2 T dz}_{D_h} - \underbrace{\frac{1}{h}\left\{\frac{\partial h}{\partial t}(\bar{T} - T_{z=-h}) - (K_{tv}\frac{\partial T}{\partial z})_{z=-h}\right\}}_B + \underbrace{\frac{q_* + q_s(1 - f_{z=-h})}{h\rho_o C_p}}_F \end{aligned} \quad (\text{B.10})$$

Appendix C

Ekman Transport

As in Gill (1982) (pp 325) a frictional term, $-r\vec{U}_E$, may be included into the horizontal transport equations for an Ekman layer:

$$\frac{\partial U_E}{\partial t} - fV_E = -rU_E + \frac{\tau_x}{\rho} \quad (C.1)$$

$$\frac{\partial V_E}{\partial t} + fU_E = -rV_E + \frac{\tau_y}{\rho} \quad (C.2)$$

Steady state is assumed so terms $\frac{\partial U_E}{\partial t}$ and $\frac{\partial V_E}{\partial t}$ are dropped.

$$U_E = -\frac{rV_E}{f} + \frac{\tau_y}{\rho f} \quad (C.3)$$

$$V_E = \frac{rU_E}{f} - \frac{\tau_x}{\rho f} \quad (C.4)$$

To obtain Equation 5.3, for the zonal Ekman transport, U_E , Equation C.4 is substituted into Equation C.3.

$$\begin{aligned} U_E &= -\frac{r \left(\frac{rU_E}{f} - \frac{\tau_x}{\rho f} \right)}{f} + \frac{\tau_y}{\rho f} \\ U_E &= -\frac{r^2 U_E}{f^2} + \frac{r\tau_x}{\rho f^2} + \frac{\tau_y}{\rho f} \\ U_E + \frac{r^2 U_E}{f^2} &= \frac{r\tau_x}{\rho f^2} + \frac{\tau_y}{\rho f} \\ U_E \left(1 + \frac{r^2}{f^2} \right) &= \frac{r\tau_x}{\rho f^2} + \frac{\tau_y}{\rho f} \\ U_E &= \frac{\frac{r\tau_x}{\rho f^2}}{1 + \frac{r^2}{f^2}} + \frac{\frac{\tau_y}{\rho f}}{1 + \frac{r^2}{f^2}} \\ U_E &= \frac{r\tau_x}{\rho f^2} + \frac{r\tau_x}{\rho r^2} + \frac{\tau_y}{\rho f} + \frac{\tau_y f}{\rho r^2} \\ U_E &= \frac{r\tau_x}{\rho(f^2 + r^2)} + \frac{\tau_y f}{\rho f^2} + \frac{\tau_y f}{\rho r^2} \\ U_E &= \frac{r\tau_x}{\rho(f^2 + r^2)} + \frac{\tau_y f}{\rho(f^2 + r^2)} \end{aligned} \quad (C.5)$$

To obtain Equation 5.4, for the meridional Ekman transport, V_E , Equation C.3 is substituted into Equation C.4.

$$\begin{aligned}
 V_E &= \frac{r \left(-\frac{rV_E}{f} + \frac{\tau_y}{\rho f} \right)}{f} - \frac{\tau_x}{\rho f} \\
 V_E &= -\frac{r^2 V_E}{f^2} + \frac{r\tau_y}{\rho f^2} - \frac{\tau_x}{\rho f} \\
 V_E + \frac{r^2 V_E}{f^2} &= \frac{r\tau_y}{\rho f^2} - \frac{\tau_x}{\rho f} \\
 V_E \left(1 + \frac{r^2}{f^2} \right) &= \frac{r\tau_y}{\rho f^2} - \frac{\tau_x}{\rho f} \\
 V_E &= \frac{\frac{r\tau_y}{\rho f^2}}{1 + \frac{r^2}{f^2}} - \frac{\frac{\tau_x}{\rho f}}{1 + \frac{r^2}{f^2}} \\
 V_E &= \frac{r\tau_y}{\rho f^2} + \frac{r\tau_y}{\rho r^2} - \frac{\tau_x}{\rho f} - \frac{f\tau_x}{\rho r^2} \\
 V_E &= \frac{r\tau_y}{\rho(f^2 + r^2)} - \frac{\tau_x f}{\rho f^2} - \frac{f\tau_x}{\rho r^2} \\
 V_E &= \frac{r\tau_y}{\rho(f^2 + r^2)} - \frac{\tau_x f}{\rho(f^2 + r^2)} \\
 V_E &= \frac{r\tau_y - \tau_x f}{\rho(f^2 + r^2)} \tag{C.6}
 \end{aligned}$$

