2013

Ubiquitous Zonal Bands in Subtropical Oceans Observed From Space

Christian E. Buckingham
University of Rhode Island, cbuckingham@gso.uri.edu

Follow this and additional works at: https://digitalcommons.uri.edu/oa_diss

Recommended Citation

Buckingham, Christian E., "Ubiquitous Zonal Bands in Subtropical Oceans Observed From Space" (2013).
Open Access Dissertations. Paper 53.
https://digitalcommons.uri.edu/oa_diss/53

This Dissertation is brought to you for free and open access by DigitalCommons@URI. It has been accepted for inclusion in Open Access Dissertations by an authorized administrator of DigitalCommons@URI. For more information, please contact digitalcommons-group@uri.edu.

UBIQUITOUS ZONAL BANDS IN SUBTROPICAL OCEANS OBSERVED FROM SPACE

BY

CHRISTIAN E. BUCKINGHAM

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY
IN
OCEANOGRAPHY

UNIVERSITY OF RHODE ISLAND

2013

DOCTOR OF PHILOSOPHY DISSERTATION
OF
CHRISTIAN E. BUCKINGHAM

APPROVED:

Dissertation Committee:

Major Professor Peter C. Cornillon

H. Thomas Rossby

Peter F. Swaszek

Nasser H. Zawia

DEAN OF THE GRADUATE SCHOOL

UNIVERSITY OF RHODE ISLAND

2013

ABSTRACT

Geostrophic turbulence predicts zonal jets in our oceans and atmospheres. Under this theory, energy cascades from small to large scales, while enstrophy—i.e., squared vorticity—cascades to increasingly smaller scales. The combined effects of the latitudinal variation in the Coriolis parameter and the inverse (forward) and cascade of energy (enstrophy) are expected to result in zonal jets [Rhines, 1975]. Other frameworks for the existence of zonal jet-like structures in the ocean include secondary instability theory [Pedlosky, 1975; Berloff *et al.*, 2009, 2011] and instabilities and/or β -plumes radiating from an eastern boundary current [Hristova *et al.*, 2008; Wang *et al.*, 2012; Centurioni *et al.*, 2008; Melnichenko *et al.*, 2010; Afanasyev *et al.*, 2012; Di Lorenzo *et al.*, 2012].

Recent studies suggest such features exist in the ocean [Maximenko *et al.*, 2005, 2008; Huang *et al.*, 2007; Scott *et al.*, 2008; Centurioni *et al.*, 2008; Ivanov *et al.*, 2009; van Sebille *et al.*, 2011; Cravatte *et al.*, 2012]. These studies make use of various combinations of altimeter measurements of sea surface height (SSH), wind reanalysis, upper ocean profiles and drift trajectories from floats to estimate ocean circulation patterns. These data are temporally averaged and generally spatially filtered to reveal jet-like structures in the worlds oceans. Given an uncertainty regarding the physics of the jet-like features, they were referred to as *striations* [Maximenko *et al.*, 2008]. Striations are zonally-elongated mesoscale (100–400 km) features observed in time-averaged zonal geostrophic velocity, $\bar{u}$. Characterized by zonal scales > 1000 km, meridional scales ~ 100 km and speeds $O(1 \text{ cm s}^{-1})$, they alternate in direction in meridional cross-sections of $\bar{u}$ and are nominally separated by 200 km. Zonal patterns with similar appearance have also been observed in front probability derived from microwave sea surface temperature (SST) [Buckingham and Cornillon, 2010] and measurements of surface winds [Maximenko *et al.*, 2010; Divakaran and Brassington, 2011].

In this study, we investigate the existence of stationary jet-like structures in subtropical oceans observed from space. Motivated by a kinematic explanation for the structures [Schlax and

Chelton, 2008], we explore the relationship between (1) mesoscale eddies and striations in SSH and (2) mesoscale eddies and zonal bands in microwave SST. Some of the more salient results of our study of striations include (i) propagating eddies explain a large fraction of the variance in striations, (ii) eddies having a broad range of amplitudes and scales are most correlated with and contribute most to the observed striations and (iii) the standard deviation of $\overline{u}$ does not decay inversely with averaging period. In our study of zonal bands, we find that large equatorward gradients explain a considerable fraction of the variance in mean equatorward SST gradient. High gradient events (HGEs) propagate westward with speed comparable to mesoscale eddies and appear to arise due to the combined effects of (i) contrasting temperatures of neighboring eddies and (ii) advection of surface waters by eddies embedded in a background temperature gradient. Results of both studies indicate mesoscale eddies are intimately related to the existence of multiple zonal jet-like structures, as postulated by *Schlax and Chelton* [2008]. Furthermore, the persistence of patterns with averaging period suggest the existence of a secondary, latent signal and/or preferred eddy paths.

ACKNOWLEDGEMENTS

I would like to thank my major professor, Dr. Peter Cornillon, for helping me to develop this topic into a fully funded research project; I am deeply grateful for his patience and wisdom. I would be remiss if I did not acknowledge the efforts of Kelsey Obenour, whose summer research project led to my dissertation, and Dr. Dudley Chelton and Michael Schlax, whose published research and several chance interactions provided a solid framework for my own research. I am grateful to the professors and scientists at the University of Rhode Island (URI) for giving me the chance to blend two traditionally unrelated areas of research – electrical engineering and oceanography. I am particularly grateful to Dr. Tetsu Hara for teaching me much of what I know about fluid dynamics, Dr. Randy Watts for encouragement over the years and committee members, Drs. Tom Rossby, Peter Swaszek, Lew Rothstein and Gavino Puggioni, for helpful suggestions throughout this study. Dr. Fabian Schlosser also provided numerous helpful suggestions. I would also like to acknowledge individuals within the STF Group at the Johns Hopkins University / Applied Physics Laboratory, who instilled in me foundational principles prior to my coming to URI and that have served me well. Finally, this research would not have been possible without the support of friends and family, including a group of surfing friends living in Narragansett, Rhode Island, the Suzuki family, Penelope Howe, Lorne Belden, and my parents, Richard and Barbara Buckingham.

This research was funded by the Earth and Space Science Fellowship from the National Aeronautics and Space Administration (NASA).

PREFACE

This dissertation is written in the manuscript format. Chapter 1 was published in the *Journal of Geophysical Research – Oceans* in 2013. Chapter 2 will be submitted to the *Journal of Geophysical Research – Oceans* later this year.

TABLE OF CONTENTS

Abstract	ii
Acknowledgements	iv
Preface	v
Table of Contents	vi
List of Tables	viii
List of Figures	ix
 MANUSCRIPT 1 The contribution of eddies to striations in absolute dynamic	
topography	1
1.1 Abstract	2
1.2 Introduction	2
1.2.1 General interpretation of striations	4
1.2.2 Focus of the study	5
1.3 Methods	6
1.3.1 Processing steps	8
1.3.2 Characterization and comparison of $\overline{u}_{total}$, $\overline{u}_{eddy}$ and $\overline{u}_{diff}$	13
1.4 Results	14
1.4.1 Zonal and meridional spectra of $\overline{u}$	14
1.4.2 Comparison of $\overline{u}_{total}$, $\overline{u}_{eddy}$ and $\overline{u}_{diff}$	15
1.4.3 Decay of $\sigma_{\overline{u}}$ with averaging period	15
1.4.4 Class of eddies most responsible for striations	16
1.5 Conclusions and discussion	18
1.5.1 Limitations	18
1.5.2 The contribution of eddies to striations in SST	22
References	24

MANUSCRIPT 2 Zonal bands in microwave sea surface temperature	39
2.1 Abstract	40
2.2 Introduction	40
2.2.1 Background	41
2.2.2 A brief overview of microwave SST	42
2.2.3 Focus of the study	43
2.3 Methods	43
2.3.1 Dataset descriptions	44
2.3.2 Description of mean equatorward SST gradient	45
2.3.3 Description of SST front probability	46
2.4 Results	47
2.4.1 Demonstration of bands in microwave SST	48
2.4.2 Distributions of equatorward gradients in the subtropics	48
2.4.3 Characteristics of large SST gradients	49
2.5 Persistence with averaging period	53
2.6 Discussion	54
References	58
 APPENDIX A Statistics of neighboring mesoscale eddies obtained from a database of tracked eddies	 76
 APPENDIX B SST, SSH and Argo data averaged within a coordinate sys- tem centered on negative HGEs	 79
Bibliography	82

LIST OF TABLES

1.1	Variances and covariances of $\overline{u}_{total}$, $\overline{u}_{eddy}$ and $\overline{u}_{diff}$. Units are $\text{cm}^2 \text{s}^{-2}$	38
1.2	Fractional variances and correlation coefficients between $\overline{u}_{total}$, $\overline{u}_{eddy}$ and $\overline{u}_{diff}$. Results of applying the contour-identification/removal algorithm to SLA (refer- enced to an Oct. 1992–Dec. 2008 mean) are highlighted by asterisks.	38
1.3	Parameters α and β obtained from least-squares fits of $\sigma_{\overline{u}_{total}}$ and $\sigma_{\overline{u}_{eddy}}$ during four time periods. The fit has the form $y = \alpha T^\beta$, where T is the averaging period in weeks. Units of α are $\text{cm s}^{-1} \text{wks}^{-\beta}$	38
2.1	Definitions of subtropical regions examined in this study.	75
2.2	The magnitude and position of extrema in average SST anomaly. Units of mag- nitude are $^\circ\text{C}$. Units of position are km and are given in terms of distances from the center of the front (positive, eastward and equatorward). Here, in contrast with plots in Figure 2.9, we define the center of the front as the extremum in average equatorward SST gradient.	75
2.3	As in Table 2.2 but for average SSH anomaly. Units of magnitude are cm. . . .	75
2.4	Fraction of variance in mean equatorward SST gradient due to HGEs. Averaging periods are one, four and nine years starting in June of the displayed year. In the case of multi-year averages, averaging periods commence in June of the first year and end in June of the second year. See text for a definition of HGEs. . . .	75

LIST OF FIGURES

- 1.1 Global maps of time-averaged zonal geostrophic velocity, $\overline{u}$, estimated from (a) observed and (b) simulated SSH, 1993-1996. Observed SSH consist of AVISO/CLS ADT and were spatially high pass-filtered prior to estimating zonal geostrophic velocity. Simulated SSH were obtained by superimposing Gaussian eddies with amplitude, scale, position and lifetime identically specified by the eddy database [CCS2011] onto an otherwise flat ocean. Black boxes denote the region examined in this study (20-50°S, 200-280°E). 30
- 1.2 (a) A westward-propagating, anticyclonic eddy in the Southern Hemisphere has zonal and meridional velocities u and v along the perimeter of the eddy as shown. (b) As velocity fields are averaged in time, $\overline{u}$ has small but finite amplitude while $\overline{v}$ approaches zero. The effect of the eddy is to create striations in $\overline{u}$. Illustration after *Scott et al.* [2008]. 31
- 1.3 Effective magnitude response of the filtering process used in this study. The filter is nearly isotropic in latitude θ and longitude ϕ in units of deg. Wavelengths in units of km assume 1 deg = 111.2 km and, thus, correspond to meridional distances. The shaded box highlights wavelengths of 300-500 km. 31
- 1.4 Illustration of (a) the contour-identification and (b) eddy removal algorithm applied to filtered ADT in the South Pacific, valid January 6, 1993. Black dots denote eddy centers obtained from the eddy database. 32
- 1.5 Histogram of eddy perimeter locations determined by the contour-identification algorithm. Eddy locations are shown as scaled distances from eddy centroids and distances are normalized by the horizontal eddy scale L as defined in the eddy database. 33

1.6	(a) Zonal and (b) meridional spectra of $\bar{u}$ in the South Pacific as a function of averaging period. Solid horizontal lines mark averaging periods of 4 years while vertical dashed lines highlight zonal and meridional wavelengths of 1000 and 400 km, respectively. Spectra were created using all available data between January 1, 1993 and December 31, 2008. Individual 4-yr averaging periods show similar structure. Note that spectral contours are illustrated on a logarithmic scale and units are $\text{cm}^2 \text{s}^{-2} \text{km}$	33
1.7	Comparison of (a-d) $\bar{u}_{total}$, (e-h) $\bar{u}_{eddy}$ and (i-l) $\bar{u}_{diff}$ estimated from ADT during four 4-yr averaging periods. Contour lines are shown every 2 cm s^{-1} and contours in difference plots (i-l) correspond to those of $\bar{u}_{total}$	34
1.8	(a) Standard deviation of $\bar{u}_{total}$ and $\bar{u}_{eddy}$ as functions of averaging period, beginning January 1993, increasing in 4-wk intervals and ending December 1996. Solid line: least-squares fit, $y = \alpha T^\beta$ with $\alpha = 9.301$ and $\beta = -0.437$. Dashed line: $y = \alpha T^\beta$ with $\alpha = 9.301$ and $\beta = -1$ as proposed by a model of random-positioned eddies. (b) Same as for (a) but for all time periods and only showing least-squares fits and proposed falloffs. Regression coefficients are listed in Table 2.3.	35
1.9	Fraction of total variance contained in $\bar{u}_{diff}$ (blue, top) and correlation coefficients between $\bar{u}_{total}$ and $\bar{u}_{eddy}$ (red, bottom) as a function of eddy (a) amplitude and (b) scale. Thin lines illustrate values for each averaging period and thick, dashed lines illustrate the means. Bin widths are 1 cm and 10 km for eddy amplitude and scale, respectively.	35
1.10	Fraction of total variance contained in $\bar{u}_{diff}$ (blue, top) and correlation coefficients between $\bar{u}_{total}$ and $\bar{u}_{eddy}$ (red, bottom) as a function of revised eddy amplitude.	36
1.11	Histogram of eddies in the South Pacific (1992-2008) as a function of revised amplitude and scale. The black box highlights the intersection of amplitudes and scales most correlated with observed striations.	36

1.12	Time-averaged SST (2005-2008) with contours of time-averaged ADT (2005-2008) overlaid. Contours are given every 0.5 cm between -1 and 1 cm and both SST and ADT fields have been spatially high pass-filtered in a manner following <i>Maximenko et al.</i> [2008]. Bathymetric features shallower than 2500 m are highlighted in black.	37
2.1	Zonal geostrophic velocity estimated from MDOT (1992–2002; color) with locations of normalized SST front probability overlaid (2002–2011; black). Normalization was performed by subtracting from each pixel the local mean and dividing by the standard deviation within this same window. We used a $4.1^\circ \times 4.1^\circ$ window centered on each pixel and a threshold of 1.2 when displaying normalized front probability.	63
2.2	Effective magnitude-squared response of the high pass-filtering methods used in this study. The filter is nearly isotropic in latitude and longitude in units of deg. Wavelengths in units of km assume $1 \text{ deg} = 111.2 \text{ km}$ and, thus, correspond to meridional distances. The shaded region highlights wavelengths of 300–500 km. For convenience, this filter response is compared with that used in an earlier study [<i>Buckingham and Cornillon</i> , 2013].	63
2.3	(a) Mean equatorward SST gradient, (b) SST front probability and (c) (locally) normalized SST front probability computed from microwave SST (June 2002–May 2011). Data in (a) were spatially high pass-filtered prior to estimating meridional gradients; data in (b) were obtained by applying the SIED to swath SST and computing SST front probability as the number of front detections divided by the number of clear pixels; normalization in (c) was performed as described in Figure 2.1.	64
2.4	Histogram of mean equatorward SST gradient (June 2002–May 2011) at locations where SST front probability exceeds 0.12 (cf. Figure 2.3b). The bulk of the histogram is found right of zero, as demonstrated by the skewness (<i>skew</i>) of the distribution.	65

2.5	(a) Probability of equatorward SST gradient as a function of time-of-year in the subtropical South Pacific (2002–2011); (b) probability of equatorward SST gradient for each subtropical basin (Table 2.1) with the probability of large positive and negative gradients in the South Pacific highlighted; (c) probability of equatorward gradient falling below $-0.0048^{\circ}\text{C km}^{-1}$ (α_1) and exceeding $+0.0048^{\circ}\text{C km}^{-1}$ (α_2). Equatorward gradients were computed from spatially high-filtered SST and do not contain large-scale background gradients.	66
2.6	Spatially-filtered SST (color; seven-day mean centered on June 14, 2006) with spatially-filtered ADT (contours; valid June 14, 2006) in the South Pacific. Positive HGEs are highlighted in black and the centroids of identified frontal features are shown in yellow. Contours range between ± 4 cm.	67
2.7	High gradient events (HGEs) in the South Pacific, 2002–2011: (a) positions of positive (red) and negative (blue) HGEs; (b) signed histogram of HGEs (color) with mean equatorward SST gradient (2002–2011) at $0.0002^{\circ}\text{C km}^{-1}$ overlaid (contours); signed histogram computed as the number of positive HGEs minus number of negative HGEs found within $0.5^{\circ}\times 0.5^{\circ}$ bins; (c) time-longitude diagram of HGEs at 35°S ; solid black line corresponds to an object with constant westward translation speed of 1.68 km d^{-1} and lifetime of 2 years; gray lines are positions of anticyclonic eddies having lifetimes ≥ 16 wks and obtained from the <i>Chelton et al.</i> [2011a] eddy database.	68
2.8	Westward propagation speeds of HGEs and mesoscale eddies in the subtropics. HGE speeds were estimated from the Radon transform of time-longitude histograms of HGEs. Speeds in the North Atlantic are not shown. Eddy speeds correspond to the average speeds of mesoscale eddies observed and tracked by <i>Chelton et al.</i> [2011a] and having lifetimes ≥ 16 wks.	69

2.9	Mean northward SST gradient, SST anomaly and SSH anomaly obtained when averaging within a coordinate system centered on positive HGEs (June 2002–May 2011). Subtropical basins include (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Oceans (Table 2.1). Black dots denote extrema and solid vertical lines highlight average x -positions of the eddy-like pair. Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.	70
2.10	As in Figure 2.10 but obtained when averaging along-track altimeter data (SLA) in the South Pacific (June 2002–May 2011). Altimeter measurements are those from the Jason-1 spacecraft and were not spatially filtered prior to averaging. This figure should be compared with Figure 2.9b, third column.	71
2.11	Mean temperature anomaly from Argo profiles (June 2002–May 2011) at depths of 20, 200 and 1000 m when averaged within a coordinate system centered on positive HGEs. To the left, we illustrate positions of profiles at 1000-m depth. Basins include the (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Ocean (Table 2.1). Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.	72
2.12	Persistence of (a) mean equatorward SST gradient (June 2002–May 2011) and (b) mean zonal geostrophic velocity, $\overline{u}$, estimated from SSH (January 1993–December 2011). Persistence is estimated from a regression of standard deviation over averaging periods $T = 1/13, 1, 2, 3, 4$ years. See text for a definition of persistence. Bathymetric features shallower than 2500 m are highlighted in black.	73

2.13	SST gradients enhanced and reduced by mesoscale eddies in the Southern Hemisphere. We ignore spatial offsets of SST relative to SSH, assume a background Pole-to-Equator temperature gradient and assume the front is dominated by the northward gradient of temperature anomalies themselves. (a) Geostrophic currents associated with positive HGEs: an anticyclonic (a cyclonic) anomaly located North (South) of the front will advect warmer (cooler) waters toward the front-axis. The net effect of advection is to enhance the existing SST gradient. (b) Geostrophic currents associated with negative HGEs: a cyclonic (an anticyclonic) anomaly located North (South) of the front will advect warmer (cooler) water toward the front axis. The net effect of advection is to reduce the existing SST gradient.	74
A.1	Histograms of the positions of neighboring eddies in the South Pacific when examining (a) cyclones closest to anticyclones and (b) anticyclones closest to anticyclones. Distances are defined as distances from eddy centers.	78
A.2	As in Figure A.1 but for (a) cyclones closest to cyclones and (b) anticyclones closest to cyclones.	78
B.1	Mean northward SST gradient, SST anomaly and SSH anomaly obtained when averaging within a coordinate system centered on negative HGEs (June 2002–May 2011). Subtropical basins include (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Oceans (Table 2.1). Black dots denote extrema and solid vertical lines highlight average x -positions of the eddy-like pair. Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.	80

B.2	Mean temperature anomaly from Argo profiles (June 2002–May 2011) at depths of 20, 200 and 1000 m when averaged within a coordinate system centered on negative HGEs. To the left, we illustrate positions of profiles at 1000-m depth. Basins include the (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Ocean (Table 2.1). Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.	81
-----	--	----

MANUSCRIPT 1

The contribution of eddies to striations in absolute dynamic topography

by

Christian E. Buckingham¹ ; Peter C. Cornillon²

is published in *Journal of Geophysical Research – Oceans*, Volume 114,

doi:10.1029/2012JC008231, 2013

¹Ph.D. Candidate, Graduate School of Oceanography, University of Rhode Island, South Ferry Road, Narragansett, RI 02882. E-mail: cbuckingham@gso.uri.edu

²Professor, Graduate School of Oceanography, University of Rhode Island, South Ferry Road, Narragansett, RI 02882.

1.1 Abstract

Distinct four-year averages of absolute dynamic topography reveal striations in all ocean basins during 1993-2008. Striations are alternating mesoscale jet-like structures observed in time-averaged zonal geostrophic velocity, $\bar{u}$. They are characterized by speeds $O(1 \text{ cm s}^{-1})$ and are nominally separated by 200 km in the meridional direction. Similar patterns have been observed in sea level anomaly, mean dynamic topography and Argo float measurements.

Use of a tracked-eddy database in concert with a contour-identification and eddy removal algorithm demonstrates that eddies are a dominant source of striations in $\bar{u}$ in the South Pacific (20–50°S, 200–280°E). Eddies with lifetimes ≥ 4 weeks account for 46-57% of the variance in $\bar{u}$ and correlation coefficients between total and eddy-only $\bar{u}$ are 0.90-0.93. Attention is given to the ability of the algorithm to correctly identify eddies and suggests that a more appropriate bound on the variance due to eddies is $\sim 30\text{--}70\%$. This permits the existence of latent zonal jets and/or β -plumes. Additional findings of the study include (1) a large number of eddies having a broad range of amplitudes and scales contribute most to the eddy-induced patterns and (2) the standard deviation of $\bar{u}$ does not decay inversely with averaging period as proposed by a model of random eddies.

1.2 Introduction

Geostrophic turbulence predicts zonal jets in the ocean and atmosphere. Under this theory, energy cascades from small to large scales while enstrophy—i.e., squared vorticity—cascades from large to small scales. The combined effects of the latitudinal variation in the Coriolis parameter and the inverse and forward cascades of energy and enstrophy, respectively, are thought to result in multiple zonal jets (MZJs) [Rhines, 1975]. Another theoretical framework for the existence of zonal jets in the ocean is secondary instability theory [Pedlosky, 1975; Berloff *et al.*, 2009].

Until recently, MZJs had not been observed in the global ocean. While numerical models consistently produced patterns supporting the existence of MZJs on the Earth (e.g., Williams [1978]; Maltrud and Vallis [1991]; Panetta [1993]; Treguier and Panetta [1994]; Cho and Polvani

[1996]; *Nakano and Hasumi* [2005]), observational support for the existence of zonal jets in the ocean have typically been confined to regional studies such as those in equatorial waters [*Firing*, 1987; *Gouriou et al.*, 2006], the Brazil Basin [*Hogg and Owens*, 1999] and the Southern Ocean [*Nowlin and Clifford*, 1982; *Orsi et al.*, 1995; *Sokolov and Rintoul*, 2007]. However, it was not until 2005, and again in 2008, that patterns resembling MZJs were observed on a global scale [*Maximenko et al.*, 2005, 2008]. In their 2005 study, Maximenko and coauthors created a time-average of sea level anomaly (SLA) and illustrated the existence of zonally-coherent, mesoscale alternating currents in all of the world's oceans. Maximenko and coauthors further demonstrated the existence of stationary jet-like structures in mean dynamic topography (MDT) in their 2008 study. The jet-like structures were referred to as “striations,” in light of uncertainty surrounding the physics of these phenomena. An example of striations is shown in Figure 2.2a.

Subsequent studies have examined striations from various points-of-view. Notable ones include those by *Centurioni et al.* [2008] and *Ivanov et al.* [2009, 2010] in the California Current System (CCS) and *van Sebille et al.* [2011] in the North Atlantic. The latter describe striations using a three-dimensional reconstruction of the density field from Argo profiling floats, while the former use altimeter-based measurements to examine striations. In addition, a recent study by *Cravatte et al.* [2012] documents a series of zonal jets at *intermediate* depth near the equator by averaging drift velocities of Argo floats at parking depths of 1000 m and 1500 m. All studies suggest the existence of zonally-coherent structures in time-averaged zonal geostrophic velocity. Most studies document quasi-stationary striations, although there is some evidence for a slight meridional migration of patterns [*Ivanov et al.*, 2009]. In addition, both *Huang et al.* [2007] and *Scott et al.* [2008] investigate measures of anisotropy for the upper ocean and find that the geostrophic velocity field, when averaged in time, has a greater degree of zonal variance than meridional variance. It is also noteworthy that patterns similar to striations have been observed in front probability estimated from microwave sea surface temperature (SST) [*Buckingham and Cornillon*, 2010], scatterometer estimates of surface winds [*Maximenko et al.*, 2010; *Divakaran*, 2011] and reanalysis of the southeast Indian Ocean [*Divakaran and Brassington*, 2011].

1.2.1 General interpretation of striations

Striations have generally been interpreted as jets [Maximenko *et al.*, 2005; Ivanov *et al.*, 2009; van Seville *et al.*, 2011; Kamenkovich *et al.*, 2009; Berloff *et al.*, 2009; Cravatte *et al.*, 2012]. Such jets have been described as *quasi-zonal* in reference to their deviation from strictly zonal flow [Ivanov *et al.*, 2009; van Seville *et al.*, 2011; Wang *et al.*, 2012] and *latent* in reference to their small amplitude relative to the background eddy field [Berloff *et al.*, 2011]. Hristova *et al.* [2008] suggest that radiating instabilities of an eastern boundary current could act as a source of zonal jets for the interior ocean. This idea was extended by Wang *et al.* [2012] using a simple, nonlinear barotropic quasi-geostrophic model. The authors find that nonlinear interactions are essential in the maintenance of jet-like features emanating from the eastern boundary. A related interpretation is that striations are the result of β -plumes radiating from the eastern boundary [Centurioni *et al.*, 2008; Melnichenko *et al.*, 2010; Afanasyev *et al.*, 2012; Di Lorenzo *et al.*, 2012]. The concept is that processes on the eastern side of the basin radiate Rossby waves and generate coherent vortices through baroclinic instability. The anomalies propagate westward under the influence of β and create striations. In this model, striations are the manifestation of both eddies and jet-like structures emanating from the eastern boundary.

While it is tempting to interpret striations as jets, Schlax and Chelton [2008] caution against this. They argue that westward-propagating anomalies with completely random positions create patterns similar to those observed. Developing a model of westward-propagating Gaussian eddies with amplitudes, scales, lifetimes and propagation speeds characteristic of observed eddies, Schlax and Chelton [2008] demonstrate that random eddies can create striations in time-averaged sea surface height (SSH) and the associated time-averaged zonal geostrophic velocity field, $\bar{u}$. In their model, eddies of large amplitude and scale contribute most to the observed striations and the standard deviation of $\bar{u}$ decays inversely with averaging period. To help illustrate this mechanism, we display in Figure 2.3 how an anticyclonic eddy can create striations in $\bar{u}$ (see also, Fig. 10 of Scott *et al.* [2008]). A similar argument can be made for cyclonic eddies. More generally, a field of mesoscale eddies with propagation direction deviant from westward might produce quasi-zonal patterns.

Figure 2.2b illustrates $\overline{u}$ estimated from four years of simulated SSH due to eddies. In this example, the eddies are not random as in the simulations of *Schlax and Chelton* [2008] but instead have position, amplitude, scale and lifetime specified by a tracked-eddy database [*Chelton et al.*, 2011a] (hereafter, CSS2011). One notes the existence of striations in eddy-generated $\overline{u}$ that are similar in appearance to observed striations (Figure 2.2a).

Such arguments leave inconclusive the presence of zonal jets in the ocean. Given the importance of MZJs to an understanding of the general circulation of the ocean and potential support for geostrophic turbulence theory [*Rhines*, 1975; *Baldwin et al.*, 2007] and/or jet formation by secondary instability mechanisms [*Pedlosky*, 1975; *Berloff et al.*, 2009], we seek to quantify what portion of the striation signal is due to propagating eddies [*Schlax and Chelton*, 2008] and what portion remains unexplained.

1.2.2 Focus of the study

The focus of the present study is on striations in the Subtropical South Pacific Ocean (20-50°S, 200-280°E; black box in Figure 2.2). The region is sufficiently distant from equatorial and coastal regions and does not enclose strong current systems such as the Gulf Stream, Kuroshio and Antarctic Circumpolar Current where a host of dynamical processes complicate interpretation of the data. For this reason, excluding differences in bathymetry, processes governing striations in the South Pacific should be representative of those in most mid-latitude, open-ocean basins.

In the present study, we examine striations in time-averaged absolute dynamic topography (ADT). The motivation for use of ADT in the examination of striations is that it provides a dynamical quantity from which to appropriately isolate and identify eddies. In addition and in contrast with SLA measurements, ADT contains stationary signals which may include zonal jets. Lastly, we mention that, to the extent that mesoscale eddies explain a portion of the striation signal in this dataset, these results may shed light on the zonal patterns in other datasets. For example, at the end of this study we suggest eddies may be responsible for some portion of the banded structure observed in SST.

This study is divided into three parts. In Section 2, we describe methods employed in the study, including isolating eddy signatures and quantifying that portion of the striation signal attributed to eddies. Section 3 summarizes the results and is followed by a discussion in Section 4.

1.3 Methods

Following *Roemmich and Gilson* [2001], we use the term eddy to refer to propagating anomalies seen in the altimeter record (and described by *Chelton and Schlax* [1996], *Chelton et al.* [2007] and CCS2011). That is, we do not distinguish between linear planetary waves (i.e., Rossby waves) and nonlinear coherent vortices in the surface ocean. We recognize that this distinction is important for understanding the dynamics associated with striations—are eddies embedded in MZJs and therefore derive energy from or feedback energy to the mean flow [*Kamenkovich et al.*, 2009]?—but argue that it is not essential for kinematically distinguishing between jets and coherent, propagating anomalies.

The data used in this study are altimeter-derived measurements of SSH and a database of tracked eddies. SSH data consist of AVISO/CLS delayed-time, merged ADT, interpolated by SSALTO/DUACS onto a quarter-degree grid [*Ducet et al.*, 2000; *Le Traon et al.*, 2003], separated in duration by 7 days and referred to as the reference series. (Dataset acronyms are expanded in the Acknowledgments section.) These data span the time period October 1992 to July 2011, although analysis in this study is limited to data between October 1992 and December 2008. The database of tracked eddies, which is described by CCS2011 and generated by D. Chelton and M. Schlax, is available for the aforementioned time period. The eddy dataset used in this study consists of locations, amplitudes, scales, azimuthal speeds and lifetimes of tracked eddies having lifetimes ≥ 4 weeks. One will note that the analysis presented in CCS2011 pertains to eddies with lifetimes ≥ 16 weeks. In spite of this discrepancy, we refer to CCS2011 when speaking of the eddy dataset.

The SSH product known as the “reference” series is used instead of the “updated” series—the reference series consists of data from two altimeters at any given time while the updated

series consists of data from up to four altimeters—because the eddy database of CCS2011 was developed from the reference series and we wish for a close correspondence between ADT fields and the eddy database.