To obtain Equation 5.2 for the vertical Ekman velocity, W_E , at the base of the Ekman layer, the continuity equation $\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$ is vertically integrated over the depth of the Ekman layer:

$$\begin{aligned}
 \int_{-h}^0 \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) dz &= 0 \\
 \frac{\partial U_E}{\partial x} + \frac{\partial V_E}{\partial y} + W_{z=0} - W_{z=-h} &= 0 \\
 W_{z=-h} &= \frac{\partial U_E}{\partial x} + \frac{\partial V_E}{\partial y} \\
 W_E &= \frac{\partial U_E}{\partial x} + \frac{\partial V_E}{\partial y} \tag{C.7}
 \end{aligned}$$

Where $z=-h$ is the depth of the Ekman layer and the vertical velocity at the surface $W_{z=0}$ is zero under the rigid lid assumption.

Appendix D

KE Evolution Equation

The Kinetic Energy of a volume filled with a hydrostatic fluid is defined as:

$$KE = \iiint \frac{\rho_o}{2} \vec{v}^2 dV \quad (D.1)$$

where ρ_o is the constant reference density, and $\vec{v} = (u, v)$ the horizontal velocity field as vertical velocities do not contribute to KE under the hydrostatic assumption. An evolution equation for KE is derived by taking the time derivative of Equation D.1:

$$\begin{aligned} \frac{dKE}{dt} &= \frac{d(\iiint \frac{\rho_o}{2} \vec{v}^2 dV)}{dt} \\ &= \iiint \frac{\rho_o}{2} \frac{\partial \vec{v}^2}{\partial t} dV + \iiint \frac{\rho_o}{2} \vec{v}^2 \frac{dV}{dt} \end{aligned} \quad (D.2)$$

$\iiint \frac{\rho_o}{2} \vec{v}^2 \frac{dV}{dt} = 0$ assuming the integration volume is fixed in time.

$$\begin{aligned} \frac{dKE}{dt} &= \iiint \frac{\rho_o}{2} \frac{\partial \vec{v}^2}{\partial t} dV \\ &= \iiint \frac{\rho_o}{2} \left(\vec{v} \cdot \frac{\partial \vec{v}}{\partial t} + \frac{\partial \vec{v}}{\partial t} \cdot \vec{v} \right) dV \\ &= \iiint \rho_o \vec{v} \cdot \frac{\partial \vec{v}}{\partial t} dV \end{aligned} \quad (D.3)$$

The temporal evolution of KE is therefore equivalent to the inner product of the horizontal momentum evolution equation and the horizontal velocity vector. The horizontal momentum evolution equation being:

$$\frac{\partial \vec{v}}{\partial t} = - \underbrace{\vec{u} \cdot \nabla \vec{v}}_{A_m} - 2\Omega \times \vec{v} - \underbrace{\frac{1}{\rho_o} \nabla p}_{P_m} + \underbrace{\nabla \cdot \tau}_{D_m} \quad (D.4)$$

where $\vec{u} = (u, v, w)$ represents the full three dimensional velocity field. τ is the viscous stress tensor $\tau = K_{mh} \frac{\partial \vec{v}}{\partial x} + K_{mh} \frac{\partial \vec{v}}{\partial y} + K_{mv} \frac{\partial \vec{v}}{\partial z}$, where K_{mh} is the horizontal viscosity coefficient and K_{mv} the vertical viscosity coefficient. Substituting Equation D.4 into Equation D.3:

$$\begin{aligned}
\frac{dKE}{dt} &= \iiint \rho_o \vec{v} \cdot \left(-\vec{u} \cdot \nabla \vec{v} - 2\Omega \times \vec{v} - \frac{1}{\rho_o} \nabla p + \nabla \cdot \tau \right) dV \\
&= \underbrace{- \iiint \vec{u} \cdot \nabla \left(\frac{\rho_o}{2} \vec{v}^2 \right) dV}_{A_{KE}} - \underbrace{\iiint \vec{v} \cdot \nabla p dV}_{P_{KE}} + \underbrace{\iiint \rho_o \vec{v} \cdot \nabla \cdot \tau dV}_{D_{KE}}
\end{aligned} \tag{D.5}$$

as $\vec{u} \cdot \vec{v} \cdot \nabla \vec{v} = \vec{u} \cdot \frac{1}{2}(\vec{v} \cdot \nabla \vec{v} + \vec{v} \cdot \nabla \vec{v}) = \vec{u} \cdot \nabla \frac{1}{2} \vec{v}^2$ and $\vec{v} \cdot (2\Omega \times \vec{v}) = 0$.