The AVISO/CLS ADT data described above are the sum of altimeter-only measurements in the form of SLA and a 1992-2008 estimate of MDT [Rio *et al.*, 2011]. The MDT field, referred to as combined mean dynamic topography (CMDT), is estimated using several datasets, including velocity estimates from drogued drifting buoys, hydrological profiles from conductivity-temperature-depth casts and Argo profiling floats and wind stress reanalysis to remove Ekman effects. The MDT field also makes use of a geoid model (EIGEN-GRGS.RL02) and an altimeter-derived mean sea surface product (CLS01).

Comment on spatial resolution of ADT

CCS2011 (Section A.3) provide a thorough summary of the resolution characteristics of SLA based on meridional and zonal spectra. Since snapshots of ADT are heavily determined by SLA, conclusions reached by CCS2011 apply to our dataset, as well. We summarize these below.

CCS2011 find that (1) filtering characteristics of the optimal interpolation scheme appear to be approximately isotropic when expressed in degrees of latitude and longitude, (2) filter attenuation can be defined as monotonically increasing with decreasing wavelength (increasing wavenumber) and (3) attenuation of the signal at 2° and 3° is 0.5 and 0.0, respectively. Signals having wavelengths of $2\text{--}3^\circ$ are therefore captured in a qualitative way. Recognizing that the spectral representation of an eddy covers a large range of wavelengths with most of the energy located at lower wavenumbers, the authors estimate the minimum eddy size that such filtering characteristics will permit. They estimate that Gaussian eddies with e -folding scale—which is a form of eddy radius—less than 0.4° are significantly attenuated, those with scale $0.4^\circ\text{--}0.6^\circ$ are somewhat attenuated and those with scale greater than 0.6° are completely resolved. At midlatitudes, these scales correspond roughly to 40 and 60 km. Therefore, 60 km represents the minimum resolvable eddy scale for our dataset.

1.3.1 Processing steps

Time-averaging and spatial filtering

One of the processing steps commonly used to observe striations is time-averaging. The argument made for the use of a time-average is that the desired signal is so small in amplitude that it is easily masked in individual snapshots of the ocean. The averaging period used in this study is 4 yr, and therefore allows us to describe stationary striations. It also affords us the ability to break the 16-yr record examined in this study into 4 distinct time intervals.

In addition to time-averaging, it is common to apply a spatial high pass-filter to the time-averaged field in order to remove spatial trends irrelevant to the processes under examination (e.g., *Maximenko et al.* [2008]; *van Sebille et al.* [2011]). Given the desire to replicate processing steps leading to striations documented elsewhere, mean ADT is spatially high pass-filtered in a manner identical to that described by *Maximenko et al.* [2008]. This filtering process consists of several steps. First, a spatial low pass filter is applied to the time-averaged field in order to isolate low-wavenumber signals. We refer to this field as *LP1*. The filter is a two-dimensional, approximately isotropic Hanning window with 4.0° half-width. The result of this step is then subtracted from the original field, resulting in a high pass-filtered field. This field is then filtered with the same Hanning window, isolating low-frequency content missed in the first step. We refer to this field as *LP2*. Finally, the sum (*LP1* + *LP2*) is subtracted from the original field, resulting in a spatially high pass-filtered field. The effective magnitude response of this filtering process is shown in Figure 2.6.

Filtering the time-averaged field in the above manner is equivalent to time-averaging spatially high pass-filtered fields. This results from the linearity of both the averaging and high pass-filtering processes. A second but equally important point is that the gradient of a time-averaged field is equal to the time-average of the gradient of underlying fields. In particular, $\overline{u}$ estimated from time-averaged ADT is equal to the time-average of *u*-fields.

Identification and removal of eddy signatures

The general assumption in this study is that one can decompose the observed velocity signal into that due to eddies and that due to other processes, which may include jets. The basis for this decomposition comes from the fact that for the large-scale (> 30 km) ocean, the sum of forces acting on a parcel of fluid in the horizontal can be written as the sum of Coriolis and pressure forces [Wunsch and Stammer, 1998]. In combination with the hydrostatic relation, this geostrophic balance permits estimation of zonal and meridional components of surface velocity from dynamic height.

An improved representation of upper ocean dynamics is the so-called gradient wind balance. This is a three-way balance of Coriolis, horizontal pressure gradient and centripetal forces, where the latter term arises due to the curvature of parcel trajectories [Holton, 1979]. It can be shown that centripetal forces do not contribute significantly to the observed height signal. In particular, the magnitude of centripetal acceleration in the vicinity of eddies has been estimated at 1-3% that of the Coriolis acceleration. Given that eddies are the dominant signal in SSH, and given that the centripetal term is larger than or equal to other nonlinear terms in the equations of motion, as well as viscous terms and time-tendency terms, we conclude that the geostrophic balance is a good approximation.³ Therefore, the ocean surface velocities are well approximated by

$$u = -\frac{g}{f} \frac{\partial \eta}{\partial y} = -\frac{g}{f} \left(\frac{\partial \eta_{eddy}}{\partial y} + \frac{\partial \eta_{other}}{\partial y} \right) = u_{eddy} + u_{other} \quad (1.1)$$

$$v = +\frac{g}{f} \frac{\partial \eta}{\partial x} = +\frac{g}{f} \left(\frac{\partial \eta_{eddy}}{\partial x} + \frac{\partial \eta_{other}}{\partial x} \right) = v_{eddy} + v_{other}, \quad (1.2)$$

³Assessing the validity of the geostrophic approximation requires a systematic scaling of the equations of motion. The importance of ageostrophic accelerations are determined by the Rossby (Ro) and Reynolds (Re) numbers applicable to the flow. It can be shown that time-tendency, nonlinear advective and viscous accelerations scale as Ro, Ro, and Ro/Re times the Coriolis acceleration, respectively. Because the Rossby number is small (10^{-2}) and the Reynolds number large (10 – 10^5), we find these accelerations are of order 10^{-2} , 10^{-2} and 10^{-3} to 10^{-5} times the Coriolis acceleration. [The latter range results from scaling the viscous acceleration term and noting that horizontal and vertical eddy viscosities have different magnitudes. We expect a significant departure from these scales in the Ekman layer where winds modify the balance through increased vertical shear – i.e., the $\partial/\partial z (\nu_z \partial u/\partial z)$ term.] In summary, the error made in making the geostrophic approximation is at most a few percent.

where x and y are eastward and northward distances, respectively, g is the acceleration due to gravity, f is the Coriolis parameter, η is the absolute dynamic height, and u and v are zonal and meridional velocities, respectively. Here, we have used the substitution $\eta = \eta_{eddy} + \eta_{other}$ and, owing to the linearity of the equations, defined zonal geostrophic velocities due to eddies and other processes. In this study, we are interested in $\bar{u}$ since $\bar{v}$ approaches zero.

To determine if eddies give rise to striations in $\bar{u}$, it is necessary to identify eddies, isolate their ADT signature and quantify their contribution to the time-averaged field. One manner of determining the overall contribution of eddies to $\bar{u}$ is to simulate a set of Gaussian eddies with amplitudes, scales and lifetimes identical to those observed, compute a time-average of these fields and compare observed and simulated $\bar{u}$. This has been done and results in $\bar{u}$ estimated from Gaussian eddies having striations with similar appearance but smaller amplitude when compared with observed striations (Figure 2.2b). Eddies of parabolic shape [CCS2011] yield improved but comparable results (not shown). The discrepancy in the magnitude of striations, coupled with an imperfect alignment of striations when using eddies of a prescribed shape motivates use of an automated algorithm to identify and subtract eddies from individual ADT fields. This process is described below. We note that Chelton and Schlax offered to provide contours of eddies identified in their 2011-study. However, differences in the underlying dataset (SLA vs. ADT) and the spatial filtering applied to SLA fields prior to identifying eddies has led us to develop our own contour-identification algorithm but one that closely follows that described by CCS2011 (Appendix B.2).

Algorithm description

The algorithm used in this paper searches for closed contours at various height levels, starting at -100 cm and increasing in intervals of 0.25 cm in its search for anticyclonic eddies while starting at $+100$ cm and decreasing in intervals of 0.25 cm in its search for cyclonic eddies. Since height contours are coincident with streamlines of geostrophic flow, one might also interpret the eddies identified from this algorithm as “closed circulation cells.”

In identifying closed contours from the combination of the ADT record and eddy centers

in the eddy database, we require that Chelton-Schlax eddy centers be located within the closed contour, or perimeter, of the eddy identified with our algorithm. In addition, we allow closed regions of the eddy to include pixels out to a distance $2L$ from the eddy center, where L is the “radius of the circle that has the same area as the region within the closed contour of SSH with maximum average speed” [CCS2011]. Additional constraints on the selection of eddies are as follows: (1) eddy pixels (i.e., those pixels found within eddy perimeters) cannot be more than a distance D from one another, where D varies linearly between approximately 670 km at 20°S and 400 km at 30°S , and is 400 km for latitudes poleward of 30°S ; (2) eddy pixels must not be assigned to more than one eddy; (3) an object identified as an eddy must contain at least 8 but no more than 1000 eddy pixels; and (4) the closed contour of the object must include an extrema with sign determined by the polarity of the eddy in the eddy database. The extrema need not be the eddy center specified in the eddy database. Figure 2.5a illustrates the identification of eddy contours in ADT.

Having identified the perimeters of eddies, one proceeds to remove eddies from ADT fields in order to determine the role eddies play in creating striations. The removal of eddies from ADT was accomplished in the following manner. ADT values within eddy perimeters were replaced with those of a smoothed interpolated field. The smoothed interpolated field was generated by (1) assigning to eddy pixels within each eddy the value of the threshold used to obtain a closed contour and (2) spatially low pass-filtering this field with a 9-by-9-pixel (2.25° -by- 2.25°) Gaussian window having approximately 1.9° half-power width. The result of removing eddies and replacing them with pixels from the smoothed field is shown in Figure 2.5b.

Finally, in order to estimate average zonal geostrophic velocity attributed to eddies, we computed the time-average of ADT with eddies removed and subtracted this from the total observed time-averaged ADT field. This results in an estimate of mean ADT due to eddies alone and from which we calculate $\bar{u}$ due to eddies. We use the terms $\bar{u}_{total}$ and $\bar{u}_{eddy}$ to denote total and eddy-only $\bar{u}$ fields, and $\bar{u}_{diff} = \bar{u}_{total} - \bar{u}_{eddy}$ to refer to the difference.

Algorithm characteristics

It is possible that zonally-elongated objects are frequently identified by the algorithm. This would reduce a considerable fraction of the striation signal while calling into question the eddy-nature of ADT anomalies identified here. From illustrations of eddy contours, this does not appear to be the case. For example, two pronounced anisotropic structures are visible in Figure 2.5a near 27-30°S, 225-232°E. These positive and negative plume-like features are preserved by the contour identification and eddy-removal process (Figure 2.5b). While qualitative, this gives us some assurance the algorithm is properly distinguishing between jets and eddies. A more systematic assessment follows.

To determine if features selected by the algorithm are anisotropic, on average, we examined the histogram of eddy perimeters and plotted these as a function of distance from the centroids of the objects. Figure 2.7 illustrates the histogram of eddy perimeters relative to the centroids of identified eddies, where zonal and meridional distances have been normalized by the eddy scale L obtained from the database. One notes two characteristics from this figure. The first is that the histogram of eddy perimeters reveals little or no anisotropy in the shapes of identified eddies. The second is that the most probable perimeter distance is approximately $1.4L$, providing confidence that the threshold distance of $2L$ (described above) is reasonable.

Note that the perimeter distance is not the same as the eddy scale. The former describes the horizontal extent of the eddy signature in SSH while the latter describes the radius at which the rotational speed of the eddy is a maximum and has potential dynamical significance: fluid within this distance is trapped by the eddy if its circulation speed U exceeds the propagation speed c of the eddy [CCS2011].

Some limitations should be noted regarding the contour-identification and eddy removal procedure. These include the following: (1) the set of eddies identified and tracked are restricted to those eddies having lifetimes greater than or equal to 4 weeks; (2) a small percentage (1%) of eddies identified and tracked by Chelton and Schlax are not identified in the corresponding ADT fields either due to the fact that (i) the tracked eddies may not have a corresponding eddy signature in the altimeter record since CCS2011 allow for the reappearance of an eddy by looking

ahead several time steps, or (ii) the latitude and longitudes of eddy tracks have been smoothed [Section B.4 of CCS2011] to the extent that eddy centers no longer fall inside the perimeters of eddies identified here; (3) the contour-identification algorithm sometimes connects two visually distinct anomalies in spatially-filtered ADT fields; and (4) removal of eddies and the subsequent replacement of eddy pixels with values makes an assumption about the nature of the SSH field in the absence of eddies. Here, this assumption is that an ADT signature exists in the absence of an eddy. Despite these limitations, we believe this method to be useful in providing an estimate of the contribution of eddies to striations. We note that items (1),(2) and (4) underestimate dynamic height due to eddies, while item (3) has the potential of overestimating mean ADT due to eddies. In the Discussion section, we attempt to bound the energy due to eddies.

1.3.2 Characterization and comparison of $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$

In determining the contribution of eddies to striations in $\bar{u}$, we characterize and compare $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$. This is accomplished by computing variances and covariances of the fields. We compute the variances, $\sigma_{\bar{u}}^2$, in order to characterize their relative amplitudes and covariances to determine the degree of similarity amongst the fields. Here, $\sigma_{\bar{u}}$ represents a deviation from a spatial average and the averaging region is chosen to be the South Pacific.

Note that variances of eddy-only and difference fields do not add to equal the total variance. This is seen in the expression for the variance of $\bar{u}$:

$$\sigma_{\bar{u}_{total}}^2 = \sigma_{\bar{u}_{eddy}}^2 + \sigma_{\bar{u}_{diff}}^2 + 2\sigma_{\bar{u}_{eddy}\bar{u}_{diff}}, \quad (1.3)$$

where the last term is twice the covariance between eddy-only and difference fields. Normalizing this expression by the total variance yields an expression for the relative contributions of $\bar{u}_{eddy}$, $\bar{u}_{diff}$ and the covariance between the two fields. In particular, the fraction of variance explained by eddies is $\sigma_{\bar{u}_{eddy}}^2 / \sigma_{\bar{u}_{total}}^2$. Both observed and fractional variances are reported in the next section.

In addition to comparing relative magnitudes of $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$, correlation coefficients are used to compare the similarity of fields. In terms of variances and covariances, the

signed correlation coefficient is defined as

$$\rho_{i,j} = \frac{\sigma_{\bar{u}_i \bar{u}_j}^2}{\sigma_{\bar{u}_i} \sigma_{\bar{u}_j}} \quad (1.4)$$

[*Bendat and Piersol*, 2000]. Here, i and j , where $i \neq j$, are subscripts used to denote total, eddy-only or difference fields. A correlation coefficient close to +1.0 suggests the two fields are linearly related through a positive scale factor, while a negative value demonstrates that $\bar{u}$ fields have features common to both but opposite in sign; a value near zero demonstrates little-to-no linear relationship between the fields.

When computing the above quantities, it is helpful to subsample $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$ in an effort to obtain decorrelated samples. Given that decorrelation scales within optimally interpolated fields are 100-300 km—although this varies with latitude (Appendix A.2 of CCS2011)—and recognizing that the signal of interest is zonally-elongated, fields were sampled every one degree in latitude and every three degrees in longitude.

1.4 Results

This section examines the role eddies play in the generation of striation patterns. Given that eddies are found to account for a fraction of the signal present in $\bar{u}$, hypotheses posed by *Schlax and Chelton* [2008] are tested, including (1) the decay of zonal standard deviation with averaging period and (2) the amplitudes and scales of eddies most responsible for striations. First, we examine zonal and meridional scales of striations to demonstrate that striations examined in this study are comparable to those in other studies.

1.4.1 Zonal and meridional spectra of $\bar{u}$

Figure 2.9 depicts zonal and meridional spectra of $\bar{u}$ as a function of averaging period. As the averaging period increases, the energy in the spectra appears to move from higher to lower zonal wavenumbers, consistent with the view that westward-propagating anomalies dominate the time-averaged field [*Schlax and Chelton*, 2008]. Meridional spectra reveal a different pattern. One notes that for all averaging periods, a large portion of energy is located within the

0.002-0.003-km⁻¹ wavenumber band, corresponding to wavelengths of 330-500 km. By $T = 60$ weeks, the spectra reveal a single peak near 0.0025 km⁻¹ (400 km), a trend extending to larger averaging periods. The $\bar{u}$ fields examined in the remainder of the study are those having averaging periods of ~ 200 weeks (4 years) and therefore correspond to $\bar{u}$ containing large (> 1000 km) zonal and ~ 400 -km meridional scales. These zonal and meridional scales show good agreement with those observed elsewhere [Maximenko *et al.*, 2008; van Seville *et al.*, 2011].

1.4.2 Comparison of $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$

Figure 2.10 illustrates $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$ estimated from four 4-year averages of ADT in the South Pacific. Variances and covariances are shown in Table 2.1 and normalized variances and correlation coefficients are shown in Table 2.2. One observes a high degree of similarity between $\bar{u}_{total}$ and $\bar{u}_{eddy}$: correlation coefficients between the two fields range from 0.90 to 0.93. In addition, an appreciable fraction of variance is explained by eddies: values range from 0.46 to 0.57. The difference, $\bar{u}_{diff}$, is smaller in amplitude but contains residual energy at locations similar to $\bar{u}_{total}$. This is reflected in moderate-to-high correlation coefficients between $\bar{u}_{total}$ and $\bar{u}_{diff}$ (0.77–0.80). The similarity of $\bar{u}_{eddy}$ and $\bar{u}_{diff}$ is not as high, with correlation coefficients range from 0.45 to 0.58. Lastly, covariance estimates between eddy-only and difference fields account for 0.29–0.32 times the total variance (column 4 in Table 2.2), leaving a small fraction attributed to the difference (0.12–0.22).

1.4.3 Decay of $\sigma_{\bar{u}}$ with averaging period

Schlax and Chelton [2008] derive an expression for the variance of $\bar{u}$ due to eddies as a function of averaging period T . In this expression, the amplitudes and scales of eddies determine to a large extent the velocity variance, with the square-root of this quantity falling off as $1/T$. To determine whether the standard deviation $\sigma_{\bar{u}}$ decays as $1/T$, we examined the SSH record as a function of averaging period, computed $\sigma_{\bar{u}}$ for each averaging period and applied a least-squares regression to the data. Here, we computed $\bar{u}$ from time-averaged ADT as a function of averaging period, increasing in four-week intervals from 4 to 200 weeks (~ 4 years). We did this for both

$\bar{u}_{total}$ and $\bar{u}_{eddy}$.

It is worth noting that the expression derived by *Schlax and Chelton* [2008] is an expected value computed at a given point. In the present case, we have defined $\sigma_{\bar{u}}$ with respect to a spatial mean and it is therefore descriptive of the South Pacific region as a whole. The comparison between the two forms of $\sigma_{\bar{u}}$ is valid so long as samples within the region are selected from the same distribution as samples at any given location.

Figure 2.11 shows the standard deviation of zonal geostrophic velocity $\sigma_{\bar{u}}$ as a function of averaging period T starting January 1993, increasing in four-week intervals and ending December 1996. Note that the data are well-approximated by an equation of the form of $y = \alpha T^\beta$ where $\alpha = 9.301$ and $\beta = -0.437$. This is compared with a $1/T$ falloff suggested by *Schlax and Chelton* [2008]. In addition to values for the 1993-1996 time period, values of α and β for the other time periods are shown in Table 2.3. In all cases, the data follow a trend distinctly different from $1/T$, with values of β ranging between -0.282 and -0.447. This suggests processes giving rise to striations are not explained by randomly-positioned eddies, alone.

1.4.4 Class of eddies most responsible for striations

An additional hypothesis posed by *Schlax and Chelton* [2008] concerning eddies and striations is that eddies of large amplitude and scale are most responsible for the observed patterns and that these eddies are few. Recall that each eddy identified in this study can be cross-referenced to the set of eddies identified by CCS2011 and are therefore associated with a given amplitude, scale, lifetime, etc. One can therefore test this hypothesis by creating $\bar{u}$ -fields attributed to eddies of each amplitude and scale and examine $\bar{u}_{total}$, $\bar{u}_{eddy}$ and $\bar{u}_{diff}$ for all amplitudes and scales. In particular, eddies of a particular class contribute most to the observed striations if the variance of $\bar{u}_{diff}$ is minimized. (Strictly speaking, eddies of a certain class contribute most to the observed striations if the mean-squared value of $\bar{u}_{diff}$ is minimized. However, the mean of $\bar{u}_{diff}$ is so close to zero that one can achieve the same effect by simply minimizing the variance.)

Figure 2.12 depicts the fraction of total variance contained in $\bar{u}_{diff}$ (i.e., $\sigma_{\bar{u}_{diff}}^2 / \sigma_{\bar{u}_{total}}^2$) as a function of eddy amplitude and scale. In addition, it illustrates the correlation coefficient

between $\bar{u}_{total}$ and $\bar{u}_{eddy}$ as a function of eddy amplitude and scale. While some variation exists for different averaging periods, the mean describes the trends in $\sigma_{\bar{u}_{diff}}^2 / \sigma_{\bar{u}_{total}}^2$ well. One notes that the fractional variance is minimized by eddies of small-to-medium amplitude (1-9 cm) and medium-to-large scale (50-150 km). This trend is reflected among correlation coefficients, as well, where $\bar{u}_{eddy}$ due to eddies of these amplitudes and scales are most correlated with $\bar{u}_{total}$. Ranges of amplitudes and scales are defined as the amplitudes and scales for which the mean in fractional variance is less than 0.95 and the mean in correlation exceeds 0.4 (horizontal lines in Figure 2.12).

It is important to note that eddy amplitudes may be biased low due to the existence of spatial gradients in SLA fields whose horizontal scale exceeds that of the eddy, thereby hindering identification of eddy contours [CCS2011]. CCS2011 apply a two-dimensional high pass filter prior to estimating eddy amplitudes in an effort to remove such bias. However, because filter dimensions are large (10° and 20° half-widths in latitude and longitude, respectively) eddy bias remains. It is also possible that bias in eddy amplitudes enters as a result of identifying eddies in SLA referenced to 1993-1999 rather than ADT or SLA referenced to a longer-duration period; however, we have determined this to be insignificant (< 0.3 cm) for our application.

In order to obtain a more accurate picture of the amplitudes of eddies giving rise to striation patterns, a set of revised eddy amplitudes were obtained from the spatially high-pass filtered fields used in our study. Given that the filter used in our study has half-width equal to 4°, large-scale gradients are attenuated. Eddy amplitudes were obtained in the following manner. Amplitudes of anticyclonic (cyclonic) eddies were estimated by taking the magnitude of the difference between the maximum (minimum) SSH value within the eddy and the average height along the perimeter of the eddy identified during the contour-identification step. In this respect, the estimation of eddy amplitudes is the same as that described by CCS2011.

The fraction of total variance contained in $\bar{u}_{diff}$ and correlation coefficients based on revised eddy amplitudes are shown in Figure 2.13. Note that the trough (peak) in fractional variance (correlation) shifts from eddies of low amplitude to those of moderate amplitude (3-12 cm), where again the range of amplitudes is defined by those amplitudes for which the mean falls

below 0.95 (exceeds 0.4). This adjusts earlier estimates and suggests that eddies of moderate amplitude and scale contribute most to the observed patterns.

While these results define the type of eddy most responsible for the observed striations, they do not tell us the number of eddies falling into this category relative to the total number of eddies observed. Figure 2.8 illustrates the histogram of eddies in the South Pacific as a function of revised eddy amplitude and scale. This can be compared with Fig. 9c of CCS2011 for the global ocean and for their eddy amplitudes. The percentage of observations falling into the intersection of the two classes described above (i.e., 3-12 cm and 50-150 km) is approximately 75%, an appreciable percentage of the total number of observations (227,000). (Note that the scale and amplitude of an eddy may change over its lifetime.) From this, we conclude that striations are not the result of a small number of large-amplitude and large-scale eddies but rather a significant number of eddies with a broad range of amplitudes and scales.

1.5 Conclusions and discussion

Two conclusions emerge from this study: (1) eddies contribute significantly to the observed striations and (2) the model of random eddies proposed by *Schlax and Chelton* [2008] does not appear to be correct. We found that a large number of moderate-amplitude (3-12 cm) and moderate-scale (50-150 km) eddies contribute significantly to the observed striations and the standard deviation of $\overline{u}$ decays with averaging period with a rate different from the inverse relationship proposed by [*Schlax and Chelton*, 2008].

1.5.1 Limitations

This study has focused on striations in the Subtropical South Pacific Ocean. Such a focus naturally restricts conclusions and prevents extrapolation to other regions of the world. However, we argue that to the extent that dynamics in the South Pacific are similar to those in other mid-latitude, open-ocean basins, results should extend to these regions, as well.

Errors in the altimeter product

Measurement and systematic errors of altimeter data include sensor noise, atmospheric errors, sea state biases, orbit errors, and errors in the removal of inverse barometer and tidal signals [Martin, 2004]. While collectively these errors can exceed several centimeters, they are estimated to be 2 cm when averaged over time and spatial scales similar to those examined in this study [Cheney *et al.*, 1994]. In addition to measurement and systematic errors, the altimeter data used in this study suffer from errors resulting from interpolation of data between tracks. Recall that the data used in this study and referred to as the reference series consist of optimally-interpolated measurements from two altimeters. Because the separation between altimeter tracks can be large (e.g., 200–300 km for TOPEX/Poseidon and Jason-1 tracks in our region; see Fig. 61 of Chelton *et al.* [2001]), the error can be large. Thus, the reference series may not suitably resolve the ocean surface in some locations.

One method of gaining insight into the magnitude of this error is to compare reference and updated products during a time when data from four satellite altimeters are available. Pascual *et al.* [2006] have done this and report larger SSH anomalies and enhanced eddy kinetic energy when using the updated series. In addition, Beron-Vera *et al.* [2010] demonstrate qualitative differences between surface structures in reference and updated data (see, for example, their Fig. 4) but ultimately conclude that mixing inferred from the two products is similar. Our own comparison during October 2002–September 2005 (not shown) reveals that differences in SSH between the two products can be as large as 6 cm (r.m.s. differences of 0.7 cm). This provides a plausible mechanism by which SSH contours identified in this study (see section 2.1.2.1) might differ from those determined from the updated series, and hence the real ocean. We acknowledge this as a potential error.

Errors in mean dynamic topography

Another potential limitation is the use of MDT. We feel this is a limitation since, in addition to errors in the optimally-interpolated altimeter fields in the form of SLA, the MDT product used in the study [Rio *et al.*, 2011] (i.e., CMTD) may contain errors. For example, it is possible

that wind stress reanalyses contain errors that then propagate to MDT during the calculation of Ekman effects. In addition, it is notable that the spatial distribution of hydrological profiles from Argo floats (2002-2008) is substantially greater than those from CTD casts (1993-2008) (see Fig. 2 of *Rio et al.* [2011]). We mention this because it is possible the mesoscale signal in CMDT is more representative of the latter-half of the averaging period than the entire 1992-2008 period. Ideally, dynamic height could be estimated by subtracting a high-resolution geoid from altimeter measurements, allowing an independent comparison of geoid- and *in-situ*-based MDTs. However, geoid models with sufficient accuracy on the mesoscale do not presently exist. We expect such models will be available in the near future from, for example, data collected by the Gravity field and steady-state Ocean Circulation Explorer (GOCE) satellite. Finally, we note that a parallel study using altimeter-only measurements (i.e., SLA referenced to a 16-yr mean) has been conducted and yielded nearly identical results. These are shown in Table 2.2 and are denoted by asterisks. This may indicate that MDT contains both eddies and jet-like processes and that the relative ratio of these signals is similarly reflected in SLA.

Recall that ageostrophic terms in the equations of motion have been neglected in this study. While the geostrophic approximation is accurate for individual snapshots of the upper ocean, it has yet to be determined whether ageostrophic terms are negligible when averaged for many years. In particular, deviations from geostrophy, such as those resulting from eddy-eddy or eddy-mean flow interactions, may accumulate in time. We have not estimated the magnitude nor the spatial distribution of these deviations relative to the observed striations.

Algorithm limitations

The most significant limitation of this study is the ability of the eddy identification and removal algorithm to correctly identify eddies. Our ability to separate eddies from the background mesoscale field is limited by two factors. First, the combined eddy tracking, contour-identification and eddy removal process is thought to underestimate the magnitude of eddies. As discussed in Section 2, this results primarily from the contour-identification and eddy removal process, in which eddy pixels are replaced with those of a smoothed interpolated field. The

second limiting factor is that the algorithm incorrectly connects distinct features, resulting in more energy being attributed to eddies than is warranted. Given a latent jet signal and eddies with comparable magnitudes, we might expect our confidence in the aforementioned percentages (i.e., 46–57%) to decrease. To this end, we estimate bounds on the percentage of variance attributed to eddies using simulated eddies and jets.

We simulated 16 years (1993–2008) of Gaussian eddies with positions, lifetimes and horizontal scales identically specified by the tracked eddy database and amplitudes specified by revised eddy amplitudes. We then applied the algorithm to the eddy-only dataset. We find that approximately 27% of the simulated eddy variance is missed by the contour-identification and removal algorithm. Similarly, we simulated both eddies and stationary jets (the latter having amplitudes of 1 cm in dynamic height), and applied the algorithm to these fields. We find that the algorithm improperly attributes approximately 32% of the jet variance to eddies. Upper and lower bounds on the percentage of variance attributed to eddies can, therefore, be defined as 68% (lower) and 127% (upper) of $\sigma_{u_{eddy}}^2 / \sigma_{u_{total}}^2$. In conjunction with values specified in Table 2.2, this suggests 31–72% of the total variance in the striation signal can be explained by eddies. While broad, these bounds on the variance attributable to eddies describe important aspects of the striation signal. They demonstrate that eddies account for an appreciable fraction of the striation signal (at least 31%) but do not account for all of the jet-like signal (at most 72%). While eddies in this study have been defined as those with lifetimes ≥ 4 weeks, we believe that the variance in the striation signal decreases with decreasing eddy lifetime. This may point to the existence of latent β -plumes and/or MZJs.

It is notable that two plume-like structures are observed in ADT near 27–30°S, 225–232°E (Figure 2.5a). Jet-like plumes such as these might result from radiating instabilities of an eastern boundary current [Hristova *et al.*, 2008; Wang *et al.*, 2012], but given the distance from the South American coast, it is more likely the result of intense flow past seamounts [Rhines, 1994] or island archipelagos.

1.5.2 The contribution of eddies to striations in SST

We close this study with a reference to striations in a secondary dataset.

Figure 1.12 displays time-averaged SST during 2005-2008, where SST measurements are those from the Advanced Microwave Scanning Radiometer for the Earth Observing System (AMSR-E) and have been spatially high pass-filtered in the same manner as ADT. One notes the existence of zonal patterns in SST much like those seen in ADT. Indeed, contours of time-averaged ADT (2005-2008) coincide with contours in SST. The correlation coefficient between the two fields is 0.54 and increases when ADT is shifted west of its present location (not shown), implying that SST and ADT are highly correlated but that mesoscale SST anomalies lie west, on average, of mesoscale anomalies in ADT.

These results are consistent with earlier studies examining the relationship between SST and SSH in the North Pacific [Roemmich and Gilson, 2001; Qiu and Chen, 2005] and we note that a similar correlation and spatial offset have been found to exist between surface chlorophyll concentration and SLA in the South Pacific [Chelton *et al.*, 2011b]. Given the role of eddies in forming striations in ADT and given moderate correlation values between SST and ADT, it is likely that eddies contribute substantially to striations in SST. Similar arguments might be made for patterns observed in density fields derived from Argo floats [van Sebille *et al.*, 2011; Cravatte *et al.*, 2012] and wind measurements over the ocean surface [Maximenko *et al.*, 2010; Divakaran, 2011], given an observed relationship between upper ocean density anomalies and eddies [Roemmich and Gilson, 2001; Qiu and Chen, 2005] and surface winds and eddies [Park *et al.*, 2006; Small *et al.*, 2008; Chelton and Xie, 2010], respectively.