In this form, the temporal evolution of KE within a fluid volume is driven by three terms:

A_{KE} - KE changes due to the horizontal advection of KE.

P_{KE} - KE changes due to horizontal pressure work.

D_{KE} - KE changes due to the combined effects of the dissipation of KE within the domain and diffusion of KE across the surfaces of the volume.

Each term in the simulated horizontal momentum evolution equation (Equation D.4) has been saved as a two day average. Using this output, the KE balance is calculated in the form given by Equation D.5. In Equation D.5, term:

A_{KE} - is obtained based on two day averages of the horizontal velocity field, $\vec{v}$, multiplied by two day averages of term A_m in Equation D.4 (the horizontal momentum advection term).

P_{KE} - is obtained based on two day averages of the horizontal velocity field, $\vec{v}$, multiplied by two day averages of term P_m in Equation D.4 (the horizontal pressure gradient term).

D_{KE} - is obtained based on two day averages of the horizontal velocity field, $\vec{v}$, multiplied by two day averages of term D_m in Equation D.4 (the horizontal momentum dissipation term).

The reference density value used is, $\rho_o = 1025 \text{ kg m}^{-3}$, which is consistent with the reference density specified for the TATL-ROMS simulation. The balance of terms achieved for Equation D.5 is depicted in Figure D.1.

However the KE evolution equation in the form given by Equation D.5 does not resolve the exchange of energy between KE and APE within the chosen domain. To acquire the buoyancy power term ($\Phi_{apk} = \iiint g \tilde{\rho} w dV$), the exchange term with the APE evolution equation, vertical dynamics need to be included. The hydrostatic equation $\frac{\partial p}{\partial z} = -\rho g$ is therefore incorporated into Equation D.5. The hydrostatic equation is multiplied by the vertical velocity component w ($\frac{\partial p}{\partial z} = -\rho g$) and considering that $-w \frac{\partial p}{\partial z} - g \rho w = 0$ these terms are then added to Equation D.5.

$$\begin{aligned}
\frac{dKE}{dt} &= - \iiint \vec{u} \cdot \nabla \left(\frac{\rho_o}{2} \vec{v}^2 \right) dV - \iiint \vec{v} \cdot \nabla p dV - \iiint w \frac{\partial p}{\partial z} dV - \iiint g \rho w dV + \iiint \rho_o \vec{v} \cdot \nabla \cdot \tau dV \\
&= - \iiint \vec{u} \cdot \nabla \left(\frac{\rho_o}{2} \vec{v}^2 \right) dV - \iiint \vec{u} \cdot \nabla p dV - \iiint \rho g w dV + \iiint \rho_o \nabla \cdot (\vec{v} \cdot \tau) dV - \iiint \rho_o \nabla \vec{v} \cdot \tau dV
\end{aligned} \tag{D.6}$$

Density is then decomposed into, $\rho^*(z, t)$, the vertical density profile associated with the hydrostatically balanced density profile of the minimum PE rest state/BPE, and $\tilde{\rho}(\vec{x}, t)$, perturbations from this reference state, $\rho = \rho^*(z, t) + \tilde{\rho}(\vec{x}, t)$. Similarly, the pressure field maybe be deconstructed into $p = p^*(z, t) + \tilde{p}(\vec{x}, t)$,