It is also notable that the striation signal in both SST and ADT appear to be influenced by bottom topography. This is evident in Figure 1.12, where bathymetric features shallower than 2500 m are overlaid in black. Note that large values of mean SST and ADT are observed east of the East Pacific Rise (running approximately North-South at 245–250°E) and striations in both datasets appear influenced by subsurface archipelagos—for example, those bathymetric features extending northwest from Easter Island (26°S, 250°E). It is likely that eddy tracks (Fig.1a of Schlax and Chelton [2008]) reveal the influence of bottom topography, as well. Such arguments

support the concept of preferred pathways for eddies [Scott *et al.*, 2008; Maximenko *et al.*, 2008] and may help explain the observed decay of $\sigma_{\overline{u}}$ with averaging period.

Acknowledgments We wish to thank the following individuals for valuable contributions to this study: Dudley Chelton, Michael Schlax, Nikolai Maximenko and Oleg Melnichenko. In addition, we are grateful for conversations with Prasanth Divakaran, Tom Farrar, Pavel Berloff, Shafer Smith, Patrice Klein and Lewis Rothstein. Finally, we wish to thank three anonymous reviewers for helpful comments and criticism on an early draft of this manuscript. ADT were produced by Segment for ALTimetry, Orbitography and precise localisation/Developing Use of ALtimetry for Climate Studies (SSALTO/DUACS) and distributed by Archivage, Validation, Interpretation des donnees des Satellite Oceanographiques (AVISO)/Collecte Localisation Satellites (CLS) with support from Centre National d'Etudes Spatiales (CNES). The database of tracked eddies was developed by Chelton and Schlax and is available at <http://cioss.coas.oregonstate.edu/eddies/>. Microwave SST measurements were provided by Chelle Gentemann of Remote Sensing Systems (RSS). This research has been made possible through a grant (NNX10AN69H) from the National Aeronautics and Space Administration.

List of References

- Afanasyev, Y. D., S. O’Leary, P. B. Rhines, and E. Lindahl (2012), On the origin of jets in the ocean, *Geophys. Astrophys. Fluid Dyn.*, 106(2), 113–137.
- Baldwin, M. P., P. B. Rhines, H.-P. Huang, and M. E. McIntyre (2007), The jet-stream conundrum, *Science*, 315, 467–468, doi:10.1126/science.1131375.
- Bendat, J. S., and A. G. Piersol (2000), *Random Data: Analysis and Measurement Procedures*, Wiley Series in Probability and Statistics, 3rd ed., 594 pp., John Wiley and Sons, New York.
- Berloff, P., I. Kamenkovich, and J. Pedlosky (2009), A mechanism of formation of multiple zonal jets in the oceans, *J. Fluid Mech.*, 628, 395–425, doi:10.1017/S0022112009006375.
- Berloff, P., S. Karabasov, J. T. Farrar, and I. Kamenkovich (2011), On latency of multiple zonal jets in the oceans, *J. Fluid Mech.*, 686, 534–567, doi:10.1017/jfm.2011.345.
- Beron-Vera, F. J., M. J. Olascoaga, G. J. Goni, 2010: Surface ocean mixing inferred from different multisatellite altimetry measurements, *J. Phys. Oceanogr.*, 40, 2466–2480, doi: 10.1175/2010JPO4458.1.
- Buckingham, C., and P. Cornillon (2010), Observed quasi-zonal banded structures in SST frontal probability: why do we see them?, Ocean Sciences 2010 Conference, Portland, Oregon.
- Centurioni, L., J. Ohlmann, and P. Niiler (2008), Permanent meanders in the California Current System, *J. Phys. Ocean.*, 38, 1690–1710, doi:10.1175/2008JPO3746.1.
- Chelton, D. B., and M. G. Schlax (1996), Global observations of oceanic Rossby waves, *Science*, 272, 234–238.
- Chelton, D. B., J. C. Ries, B. J. Haines, L.-L. Fu, and P. S. Callahan (2001), Satellite altimetry, in *Satellite Altimetry and Earth Sciences: A Handbook of Techniques and Applications*, edited by L.-L. Fu and A. Cazenave, pp. 1–131, Academic, San Diego, Calif.

- Chelton, D. B., M. G. Schlax, R. M. Samelson, and R. A. de Szoeke (2007), Global observations of large oceanic eddies, *Geophys. Res. Lett.*, *34*, L15606, doi:10.1029/2007GL030812.
- Chelton, D. B., and S.-P. Xie (2010), Coupled ocean-atmosphere interaction at oceanic mesoscales, *Oceanog.*, *23*(4), 52–69, doi:10.5670/oceanog.2010.05.
- Chelton, D. B., M. G. Schlax, and R. M. Samelson (2011a), Global observations of nonlinear mesoscale eddies, *Prog. Oceanog.*, *91*, 167–216, doi:10.1016/j.pocean.2011.01.002.
- Chelton, D. B., P. Gaube, M. G. Schlax, J. J. Early, and R. M. Samelson (2011b), The influence of nonlinear mesoscale eddies on near-surface oceanic chlorophyll, *Science*, *334*(6054), 328–332.
- Cheney, R., L. Miller, R. Agreen, N. Doyle, and J. Lillibridge (1994), TOPEX/POSEIDON: the 2-cm solution, *J. Geophys. Res.*, *99*, 24 555–24 563, doi:10.1029/94JC02050.
- Cho, J. Y-K., and L. M. Polvani (1996), The emergence of jets and vortices in freely evolving, shallow-water turbulence on a sphere, *Phys. Fluids*, *8*, 5531–1552.
- Cravatte, S., and W. Kessler, and F. Marin (2012), Intermediate zonal jets in the tropical Pacific Ocean observed by Argo floats, *J. Phys. Oceanogr.*, doi:10.1175/JPO-D-11-0206.1.
- Di Lorenzo, E., A. Belmadani, H. Luo, O. Melnichenko, N. Schneider, and N. Maximenko (2012), The emergence of striations in the North Pacific circulation, Ocean Sciences 2012 Conference, Salt Lake City, Utah.
- Divakaran, P., and G. B. Brassington (2011), Arterial ocean circulation of the southeast Indian Ocean, *Geophys. Res. Lett.*, *38*, L01802, doi:10.1029/2010GL045574.
- Divakaran, P. (2011), Mesoscale circulation and variability of the Indian Ocean, Ph.D. thesis, School of Earth Sc., Univ. of Melbourne, Melbourne, Australia.
- Ducet, N., P. Y. LeTraon, and G. Reverdin (2000), Global high-resolution mapping of ocean circulation from TOPEX/Poseidon and ERS-1 and -2, *J. Geophys. Res.*, *105*, 477–498.

- Firing, E. (1987), Deep zonal currents in the central equatorial Pacific, *J. Mar. Res.*, *45*, 791–812.
- Gouriou, Y., T. Delcroix, and G. Eldin (1987), Upper and intermediate circulation in the western equatorial Pacific Ocean in October 1999 and April 2000, *Geophys. Res. Lett.*, *33*, L10603, doi:10.1029/2006GL025941.
- Hogg, N. G., and W. Brechner Owens (1999), Direct measurement of the deep circulation within the Brazil Basin, *Deep-Sea Res. II*, *46*, 335–353.
- Holton, J. R. (1979), *An Introduction to Dynamic Meteorology*, 3rd ed., 391 pp., Elsevier Academic Press, San Diego.
- Hristova, H. G., J. Pedlosky, and M. A. Spall (2008), Radiating instability of a meridional boundary current, *J. Phys. Oceanogr.*, *38*, 2294–2307, doi:10.1175/2008JPO3853.1.
- Huang, H.-P., A. Kaplan, E. N. Curchitser, and N. A. Maximenko (2007), The degree of anisotropy for mid-ocean currents from satellite observations and an eddy-permitting model simulation, *J. Geophys. Res.*, *36*, C09005, doi:10.1029/2007JC004105.
- Ivanov, L., C. Collins, and T. Margonina (2009), System of quasi-zonal jets off California revealed from satellite altimetry, *Geophys. Res. Lett.*, *36*, L03609, doi:10.1029/2008GL036327.
- Ivanov, L., C. Collins, T. Margonina, and V. Eremeev (2010), Nonlinear Rossby waves off California, *Geophys. Res. Lett.*, *37*, L13602, doi:10.1029/2010GL043708.
- Kamenkovich, I., P. Berloff, and J. Pedlosky (2009), Role of eddy forcing in the dynamics of multiple zonal jets in a model of the North Pacific, *J. Phys. Ocean.*, *39*, 1361–1379, doi:10.1175/2008JPO4096.1.
- Le Traon, P. Y., Y. Faugère, F. Hernandez, F. M. J. Dorandeu, and M. Ablain (2003), Can we merge GEOSAT follow-on with TOPEX/Poseidon and ERS-2 for an improved description of the ocean circulation?, *J. Atmos. Oceanic Technol.*, *20*, 889–895.
- Maltrud, M. E., and G. K. Vallis (1991), Energy spectra and coherent structures in

- forced two-dimensional and beta-plane turbulence, *J. Fluid Mech.*, 228, 321–342, doi:10.1017/S0022112091002720.
- Martin, S. (2004), *An Introduction to Ocean Remote Sensing*, 426 pp., Cambridge University Press, Cambridge.
- Maximenko, N., B. Bang, and H. Sasaki (2005), Observational evidence of alternating zonal jets in the World Ocean, *Geophys. Res. Lett.*, 32, L12607, doi:10.1029/2005GL022728.
- Maximenko, N., O. V. Melnichenko, P. P. Niiler, and H. Sasaki (2008), Stationary mesoscale jet-like features in the ocean, *Geophys. Res. Lett.*, 35, L08603, doi:10.1029/2008GL033267.
- Maximenko, N., O. V. Melnichenko, S.-P. Xie, and N. Schneider (2010), Atmospheric signatures of oceanic striations, poster, Altimetry for Oceans and Hydrology - OSTST Meeting, Lisbon, Portugal.
- Melnichenko, O. V., N. A. Maximenko, N. Schneider, and H. Sasaki (2010), Quasi-stationary striations in basin-scale oceanic circulation: vorticity balance from observations and eddy-resolving model, *Ocean Dyn.*, 60, 653–666.
- Nakano, H., and H. Hasumi (2005), A series of zonal jets embedded in the broad zonal flows in the Pacific obtained in eddy-permitting ocean general circulation models, *J. Phys. Ocean.*, 35, 474–488.
- Nowlin, W. D. , Jr. and M. Clifford (1982), The kinematic and thermohaline zonation of the Antarctic Circumpolar Current at Drake Passage, *J. Mar. Res.*, 40, 481–507.
- Orsi, A. H., T. Whitworth, III, and W. D. Nowlin, Jr. (1995), On the meridional extent and fronts of the Antarctic Circumpolar Current, *Deep-Sea Res. I*, 42, 641–673.
- Panetta, R. L. (1993), Zonal jets in wide baroclinically unstable regions: persistence and scale selection, *J. Atmos. Sci.*, 50, 2073–2106.
- Park, K., P. Cornillon, and D. L. Codiga (2006), Modification of surface winds near ocean fronts:

- Effects of Gulf Stream rings on scatterometer (QuikSCAT, NSCAT) wind observations, *J. Geophys. Res.*, *111*, C03021, doi:10.1029/2005JC003016.
- Pascual, A., Y. Faugère, G. Larnicol, and P.-Y. Le Traon (2006), Improved description of the ocean mesoscale variability by combining four satellite altimeters, *Geophys. Res. Lett.*, *33*, L02611, doi:10.1029/2005GL024633.
- Pedlosky, J. (1975), On secondary baroclinic instability and the meridional scale of motion in the ocean, *J. Phys. Ocean.*, *5*, 603–607.
- Qiu, B., and S. Chen (2005), Eddy-induced heat transport in the subtropical North Pacific from Argo, TMI, and altimetry measurements, *J. Phys. Ocean.*, *35*, 458–473.
- Rhines, P. B. (1975), Waves and turbulence on a beta-plane, *J. Fluid Mech.*, *69*, 417–443.
- Rhines, P. B. (1994), Jets, *Chaos*, *4*, 313–339.
- Rio, M. H., S. Guinehut, and G. Larnicol (2011), New CNES-CLS09 global mean dynamic topography computed from the combination of GRACE data, altimetry, and in situ measurements, *J. Geophys. Res.*, C07018, doi:10.1029/2010JC006505.
- Roemmich, D., and J. Gilson (2001), Eddy transport of heat and thermocline waters in the North Pacific: a key to interannual/decadal climate variability?, *J. Phys. Ocean.*, *31*, 675–687.
- Schlax, M. G., and D. B. Chelton (2008), The influence of mesoscale eddies on the detection of quasi-zonal jets in the ocean, *Geophys. Res. Lett.*, *35*, L24602, doi:10.1029/2008GL035998.
- Scott, R. B., B. K. Arbic, C. L. Holland, A. Sen, and B. Qiu (2008), Zonal versus meridional velocity variance in satellite observations and realistic and idealized ocean circulation models, *Ocean Modell.*, *23*, 102–112, doi:10.1016/j.ocemod.2008.04.009.
- Small, R. J., S. P. de Szoeke, S. P. Xie, L. O'Neill, H. Seo, Q. Song, P. Cornillon, M. Spall, S. Minobe (2008), Air-sea interaction over fronts and eddies, *Dyn. Atm. Oceans*, *45*, 274–319, doi:10.1016/j.dynatmoce.2008.01.001.

- Sokolov, S., and S. R. Rintoul (2007), Multiple jets of the Antarctic Circumpolar Current south of Australia, *J. Phys. Ocean.*, 37, 1394–1412, doi:10.1175/JPO3111.1.
- Treguier, A. M., and R. L. Panetta (1994), Multiple zonal jets in a quasigeostrophic model of the Antarctic Circumpolar Current, *J. Phys. Oceanogr.*, 24, 2263–2277.
- van Sebillie, E., I. Kamenkovich, and J. K. Willis (2011), Quasi-zonal jets in 3D Argo data of the northeast Atlantic, *Geophys. Res. Lett.*, 38, L02606, doi:10.1029/2010GL046267.
- Wang, J., M. A. Spall, G. R. Flierl, and P. Malanotte-Rizzoli (2012), A new mechanism for the generation of quasi-zonal jets in the ocean, *Geophys. Res. Lett.*, 39, L10601, doi: 10.1029/2012GL051861.
- Williams, G. P. (1978), Planetary circulations: 1. Barotropic representation of Jovian and terrestrial turbulence, *J. Atmos. Sci.*, 35, 1399–1426.
- Wunsch, C., and D. Stammer (1998), Satellite altimetry, the marine geoid, and the ocean general circulation, *Annu. Rev. Earth Planet. Sci.*, 26, 219–253.

Figure 1.1. Global maps of time-averaged zonal geostrophic velocity, $\bar{u}$, estimated from (a) observed and (b) simulated SSH, 1993-1996. Observed SSH consist of AVISO/CLS ADT and were spatially high pass-filtered prior to estimating zonal geostrophic velocity. Simulated SSH were obtained by superimposing Gaussian eddies with amplitude, scale, position and lifetime identically specified by the eddy database [CCS2011] onto an otherwise flat ocean. Black boxes denote the region examined in this study (20°S-50°S, 200°E-280°E).

Figure 1.2. (a) A westward-propagating, anticyclonic eddy in the Southern Hemisphere has zonal and meridional velocities u and v along the perimeter of the eddy as shown. (b) As velocity fields are averaged in time, $\bar{u}$ has small but finite amplitude while $\bar{v}$ approaches zero. The effect of the eddy is to create striations in $\bar{u}$. Illustration after *Scott et al.* [2008].