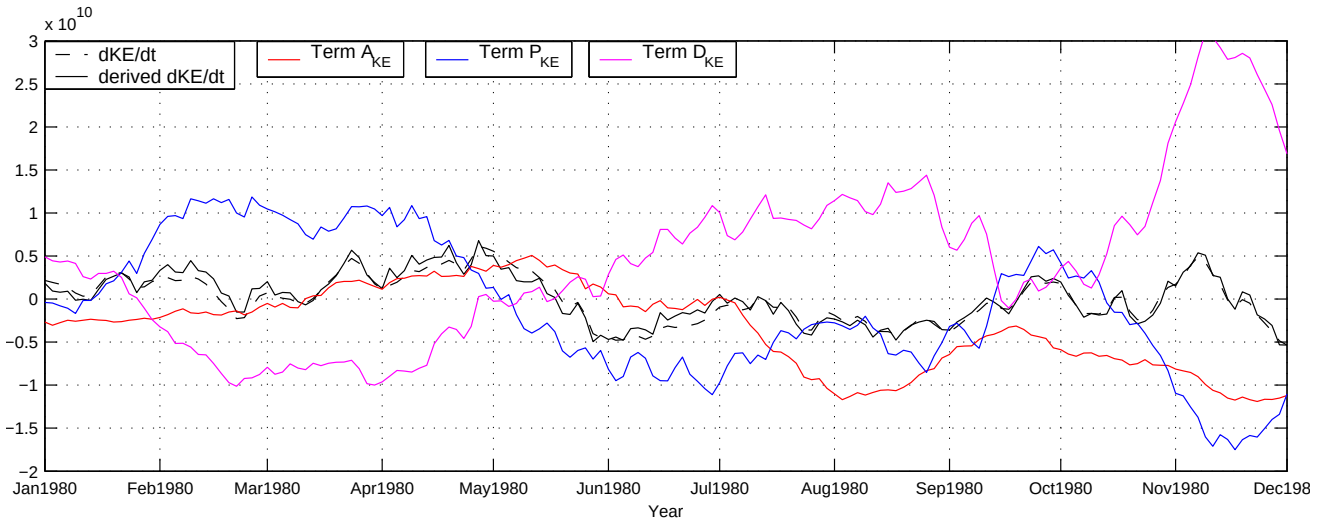


Figure D.1: Balance of terms for the KE evolution equation in the form given by Equation D.5, assessed over the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m). dKE/dt represents the actual rate of change of KE where as *derived* dKE/dt represents the rate of change of the KE estimated by balancing the terms within Equation D.5. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

where $p^*(z, t)$ is the pressure field in which all pressure surfaces are level and in hydrostatic balance with $\rho^*(z, t)$ ($\frac{\partial p^*}{\partial z} = -\rho^*g$). Substituting $\rho = \rho^*(z, t) + \tilde{\rho}(\vec{x}, t)$ and $p = p^*(z, t) + \tilde{p}(\vec{x}, t)$ into $\iiint \rho g w dV$ and $\iiint \vec{u} \cdot \nabla p dV$ respectively, it becomes apparent that these two terms may be rewritten as, $-\iiint \vec{u} \cdot \nabla (p^* + \tilde{p}) dV - \iiint (\rho^* + \tilde{\rho}) g w dV = -\iiint \vec{u} \cdot \nabla \tilde{p} dV - \iiint \tilde{\rho} g w dV$, considering $-\iiint \vec{v} \cdot \nabla p^* dV = 0$.

Invoking the continuity equation $\nabla \cdot \vec{u} = 0$

$$\frac{dKE}{dt} = -\iiint \nabla \cdot \left(\frac{\rho_o}{2} \vec{v}^2 \vec{u} \right) dV - \iiint \nabla \cdot (\tilde{p} \vec{u}) dV - \iiint \tilde{\rho} g w dV + \iiint \rho_o \nabla \cdot (\vec{v} \cdot \tau) dV - \iiint \rho_o \nabla \vec{v} \cdot \tau dV \quad (D.7)$$

Invoking the divergence theorem:

$$\begin{aligned} \frac{dKE}{dt} = & -\oint \left(\frac{\rho_o}{2} \vec{v}^2 \vec{u} \right) \cdot \hat{n} dS - \oint (\tilde{p} \vec{u}) \cdot \hat{n} dS - \iiint \tilde{\rho} g w dV \\ & + \rho_o \oint (\vec{v} \cdot \tau) \cdot \hat{n} dS - \rho_o \iiint \nabla \vec{v} \cdot \tau dV \end{aligned} \quad (D.8)$$

Assuming that the diffusion of KE through the surfaces of the domain is negligible with the exception of the surface $z=0$, $\rho_o \oint (\vec{v} \cdot \tau) \cdot \hat{n} dS = \iint_{z=0} \vec{v} \cdot \tau_{z=0} dS$. A further assumption is that $\tau_{z=0} = (\rho_o K_{mv} \frac{d\vec{v}}{dz})_{z=0} \approx \tau_s$ where τ_s is the surface wind stress. The surface wind stress τ_s can be decomposed into it's x (τ_s^x) and y (τ_s^y) components $\tau_s = \tau_s^x + \tau_s^y$, where $\tau_s^x = \rho_o K_{mv} \frac{du}{dz} |_{z=0}$ and $\tau_s^y = \rho_o K_{mv} \frac{dv}{dz} |_{z=0}$.

This brings us to the final form of the KE evolution equation used here:

$$\begin{aligned}
\frac{dKE}{dt} = & - \underbrace{\oint \left(\frac{\rho_o}{2} \vec{v}^2 \vec{u} \right) \cdot \hat{n} dS}_{\Phi_{ka}} - \underbrace{\oint (\tilde{p} \vec{u}) \cdot \hat{n} dS}_{\Phi_{pw}} - \underbrace{\iiint \tilde{\rho} g w dV}_{\Phi_{apk}} + \underbrace{\iint_{z=0} \vec{v} \cdot \tau_s dS}_{\Phi_{ww}} \\
& - \underbrace{\iiint \rho_o \left(K_{mh} \left(\frac{\partial \vec{v}}{\partial x} \cdot \frac{\partial \vec{v}}{\partial x} \right) + K_{mh} \left(\frac{\partial \vec{v}}{\partial y} \cdot \frac{\partial \vec{v}}{\partial y} \right) + K_{mv} \left(\frac{\partial \vec{v}}{\partial z} \cdot \frac{\partial \vec{v}}{\partial z} \right) \right) dV}_{\Phi_{ss}} \quad (D.9)
\end{aligned}$$

In this form, the temporal evolution of KE within a fluid volume is driven by five terms:

- Φ_{ka} -the advection of KE across the surfaces of the volume.
- Φ_{pw} -pressure work across the surfaces of the volume (Winters et al., 1995).
- Φ_{apk} -the buoyancy power term which is a reversible exchange term with the APE evolution equation (Equation 5.29).
- Φ_{ww} -work done by the wind stress acting on surface currents (Goddard and Philander, 2000).
- Φ_{ss} -work done by shear stresses of horizontal flows within the domain (Winters et al., 1995).

Having obtained an adequate balance between terms for Equation D.5 (Figure D.1), following the method outlined below, these terms are then modified to obtain the KE evolution terms in their final form as given by Equation D.9:

- Φ_{apk} - the buoyancy power term is calculated by based on two day averages of vertical velocity, w , multiplied by perturbation density values, $\tilde{\rho}$, which have been calculated based two day averages of temperature and salinity according to the linear equation of state given by Equation 5.9.
- Φ_{pw} - pressure work across the surfaces of the volume is calculated by adding the buoyancy power term to the P_{KE} term of Equation D.5, as $g\tilde{\rho}w = -w\frac{\partial \tilde{p}}{\partial z}$ and as it has been shown in the derivation of Equation D.9 $-\oint (\tilde{p} \vec{u}) \cdot \hat{n} dS = -\iiint \nabla \cdot (\tilde{p} \vec{u}) dV = -\iiint \vec{u} \cdot \nabla \tilde{p} dV = -\iiint \vec{v} \cdot \nabla \tilde{p} dV - \iiint w \frac{\partial \tilde{p}}{\partial z} dV = -\iiint \vec{v} \cdot \nabla \tilde{p} dV - \iiint w \frac{\partial \tilde{p}}{\partial z} dV$ as $-\iiint \vec{v} \cdot \nabla \tilde{p} dV = 0$.
- Φ_{ka} - the advection of KE across the surfaces of the volume is the equivalent of term A_{KE} in Equation D.5.
- Φ_{ww} - the work done by the wind stress acting on surface currents is calculated based on two day averages of the x (τ^x) and y (τ^y) components of the wind stress used to force the simulation multiplied by two day averages of the x (u) and y (v) velocity components.
- Φ_{ss} - the work done by shear stresses of horizontal flows within the domain is calculated as the difference between term D_{KE} in Equation D.5 and the wind work term, Φ_{ww} .

The balance of terms achieved for the KE evolution equation in its final form (Equation D.9) is depicted in Figure D.2.