Figure 1.3. Effective magnitude response of the filtering process used in this study. The filter is nearly isotropic in latitude θ and longitude ϕ in units of deg. Wavelengths in units of km assume $1 \text{ deg} = 111.2 \text{ km}$ and, thus, correspond to meridional distances. The shaded box highlights wavelengths of 300-500 km.

Figure 1.4. Illustration of (a) the contour-identification and (b) eddy removal algorithm applied to filtered ADT in the South Pacific, valid January 6, 1993. Black dots denote eddy centers obtained from the eddy database.

Figure 1.5. Histogram of eddy perimeter locations determined by the contour-identification algorithm. Eddy locations are shown as scaled distances from eddy centroids and distances are normalized by the horizontal eddy scale L as defined in the eddy database.

Figure 1.6. (a) Zonal and (b) meridional spectra of $\bar{u}$ in the South Pacific as a function of averaging period. Solid horizontal lines mark averaging periods of 4 years while vertical dashed lines highlight zonal and meridional wavelengths of 1000 and 400 km, respectively. Spectra were created using all available data between January 1, 1993 and December 31, 2008. Individual 4-yr averaging periods show similar structure. Note that spectral contours are illustrated on a logarithmic scale and units are $\text{cm}^2 \text{s}^{-2} \text{km}$.

Figure 1.7. Comparison of (a-d) $\bar{u}_{total}$, (e-h) $\bar{u}_{eddy}$ and (i-l) $\bar{u}_{diff}$ estimated from ADT during four 4-yr averaging periods. Contour lines are shown every 2 cm s^{-1} and contours in difference plots (i-l) correspond to those of $\bar{u}_{total}$.

Figure 1.8. (a) Standard deviation of $\overline{u}_{total}$ and $\overline{u}_{eddy}$ as functions of averaging period, beginning January 1993, increasing in 4-wk intervals and ending December 1996. Solid line: least-squares fit, $y = \alpha T^\beta$ with $\alpha = 9.301$ and $\beta = -0.437$. Dashed line: $y = \alpha T^\beta$ with $\alpha = 9.301$ and $\beta = -1$ as proposed by a model of random-positioned eddies. (b) Same as for (a) but for all time periods and only showing least-squares fits and proposed falloffs. Regression coefficients are listed in Table 2.3.

Figure 1.9. Fraction of total variance contained in $\overline{u}_{diff}$ (blue, top) and correlation coefficients between $\overline{u}_{total}$ and $\overline{u}_{eddy}$ (red, bottom) as a function of eddy (a) amplitude and (b) scale. Thin lines illustrate values for each averaging period and thick, dashed lines illustrate the means. Bin widths are 1 cm and 10 km for eddy amplitude and scale, respectively.

Figure 1.10. Fraction of total variance contained in $\bar{u}_{diff}$ (blue, top) and correlation coefficients between $\bar{u}_{total}$ and $\bar{u}_{eddy}$ (red, bottom) as a function of revised eddy amplitude.

Figure 1.11. Histogram of eddies in the South Pacific (1992-2008) as a function of revised amplitude and scale. The black box highlights the intersection of amplitudes and scales most correlated with observed striations.

Figure 1.12. Time-averaged SST (2005-2008) with contours of time-averaged ADT (2005-2008) overlaid. Contours are given every 0.5 cm between -1 and 1 cm and both SST and ADT fields have been spatially high pass-filtered in a manner following *Maximenko et al.* [2008]. Bathymetric features shallower than 2500 m are highlighted in black.

Table 1.1. Variances and covariances of $\overline{u}_{total}$, $\overline{u}_{eddy}$ and $\overline{u}_{diff}$. Units are $\text{cm}^2 \text{s}^{-2}$.

Avg. Period	$\sigma_{\overline{u}_{total}}^2$	$\sigma_{\overline{u}_{eddy}}^2$	$\sigma_{\overline{u}_{diff}}^2$	$\sigma_{\overline{u}_{total}\overline{u}_{eddy}}^2$	$\sigma_{\overline{u}_{total}\overline{u}_{diff}}^2$	$\sigma_{\overline{u}_{eddy}\overline{u}_{diff}}^2$
1993–1996	0.758	0.351	0.170	0.470	0.288	0.119
1997–2000	0.972	0.506	0.187	0.646	0.326	0.139
2001–2004	1.707	0.934	0.233	1.204	0.503	0.270
2005–2008	2.286	1.301	0.283	1.652	0.634	0.351

Table 1.2. Fractional variances and correlation coefficients between $\overline{u}_{total}$, $\overline{u}_{eddy}$ and $\overline{u}_{diff}$. Results of applying the contour-identification/removal algorithm to SLA (referenced to an Oct. 1992–Dec. 2008 mean) are highlighted by asterisks.

Avg. Period	$\sigma_{\overline{u}_{eddy}}^2/\sigma_{\overline{u}_{total}}^2$	$\sigma_{\overline{u}_{diff}}^2/\sigma_{\overline{u}_{total}}^2$	$2\sigma_{\overline{u}_{eddy}\overline{u}_{diff}}^2/\sigma_{\overline{u}_{total}}^2$	$\rho_{total,eddy}$	$\rho_{total,diff}$	$\rho_{eddy,diff}$
1993–1996	0.46	0.22	0.32	0.91	0.80	0.49
1997–2000	0.52	0.19	0.29	0.92	0.77	0.45
2001–2004	0.55	0.14	0.31	0.95	0.80	0.58
2005–2008	0.57	0.12	0.31	0.96	0.79	0.58
1993–1996*	0.41	0.23	0.36	0.92	0.85	0.58
1997–2000*	0.49	0.23	0.28	0.90	0.76	0.40
2001–2004*	0.58	0.19	0.23	0.91	0.70	0.35
2005–2008*	0.61	0.15	0.24	0.93	0.69	0.38

Table 1.3. Parameters α and β obtained from least-squares fits of $\sigma_{\overline{u}_{total}}$ and $\sigma_{\overline{u}_{eddy}}$ during four time periods. The fit has the form $y = \alpha T^\beta$, where T is the averaging period in weeks. Units of α are $\text{cm s}^{-1} \text{wks}^{-\beta}$.

Time Period	$\alpha(total)$	$\beta(total)$	$\alpha(eddy)$	$\beta(eddy)$
1993-1996	9.301	-0.437	7.530	-0.461
1997-2000	10.216	-0.447	9.022	-0.488
2001-2004	8.144	-0.357	6.490	-0.366
2005-2008	6.695	-0.282	5.427	-0.295

MANUSCRIPT 2

Zonal bands in microwave sea surface temperature

by

Christian E. Buckingham¹ ; Peter C. Cornillon² ; Kelsey M. Obenour³

to be submitted to the *Journal of Geophysical Research – Oceans*.

¹Ph.D. Candidate, Graduate School of Oceanography, University of Rhode Island, South Ferry Road, Narragansett, RI 02882. E-mail: cbuckingham@gso.uri.edu

²Professor, Graduate School of Oceanography, University of Rhode Island, South Ferry Road, Narragansett, RI 02882.

³Masters Candidate, Graduate School of Oceanography, University of Rhode Island, South Ferry Road, Narragansett, RI 02882.

2.1 Abstract

A multi-year (2002–2011) average of microwave measurements of sea surface temperature (SST) from the Advanced Microwave Scanning Radiometer reveals bands in all ocean basins. Bands are quasi-zonal patterns observed in both mean equatorward SST gradient and SST front probability computed over the nine-year time period and characterized by repeated meridional separation distances. Inspection of seven-day averages of SST in five subtropical basins reveal that high gradient events (HGEs) explain a significant fraction (0.19–0.38) of the variance in mean equatorward SST gradient. HGEs are defined as isolated, mesoscale frontal features whose equatorward SST gradient exceeds $|\pm 0.0048^{\circ}\text{C km}^{-1}|$ above the background temperature gradient and whose total connected area exceeds 0.5 degrees squared. Time-longitude plots demonstrate that the features propagate westward with speeds comparable to mesoscale eddies tracked by *Chelton et al.* [2011a]. An analysis of SST, sea surface height (SSH) and temperature measurements made from Argo profiling floats demonstrate that HGEs arise due to (1) contrasting temperatures of neighboring anomalies and (2) the strain field associated with the eddies. Together, these results demonstrate mesoscale eddies play important roles in the generation of zonal bands in microwave SST. Lastly, motivated by an overlay of bands in SST and striations in mean dynamic topography, an examination of the decay of standard deviation with averaging period demonstrates that patterns are persistent in time.

2.2 Introduction

One of the characteristics of geophysical flow is its tendency to develop zonally coherent structures [*Rhines*, 1994]. Examples of these in the ocean include the Equatorial Under Current, North and South Equatorial Counter-Currents, Mediterranean outflow, the Gulf Stream, Kuroshio and Antarctic Circumpolar Currents. Their existence is predicted by theory and supported by observations. In contrast, a class of zonal structures predicted by theory [*Rhines*, 1975; *Baldwin et al.*, 2007; *Dritschel and McIntyre*, 2008; *Berloff et al.*, 2009] but not yet established to exist in the ocean are multiple zonal jets (MZJs).

MZJs are mesoscale coherent zonal structures thought to exist in the global ocean [Berloff *et al.*, 2009, 2011]. Alternating in sign in meridional cross-sections of the upper ocean, characterized by repeated meridional spacing and extending for thousands of kilometers, MZJs have the potential to modify heat, momentum and biogeochemical properties in the ocean via circulation patterns and interaction with mesoscale eddies [Kamenkovich *et al.*, 2009; Robinson, 1983]. The main distinctions between MZJs and those flows mentioned above are their alternating appearance and regular spacing.

2.2.1 Background

Patterns resembling MZJs have recently been observed in the ocean. Referred to as striations [Maximenko *et al.*, 2008] or quasi-zonal jets [Ivanov *et al.*, 2009; van Seville *et al.*, 2011; Schlax and Chelton, 2008], the jet-like patterns have been observed in time-averaged sea level anomaly (SLA) [Maximenko *et al.*, 2005; Huang *et al.*, 2007; Scott *et al.*, 2008; Ivanov *et al.*, 2009, 2010], mean dynamic topography (MDT) [Maximenko *et al.*, 2008; Centurioni *et al.*, 2008; Buckingham and Cornillon, 2013], sea surface temperature (SST) [Buckingham and Cornillon, 2010; Divakaran and Brassington, 2011] and Argo float measurements [van Seville *et al.*, 2011]. Two studies of particular note are those by Hogg and Owens [1999] and Cravatte *et al.* [2012]. Hogg and Owens [1999] find that subsurface isopycnal floats at 2500- and 4000-m depths translate along quasi-zonal trajectories and Cravatte *et al.* [2012] uncover a similar result from the displacement of Argo profiling floats at parking depths (i.e., 1000 and 1500 m) along the Equator.

While supportive of MZJs, these studies suffer from an ambiguity created by mesoscale oceanic phenomena. Developing a model of Gaussian eddies, Schlax and Chelton [2008] demonstrate that westward propagating eddies with completely random position but with characteristics estimated from observed eddies [Chelton *et al.*, 2007] create jet-like patterns nearly identical to those observed. Furthermore, Buckingham and Cornillon [2013] demonstrated that a significant fraction of variance in time-averaged zonal geostrophic velocity in the South Pacific can be attributed to eddies tracked by Chelton *et al.* [2011a] but that a small percentage may not be accounted for by mesoscale eddies.

Many of the observational studies described above make use of altimeter-derived sea surface height (SSH) in their investigations of striations. While SSH is of immense value in understanding the dynamics of striations, the sparse sampling of the altimeter ultimately limits analysis (Section 4.1 of *Buckingham and Cornillon* [2013]). Float-based studies suffer from similar undersampling problems. For this reason, it is worth considering the existence of zonal patterns in microwave sea surface temperature (SST), a dataset with significantly higher temporal and spatial coverage and whose measurements are independent of the altimeter- and Argo-derived observations.

2.2.2 A brief overview of microwave SST

Passive microwave sensors are unique in their ability to measure ocean surface temperatures from space. While infrared measurements of SST are affected significantly by atmospheric water vapor and precipitation, microwave radiation emitted from Earth’s surface penetrates clouds, providing an unhindered view of Earth for a large fraction of the time the sensor is in orbit. However, this comes at a cost: upwelling radiation from Earth in the microwave is several orders of magnitude less than in the infrared such that microwave sensors must sample from a larger area in order to obtain comparable thermal resolution. The Advanced Microwave Scanning Radiometer for the Earth Observing System (AMSR-E) on-board the Aqua spacecraft had an effective footprint of 56 km, a swath width of approximately 1450 km [*Martin*, 2004] and provided approximately 89% coverage each day (98% coverage every 2 days) [*Chelton and Wentz*, 2005]⁴. In contrast, the repeat orbit of the altimeter occurs at most once every 10 days and distances between altimeter tracks can be large—greater than 300 km for TOPEX/Poseidon tracks [*Chelton et al.*, 2001]. The merger and optimal-interpolation of altimeter data overcomes this limitation to some degree by combining measurements from multiple spacecraft but assumes certain temporal and spatial decorrelation scales during interpolation.

Buckingham and Cornillon [2010] demonstrated the existence of zonally-elongated bands in the probability of front detection derived from AMSR-E SST (hereafter, SST front proba-

⁴The operational lifetime of the AMSR-E sensor ended prematurely following antenna failure on October 4, 2011.

bility). When computed over a multi-year averaging period, bands in high front probability were found to resemble striations in zonal geostrophic velocity computed from mean dynamic ocean topography (MDOT) [Maximenko *et al.*, 2008]. We reproduce the result of *Buckingham and Cornillon* [2010] in Figure 2.1 but extend the averaging period associated with the AMSR-E record. A noteworthy observation is that the patterns are similar in appearance despite being obtained from different averaging periods—i.e., 2002–2011 for AMSR-E and 1992–2002 for MDOT. The existence of zonal patterns in a secondary and independent dataset and whose appearance resembles striations in MDOT highlights the potential for microwave SST to provide complimentary and additional information regarding quasi-zonal jet-like features.

2.2.3 Focus of the study

In this study, we demonstrate the existence of bands in microwave SST. We do this in two forms: mean equatorward SST gradient and SST front probability. We focus this study on the occurrence of large gradients observed in subtropical oceans and ask the following questions: (1) what gives rise to large equatorward gradients in SST? and (2) what fraction of the banded signal can be explained by these large gradient events? Finally, motivated by the similarity of bands and striations (cf. Figure 2.1), we seek to determine if patterns in SST and SSH are persistent in time.

The outline of this study is as follows. In section 2, we describe datasets and processing methods used in the study. In section 3, we demonstrate the existence of zonal bands in microwave SST and seek to relate large gradient events observed in subtropical oceans to zonal bands in time-averaged fields. In section 4, we estimate the persistence of patterns in both microwave SST and SSH and conclude this study with a discussion in section 5.

2.3 Methods

In the following section, we associate zonal bands in microwave SST with mesoscale eddies. Here, as in an earlier study [*Buckingham and Cornillon*, 2013], we use the term mesoscale eddy to refer to predominantly westward-propagating anomalies seen in the altimeter record

and described by *Chelton and Schlax* [1996], *Chelton et al.* [2007] and *Chelton et al.* [2011a]. Again, we recognize the importance of distinguishing between baroclinic Rossby waves and mesoscale eddies in the context of striations/bands but use the term to be consistent with the literature [*Schlax and Chelton*, 2008; *Scott et al.*, 2008; *Maximenko et al.*, 2008; *Buckingham and Cornillon*, 2013].

2.3.1 Dataset descriptions

The primary data used in this study consist of Level 2P microwave measurements of SST from the AMSR-E sensor produced by Remote Sensing Systems (RSS)⁵. We used Version-7 mapped and swath SST measurements, with the mapped product having a dataset spatial resolution of 0.25° and swath measurements provided at a resolution of $\sim 0.1^\circ$ (10-km). (Recall, the inherent resolution of AMSR-E SST is 56 km.)