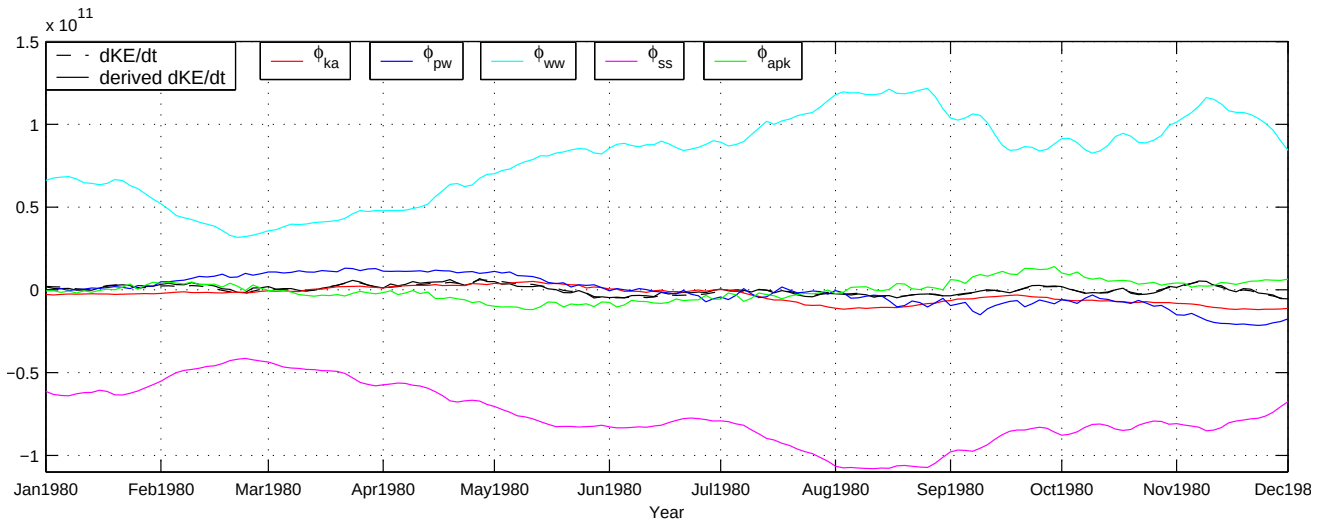


Figure D.2: Balance of terms for the KE evolution equation in the form given by Equation D.9, assessed over the equatorial Atlantic domain (3°S-3°N 60°W-15°E 0-400m). dKE/dt represents the actual rate of change of KE where as *derived* dKE/dt represents the rate of change of the KE estimated by balancing the terms within Equation D.9. A 14 day running mean has been applied to the time series to smooth out the high frequency variability.

Appendix E

List of Acronyms

Listed as they appear in the text.

NRF - National Research Foundation

CNRS - French National Centre for Scientific Research

CHPC - The Centre for High Performance Computing

LPO - Laboratoire de Physique des Oceans

ENSO - El Niño Southern Oscillation

SST - Sea Surface Temperature

WES - Wind-Evaporation-SST

SLP - Sea Level Pressure

COADS - Comprehensive Ocean-Atmosphere Data Set

ROMS - Regional Ocean Modelling System

APE - Available Potential Energy

SSH - Sea Surface Height

ITCZ - Inter Tropical Convergence Zone

STC - SubTropical Cell

MOC - Meridional Over-turning Circulation

SEC - South Equatorial Current

NBC - North Brazil Current

EUC - Equatorial Under Current

NECC - North Equatorial Counter Current

NEC - North Equatorial Current

NEUC - North Equatorial Under Current

SEUC - South Equatorial Under Current
GC - Guinea Current
AC - Angola Current
BC - Brazil Current
NEUC - North Equatorial Under Currents
SEUC - South Equatorial Under Currents
NBUC - North Brazil Under Current
ABF - Angola–Benguela Front
PIRATA - Pilot Research Moored Array in the Tropical Atlantic
ABA - Angola Benguela Area
EOF - Empirical Orthogonal Function
NAO - North Atlantic Oscillation
PE - Potential Energy
SODA - Simple Ocean Data Assimilation
WOA05 - World Ocean Atlas 2005
BPE - Background Potential Energy
ECMWF - European Center for Medium Range Weather Forecast
ROMS-TAtl - Regional Ocean Modelling System Tropical Atlantic Simulation
KPP - K-Profile Planetary
NCEP - National Center of Environmental Prediction
NCAR - National Center for Atmospheric Research
SML - Surface Mixed Layer
SSL - SubSurface Layer
KE - Kinetic Energy