To complement microwave measurements of SST, we examined altimeter-derived measurements of SSH at different points in the study. Along-track SSH were provided by the Delft Institute for Earth-Oriented Space research (DEOS) and consist of SLA from the Jason-1 spacecraft. Mapped SSH consist of merged, optimally-interpolated absolute dynamic topography (ADT) distributed by Archiving, Validation, Interpretation des donnees des Satellite Oceanographiques (AVISO)/Collecte Localisation Satellites (CLS) with support from Centre National d'Etudes Spatiales (CNES) and referred to as the reference series. These data are the sum of altimeter-only SLA and a 1992-2008 estimate of MDT [*Rio et al.*, 2011]. Along-track measurements are sampled at 1-Hz (resulting in a nominal 10-km resolution product), while the mapped data are provided on a 0.25° -grid. See *Chelton et al.* [2011a] and *Buckingham and Cornillon* [2013] for discussions of the resolution capability of the mapped product.

In addition to the aforementioned datasets, we examined temperature measurements from vertically profiling floats and used a database of tracked eddies [*Chelton et al.*, 2011a]. The former data were made available by the International Argo Program and consist of temperature measurements made within the upper 1500 m of the ocean. Climatological profiles from the

⁵For a definition of Level 2P products, see <https://www.ghrsst.org/data/data-descriptions/l2p-observations/>.

World Ocean Atlas 2009 (WOA2009) were then used to obtain anomalous temperatures within the upper ocean. The database of tracked eddies were used within the study in order to link large gradients in SST to specific oceanic phenomena. In doing so, we estimate the speeds of mesoscale eddies having lifetime ≥ 16 weeks.

Datasets were restricted to measurements made during June 2002–May 2011. The reason for imposing such a restriction is that AMSR-E SST is available from June 2002–October 2011 and we wish to eliminate seasonal bias in our results.

2.3.2 Description of mean equatorward SST gradient

Zonal bands are demonstrated in both mean equatorward SST gradient and SST front probability. Mean equatorward SST gradient is computed as the equatorward gradient of time-averaged, mapped AMSR-E SST. This is in many ways analogous to time-averaged zonal geostrophic velocity, $\bar{u}$, computed from SSH [Buckingham and Cornillon, 2013]. In addition to time-averaging, we apply a spatial high pass filter to the data prior to computing an equatorward gradient. The motivation for this step is that it preserves mesoscale features while removing large spatial trends (e.g., the Pole-to-Equator SST gradient). The high pass filter is different than that used by *Maximenko et al.* [2008] and described by *Buckingham and Cornillon* [2013] but results in a smoother filter response (Figure 2.2). The filtering process can be described as follows. We apply a spatial low pass-filter to the data, subtract this from the original data, and repeat this step a second time. This implements a spatial high pass-filter twice. The low pass filter is a two-dimensional, nearly isotropic (in degrees), minimum 4-term Blackman-Harris window [Harris, 1978] with $N = 51$. It has an effective cutoff of 4.5° , which we define as the wavelength at which the filter has 0.7079 gain (i.e., 3 dB below the maximum gain). We also apply this spatial filtering process to mapped (i.e., merged and interpolated) ADT. While the filtering process described here restricts anomalies to those characterized by wavelengths less than 450 km, we do not believe we are preferentially selecting zonal or meridional waves/eddies owing to the isotropic nature of the filter.

2.3.3 Description of SST front probability

We apply an ocean front detection algorithm [Cayula and Cornillon, 1992] to swath measurements of SST in order to compute SST front probability. The motivation for using the detection algorithm is that it often identifies fronts associated with weak SST gradients. In contrast, fronts determined by thresholding SST gradient at a low threshold can result in spurious detections. The reason for using swath SST in place of mapped SST is that it removes potential for false detections when applying the algorithm to mapped SST. These can occur, for example, when the algorithm encounters a portion of the mapped SST product that was measured at different times of the day—temperatures can be significantly altered relative to neighboring pixels either due to advection of surface temperatures or air-sea dynamics that may have changed over the course of the day. (We attempt to eliminate front detections resulting from or obscured by diurnal warming events by excluding pixels whose estimated diurnal warming amplitude exceeds 0.4°C .) A brief description of the detection algorithm follows. See Cayula and Cornillon [1992] for a detailed description of the algorithm.

The algorithm, known as the Single Image Edge Detector (SIED), locally searches for temperature populations within an image. It consists of three steps: (1) histogram, (2) cohesion and (3) contour-following steps. In the first step of the algorithm, the SIED computes the histogram of temperatures within overlapping 32×32 -pixel tiles. If the algorithm flags a tile as consisting of two populations, the SIED then calculates the *cohesion* of the tile. The meaning of cohesion is best understood using the following two examples. A tile consisting of two temperatures, T_1 and T_2 , in a checkerboard pattern is not cohesive; in contrast, a tile having a equal percentages of temperatures T_1 and T_2 but with left-half equal to T_1 and right-half equal to T_2 is cohesive. In identifying valid SST fronts, we wish to detect the latter of these while rejecting the former. Finally, if the tile passes both histogram and cohesive tests, those pixels having temperatures falling between the two populations are flagged as frontal pixels and a contouring algorithm connects and extends the pixels associated with this front segment. Since the swath measurements are provided by RSS at nominally 0.1° resolution, we binned front detections on a $0.1^{\circ}\times 0.1^{\circ}$ grid and calculated SST front probability as the number of detections divided by the number of clear

pixels⁶ within the data identified within each bin. Defined in this manner, SST front probability is a function of latitude, longitude and averaging period for the given space-time bin.

Two factors must be considered when applying the SIED to swath measurements of AMSR-E SST. The first is that AMSR-E data are distributed in along- and cross-scan coordinates. Because coordinates are not, in general, orthogonal, care must be taken to ensure that front probabilities computed from detections of the SIED are consistent with those obtained using orthogonal coordinates. This has been done. The second factor that must be considered is the effect large-scale temperature gradients have on detector performance. During the course of this study, the SIED was found to detect fronts more readily in the presence of a background temperature gradient. Furthermore, depending on noise characteristics of the underlying dataset, false detections are possible (Buckingham and Cornillon, manuscript in preparation, 2013).

Rather than present a modified version of the original algorithm that attempts to account for false detections resulting from background gradients, we present front probability resulting from the algorithm of *Cayula and Cornillon* [1992] and parenthetically note that modification of the algorithm to account for such concerns results in zonal patterns in SST front probability nearly identical to those displayed in Figure 2.2.

In the next section, we apply a normalization scheme to SST front probability. Our reason for doing so is to enhance weak zonal signals present in front probability but that are small relative to surrounding signals. Normalization is performed by subtracting from each pixel the local mean and dividing by the standard deviation within a $4.1^\circ \times 4.1^\circ$ window. This field is referred to as locally normalized SST front probability.

2.4 Results

In this section, we demonstrate the existence of zonal bands in microwave SST, examine a particular oceanographic feature that is coincident with the observed zonal bands, and estimate the persistence of zonal bands in SST and striations in SSH with averaging period.

⁶We defined clear pixels as those pixels identified by Remote Sensing Systems as “excellent” (i.e., proximity confidence flag = 4) and model diurnal warming amplitude less than 0.4°K .

2.4.1 Demonstration of bands in microwave SST

Figure 2.3a-b displays mean equatorward SST gradient and SST front probability computed from microwave SST measurements obtained during 2002–2011. One observes zonally-elongated structures regularly separated in space in all major ocean basins. Patterns in mean equatorward SST gradient alternate in sign while those in SST front probability are associated primarily with positive, equatorward SST gradients (Figure 2.4). In addition, in 2.3c we display locally normalized SST front probability, which reveals the patterns in greater detail. Such patterns, hereafter referred to as bands, are similar to striations observed by *Maximenko et al.* [2005, 2008] (cf. Figure 2.1).

2.4.2 Distributions of equatorward gradients in the subtropics

The distributions of mesoscale equatorward gradients in the subtropics generally follow a similar trend. In winter months, there is an increase in large gradients and decrease in small gradients. This change is followed by a reduction in the number of large gradients and increase in small gradients in summer months. To illustrate this seasonal trend, we display in Figure 2.5a the distribution of equatorward SST gradients in the South Pacific as a function of time-of-year, where equatorward gradients were estimated from spatially high pass-filtered SST (see section 2.3, Methods). We also illustrate in Figure 2.5b the likelihood of observing each gradient value, independent of time-of-year, for five subtropical ocean basins (see Table 2.1). From Figures 2.5a and 2.5b we conclude that, while the distributions of equatorward SST gradient vary by basin, they have Gaussian-like shape and illustrate that large SST gradients are uncommon. Gradients having value greater than $|\pm 0.0048^{\circ}\text{C km}^{-1}|$ account for only about 11% of observed values. This estimate is refined in Figure 2.5c, where we display the likelihood of observing gradients less than $-0.0048^{\circ}\text{C km}^{-1}$ and greater than $+0.0048^{\circ}\text{C km}^{-1}$ in different ocean basins. We find such gradients occur approximately 10–16% of the time. Furthermore, the probability of observing positive gradients is, in all cases, somewhat larger than that of observing negative gradients. The reason for such asymmetry in the probability of positive and negative gradients greater than $|0.0048^{\circ}\text{C km}^{-1}|$ is unknown at present, but we postulate that the horizontal ad-

vection by mesoscale eddies plays a role (see Discussion).

2.4.3 Characteristics of large SST gradients

High gradient events (HGEs) are defined as equatorward temperature gradients in excess of $|\pm 0.0048^{\circ}\text{C km}^{-1}|$ above the background temperature gradient and whose total connected area exceeds 8 pixels (i.e., 0.5 degrees squared). To illustrate what is meant by the term HGE, we depict in Figure 2.6 a seven-day average of spatially-filtered microwave SST in the South Pacific with contours of identically-filtered SSH overlaid. In addition, we highlight locations at which equatorward SST gradient exceeds $+0.0048^{\circ}\text{C km}^{-1}$ and connected area exceeds 8 pixels (i.e., positive HGEs). Note that an anomaly moving westward at 4 km d^{-1} will propagate a distance approximately equal to the resolution of the mapped SST product over the averaging period (i.e., 7 days).

Several characteristics are evident in this figure. The first is that mesoscale SST and SSH are strongly correlated. The second is that large, positive equatorward SST gradients are located midway between positive and negative SSH anomalies. The third characteristic is somewhat more difficult to see but we believe evident: the magnitudes of SST anomalies are greatest on the western sides of closed contours in SSH. We demonstrate these characteristics in an average sense below. Prior to doing so, however, we examine the spatio-temporal characteristics of HGEs.

Spatial and temporal characteristics of HGEs

HGEs were identified in seven-day averages of SST gradient and the centroid of each feature was retained (i.e., yellow dots in Figure 2.6). In practice, morphological image processing operations [Gonzalez and Woods, 2002] were used to reduce the number of spurious classifications during the identification process and to identify distinct (i.e., unconnected) frontal features. We identified both positive and negative HGEs.

Figure 2.7a displays the positions of HGEs identified from seven-day averages of SST in the South Pacific during 2002–2011. Figure 2.7b displays the signed histogram of these positions,

where the signed histogram is computed as the number of positive HGEs minus the number of negative HGEs falling within $0.5^\circ \times 0.5^\circ$ bins. Finally, Figure 2.7c depicts these events in a Hovmöller plot and contain HGEs located within the region $34^\circ\text{--}36^\circ\text{S}$. For reference, we overlay the positions of anticyclonic eddies occurring in the same region and having lifetime ≥ 16 weeks as determined from a database of tracked eddies [Chelton *et al.*, 2011a].

We note three characteristics from these plots. First, HGEs appear to occur in regularly-spaced intervals, with positive and negative HGEs alternating in dominance (Figure 2.7a). Second, these patterns are coincident with contours of bands obtained from multi-year averages of equatorward SST gradient (Figure 2.7b). Third, HGEs propagate westward with speed visually comparable to mesoscale eddies observed and tracked by Chelton *et al.* [2011a] (Figure 2.7c). Plots similar to Figure 2.7c were generated at multiple latitudes (i.e., $20\text{--}50^\circ$ N/S) and within the five subtropical regions (Table 2.1). The positions of HGEs were binned as a function of longitude and time and resulting histograms were used in conjunction with the Radon transform [Deans, 1993] to objectively estimate the speeds of HGEs. In each case, HGEs translated westward with speed comparable to the average propagation speed of mesoscale eddies observed by Chelton *et al.* [2011a] (Figure 2.8). (An exception occurs in the North Atlantic and is attributed to both low sample number and variation in the propagation of HGEs with longitude.) Mean propagation speeds of mesoscale eddies were estimated by selecting eddies with lifetimes ≥ 16 weeks, estimating displacements from the positions of eddies and smoothing associated speeds using a 3-week averaging window.

SST, SSH and Argo profiles averaged within a coordinate system centered on HGEs

Motivated by qualitative characteristics observed in Figure 2.6 (see above), we sought to determine if such characteristics held for the majority of seven-day averages. To do so, we established an average relationship between SST and SSH in a reference system centered on HGEs. SST gradient, SST anomaly and SSH anomaly were interpolated to an equal-area grid centered on positive HGEs, each having dimensions of 300×300 km in eastward and northward directions. Here, we use the terms SST and SSH anomaly to refer to spatially high pass-filtered SST

and SSH, respectively. In an effort to validate use of the interpolated altimeter product, we also identified and averaged along-track altimeter measurements falling within the vicinity of frontal features in the South Pacific. For purposes of discussion, we present average fields associated with positive HGEs only. Negative HGEs are discussed in the Appendix of *Buckingham* [2013].

Averages of SST gradient, SST anomaly and SSH anomaly obtained in five subtropical basins (Table 2.1) are shown in Figure 2.9. In addition, Figure 2.10 displays along-track measurements of SLA averaged in the coordinate system centered on positive HGEs in the South Pacific. Finally, the positions of extrema in average SST and SSH are summarized in Tables 2.2 and 2.3, respectively.

Several observations can be made from Figures 2.9 and 2.10. First, one notes that average SST gradient contains a peak at the origin and falls off with distance from this location, consistent with a field properly averaged in a coordinate system centered on HGEs. Second, average SST anomaly displays an antisymmetric structure about the front location, with warm anomalies equatorward and cool anomalies poleward of the front. Third, average SSH anomaly depicts a similar geometrical structure, consisting of a positive (anticyclonic) anomaly on the equatorward side and negative (cyclonic) anomaly on the poleward side. However, the antisymmetric structure in SSH differs from that in SST in that the axis of the SSH dipole-like pair is shifted approximately 40–50 km East of the SST front location. Fourth, estimates of the SSH structure near the front obtained from along-track altimeter data compare well with those obtained from merged altimeter data averaged within the same coordinate system. Along-track measurements yield a nearly identical structure, although we note that amplitudes in the along-track average are 20% larger. There is also a slight but consistent equatorward shift of SSH with respect to SST in all basins, a feature that is greatest in the Indian Ocean (Tables 2.2 and 2.3).

The above relationship between mesoscale SST and SSH can be understood by examining the geostrophic flow field associated with these anomalies (not shown): west of the front location, there is a convergence of meridional currents toward the front, advecting warm water from the equatorward side and cool water from the poleward side, while currents to the east advect warm and cool waters parallel to and away from the front axis. This results in little or no SST gradient

east of the front but a large SST gradient at the front location. We hypothesize that advection due to mesoscale eddies in the presence of a background gradient accounts for a portion of the SST gradient, with the remaining portion accounted for by contrasting temperatures of the eddies themselves—i.e., contrasting temperatures associated with neighboring eddies of opposite polarity.

To verify that the zonal offset of SSH with respect to SST is an observable feature in the ocean, we obtained temperature measurements made by Argo profiling floats in the vicinity of the SST front. Selecting profiles taken within ± 3.5 days of the occurrence of the front and subtracting the nearest WOA2009 climatological profile, we produced average snapshots of temperature anomalies at 20-, 200- and 1000-m depths (Figure 2.11). The axis of the vortex pair is centrally-located at 20 m, moves eastward at 200 m and can be found East of this location at greater depths. The structure even appears coherent at 1500-m depth (not shown).

Fraction of zonal bands due to HGEs

We have shown that positive HGEs are likely the result of neighboring mesoscale eddies of opposite polarity. In this case, warm, anticyclonic anomalies are found equatorward and cool, cyclonic anomalies are found poleward of the front in SST. In a similar manner, one can identify negative HGEs, compute mean SST and SSH fields in this coordinate system and demonstrate that negative HGEs are associated with cool, cyclonic anomalies equatorward of the front and warm, anticyclonic anomalies poleward of the front. (It might also be noted that the axis of the dipole-like structure, while present in average SSH, is rotated slightly from North.) Given that patterns in the positions of HGEs observed in Figure 2.7 resemble zonal bands, we are motivated to answer the following question: what fraction of zonal bands is due to HGEs?

To answer this question, we identified both positive and negative HGEs. Setting pixels associated with HGEs (e.g., black regions in Figure 2.6) to zero and averaging in time, we estimate that the fraction of variance explained by HGEs in one-, four- and nine-year averages of equatorward SST gradient is 0.19–0.38 (Table 2.4).

2.5 Persistence with averaging period

Persistence is defined as the decay of standard deviation with averaging period. Its definition is motivated by the work of *Schlax and Chelton* [2008], who find that westward-propagating eddies with random initial position produce a time-averaged zonal geostrophic velocity field, $\bar{u}$, whose standard deviation decays inversely with averaging period⁷. Other dependencies remain (e.g., propagation speeds, amplitudes and scales of eddies) but are represented by expected values and therefore remain constant with respect to averaging period. We generalize this model as follows.

We assume that the standard deviation of the quantity p has the form $\sigma_p = \alpha T^\gamma$. Here, T is the averaging period, α is a quantity that measures the amplitude of p at $T = 0$ (i.e., instantaneous snapshots) and γ is the persistence. A value of $\gamma = 0$ corresponds to a stationary signal; a value of $\gamma = -1$ corresponds to a time-averaged quantity statistically identical to a field of randomly positioned eddies that then propagate westward with constant speed. Values between these limits indicate two possibilities: signals characterized by a small number of large-amplitude events or repeated position. Note that we have defined σ_p , α and γ to be functions of latitude and longitude in this study.

We computed the standard deviation of mean equatorward SST gradient (with respect to a spatial mean) and estimated persistence using $15^\circ \times 15^\circ$ tiles with 50% overlap. We used averaging periods $T = 1/13, 1, 2, 3, 4$ years in an effort to eliminate seasonal bias and regressed this quantity as a function of averaging period using the above functional form. For comparison purposes and since the altimeter record provides a longer record from which to estimate persistence, we perform this regression on zonal geostrophic velocity, $\bar{u}$, estimated from ADT. An example of such analysis applied to ADT is given in *Buckingham and Cornillon* [2013].

Figures 2.12a and 2.12b display the persistence of mean equatorward SST gradient (2002–2011) and $\bar{u}$ (1993–2011) with averaging period, respectively. In the following discussion, we

⁷There is some question as to the value with which the Schlax-Chelton model decays (F. Schlosser, 2013, personal communication). Close examination of Equation 2 of *Schlax and Chelton* [2008] suggests $\sigma_{\bar{u}}$ decays as $T^{-0.5}$ owing to the dependence of $\sigma_{\bar{u}}$ on N^* , the effective number of eddies crossing a meridional section within a time period T .

refer to persistence in mean equatorward SST gradient as γ_{SST} and zonal geostrophic velocity γ_{SSH} .

The similarity in γ_{SST} and γ_{SSH} is remarkable, given that SST is a variable influenced by a multitude of factors and γ_{SST} is estimated from a small amount of data in comparison to SSH. One observes large values of persistence (i.e., more positive) in western boundary current regions, significant values at locations in the Southern Ocean, moderate values near coasts and reduced persistence at midlatitudes. $\gamma_{SST} > \gamma_{SSH}$ immediately west of Africa. Slight differences arise in the subtropics, where values of γ_{SST} are low (as low as -0.5) when compared with γ_{SSH} (closer to -0.4). The most significant characteristic in Figure 2.12 is that persistence exceeds -0.5 at all locations, demonstrating that zonal bands in SST and striations in SSH are more persistent than would be expected from random eddies.

2.6 Discussion

Previous studies have discussed the role of mesoscale eddies in creating zonal patterns in time-averaged currents. *Schlax and Chelton* [2008] demonstrate that westward-propagating eddies with characteristics determined from observed eddies give rise to jet-like structures comparable in magnitude and appearance to striations. *Scott et al.* [2008] have also suggested that preferential paths of mesoscale eddies could explain zonal patterns in a measure of anisotropy (i.e., a ratio of velocity variance). Motivated by these studies, *Buckingham and Cornillon* [2013] estimated that a significant fraction of the variance in time-averaged zonal geostrophic velocity, $\bar{u}$, estimated from ADT in the South Pacific can be attributed to eddies and that zonal patterns created by eddies are nearly coincident with observed patterns.

In the present study, we observe a similar relationship between mesoscale eddies and zonal bands in microwave SST. The preceding results suggest that mesoscale eddies of opposite polarity give rise to large SST gradients (i.e., HGEs) through a combination of (i) contrasting warm and cool temperatures associated with (outcropped) mesoscale eddies and (ii) advection of surface waters by the eddies. The analysis also demonstrates that HGEs propagate westward with speed comparable to the westward propagation of mesoscale eddies. Finally, we demonstrate

that HGEs form band-like patterns coincident with observed bands and that explain a significant fraction (0.19–0.38) of the observed variance. While we cannot say for certain, we suspect that the remaining energy can be attributed to zonal flows and propagating baroclinic waves/eddies with smaller amplitude. Given the high degree of similarity between zonal bands in SST and striations in SSH (cf. Figure 2.1), similarity between SST and SSH in the vicinity of HGEs (cf. Figure 2.9) and correlation between mean SST and mean ADT (cf. Fig. 12 of *Buckingham and Cornillon* [2013]), the results of this study lend support to the view that eddies are (at least) embedded in the mean zonal flows referred to as striations and quasi-zonal jets.

One of the more remarkable results of this study is the observed relationship between SST and SSH in the vicinity of HGEs and its consistency with temperature measurements made from Argo profiling floats. We note that Argo measurements display coherent structures to depths of at least 1000 m and possibly deeper, suggesting that upper ocean anomalies seen by the altimeter have a significant barotropic component. Also of note is the westward offset of SST with respect to SSH. Earlier studies have documented a similar offset [*Bernstein and White*, 1982; *Bennett and White*, 1986; *Roemmich and Gilson*, 2001; *Qiu and Chen*, 2005; *Hausmann and Czaja*, 2012] and have used this information to estimate meridional heat flux due to baroclinic waves and/or eddies. In addition, *Chelton et al.* [2011b] observe westward phase shifts of surface chlorophyll and SSH in lagged-correlation estimates that suggest chlorophyll is advected by mesoscale eddies. Thus, while we cannot identify for certain the cause of this spatial offset, advection of surface waters by mesoscale eddies [*Bernstein and White*, 1982] is thought to play an important role.

In our discussion of Figure 2.5c, we identify observed asymmetry in the probability of observing positive and negative HGEs. One explanation for this asymmetry is that neighboring mesoscale eddies of opposite polarity produce large, temperature gradients owing to the temperatures of the anomalies themselves and that the rotational motion of the eddies in the presence of a background temperature gradient enhances (diminishes) SST gradients associated with positive (negative) HGEs. This is illustrated schematically in Figure 2.13, where (a) positive and (b) negative HGEs are depicted as being created by mesoscale eddies of opposite sign. In the first

case, the associated geostrophic currents enhance the existing SST gradient; in the second case, currents reduce the existing SST gradient.

Despite this asymmetry, we find it notable that the likelihood of observing positive HGEs is nearly equal to that of observing negative HGEs (cf. Figure 2.5b–2.5c). That this is so suggests the net effect of advection by eddies in the presence of a background gradient is small in comparison to temperature gradients resulting from temperatures of the mesoscale eddies, themselves.

Some discretion is necessary in interpreting average fields (cf. Figures 2.9–2.11; see also Fig. 5 of *Roemmich and Gilson* [2001] and Fig. 7 of *Bernstein and White* [1982]). In particular, while average fields depict a symmetric dipole structure in both SST and SSH, it is not the case that all HGEs are created by such ideal structures; nor do we believe that in such cases, that they remain coupled for extended periods of time. We have tracked several HGEs in time for periods exceeding 6 months. However, it remains to be seen whether, in general, HGEs can be ascribed such characteristics. Thus, returning to our original word of caution, we summarize our interpretation of average fields as follows: we interpret Figures 2.9–2.11 as reflective of qualitative relationships believed to exist in instantaneous fields (cf. Figure 2.6) that may shed light on dynamics giving rise to these relationships.

A final notable result of this study is the persistence of patterns in SST and SSH. Values in Figures 2.12a and 2.12b suggest one or more processes is at work in the creation of bands and striations. Explanations for such persistence include that weak zonal jets govern the positions of eddies [*Maximenko et al.*, 2005], deep bathymetry has an influential role on the paths of eddies [*Scott et al.*, 2008] and eddies are generated at preferred locations [*Schlax and Chelton*, 2008]. Freely-evolving geostrophic turbulence [*Rhines*, 1975; *Cho and Polvani*, 1996] and potential vorticity staircases [*Baldwin et al.*, 2007; *Dritschel and McIntyre*, 2008] have been eliminated as likely mechanisms [*Maximenko et al.*, 2008] but several possibilities remain [*Pedlosky*, 1975; *Berloff et al.*, 2009, 2011; *Hristova et al.*, 2008; *Wang et al.*, 2012; *Melnichenko et al.*, 2010; *Afanasyev et al.*, 2012].

Progress in interpreting the persistence of zonal patterns might be made by examining stri-

ations/bands relative to bottom topography. Indeed, the positions of bands in SST appear influenced by bathymetric features in the South Pacific (Fig. 12 of *Buckingham and Cornillon* [2013]) and persistence estimates contain suggestions of topographic influence. Another avenue worth exploring is the extension of the Schlax-Chelton model to include more complex probability distributions for the locations of eddies. There is some suggestion from histograms of eddy tracks that eddies have preferred paths (e.g., Fig. 1 of *Schlax and Chelton* [2008]; Fig. 8 of *Chelton et al.* [2011a]). In addition, histograms of distances between neighboring eddies demonstrate that eddies of opposite polarity are closer to one another, on average, than are eddies of the same polarity [*Buckingham*, 2013], indicating a degree of interaction between eddies. These two observations suggest the probability distribution for the positions of eddies is not uniform.

Acknowledgments We would like to thank D. Chelton and P. Klein for conversations regarding an observed relationship between SST and SSH, N. Maximenko and O. Melnichenko for discussions regarding striations in MDOT, and D. Chelton for his review of an early manuscript. F. Schlosser provided valuable input regarding persistence and C. Gentemann provided guidance on the use of microwave SST. The normalization method was inspired by code written by E. C. Larsen. AMSR-E SST were produced and distributed by RSS and sponsored by the NASA Earth Science MEaSUREs DISCOVER Project. The mapped altimeter product was produced by SSALTO/DUACS and distributed by AVISO/CLS with support from CNES. Along-track measurements were provided by DEOS and obtained from the Radar Altimeter Database System (RADS) database. Argo profiles were made available by the International Argo Program and the national programs that contribute to it and were distributed by USGODAE. We thank N. Maximenko and O. Melnichenko for use of MDOT and D. Chelton and M. Schlax for access to the database of tracked eddies. This research has been made possible through two grants and NNX10AN69H from the National Aeronautics and Space Administration. Salary support for P. Cornillon was provided by the State of Rhode Island and Providence Plantations.

List of References

- Afanasyev, Y. D., S. O’Leary, P. B. Rhines, and E. Lindahl (2012), On the origin of jets in the ocean, *Geophys. Astrophys. Fluid Dyn.*, 106(2), 113–137.
- Baldwin, M. P., P. B. Rhines, H.-P. Huang, and M. E. McIntyre (2007), The jet-stream conundrum, *Science*, 315, 467–468, doi:10.1126/science.1131375.
- Bennett, A. F., and W. B. White (1986), Eddy heat flux in the subtropical North Pacific, *J. Phys. Ocean.*, 16, 728–740.
- Berloff, P., I. Kamenkovich, and J. Pedlosky (2009), A mechanism of formation of multiple zonal jets in the oceans, *J. Fluid Mech.*, 628, 395–425, doi:10.1017/S0022112009006375.
- Berloff, P., S. Karabasov, J. T. Farrar, and I. Kamenkovich (2011), On latency of multiple zonal jets in the oceans, *J. Fluid Mech.*, 686, 534–567, doi:10.1017/jfm.2011.345.
- Bernstein, R. L., and W. B. White (1982), Meridional eddy heat flux in the Kuroshio Extension Current, *J. Phys. Ocean.*, 12, 154–159.
- Buckingham, C. E. (2013), Ubiquitous zonal bands in subtropical oceans observed from space, Ph.D. thesis, Grad. School of Oceano., Univ. of Rhode Island, Kingston, Rhode Island.
- Buckingham, C. E., and P. C. Cornillon (2010), Observed quasi-zonal banded structures in SST frontal probability: why do we see them?, Ocean Sciences 2010 Conference, Portland, Oregon.
- Buckingham, C. E., and P. C. Cornillon (2013), The contribution of eddies to striations in absolute dynamic topography, *J. Geophys. Res.*, 118, 1–14, doi:10.1029/2012JC008231.
- Cayula, J-F. and P. C. Cornillon (1992): Edge detection algorithm for SST images, *J. Atmos. Oc. Tech.*, 9, 67–80.

- Centurioni, L., J. Ohlmann, and P. Niiler (2008), Permanent meanders in the California Current System, *J. Phys. Ocean.*, 38, 1690–1710, doi:10.1175/2008JPO3746.1.
- Chelton, D. B., and M. G. Schlax (1996), Global observations of oceanic Rossby waves, *Science*, 272, 234–238.
- Chelton, D. B., J. C. Ries, B. J. Haines, L.-L. Fu, and P. S. Callahan (2001), Satellite altimetry, in *Satellite Altimetry and Earth Sciences: A Handbook of Techniques and Applications*, edited by L.-L. Fu and A. Cazenave, pp. 1–131, Academic, San Diego, Calif.
- Chelton, D. B., and F. J. Wentz (2005), Global observations of nonlinear mesoscale eddies, *Bull. Amer. Meteor. Soc.*, 86, 1097–1115, doi:10.1175/BAMS-86-8-1097.
- Chelton, D. B., M. G. Schlax, R. M. Samelson, and R. A. de Szoeke (2007), Global observations of large oceanic eddies, *Geophys. Res. Lett.*, 34, L15606, doi:10.1029/2007GL030812.
- Chelton, D. B., M. G. Schlax, and R. M. Samelson (2011a), Global observations of nonlinear mesoscale eddies, *Prog. Oceanog.*, 91, 167–216, doi:10.1016/j.pocean.2011.01.002.
- Chelton, D. B., P. Gaube, M. G. Schlax, J. J. Early, and R. M. Samelson (2011b), The influence of nonlinear mesoscale eddies on near-surface oceanic chlorophyll, *Science*, 334(6054), 328–332.
- Cho, J. Y.-K., and L. M. Polvani (1996), The emergence of jets and vortices in freely evolving, shallow-water turbulence on a sphere, *Phys. Fluids*, 8, 1531–1552.
- Cravatte, S., and W. Kessler, and F. Marin (2012), Intermediate zonal jets in the tropical Pacific Ocean observed by Argo floats, *J. Phys. Oceanogr.*, doi:10.1175/JPO-D-11-0206.1.
- Deans, S. R., (1993), *The Radon transform and some of its applications*, 2nd ed., Krieger, Malabar, Florida.
- Di Lorenzo, E., A. Belmadani, H. Luo, O. Melnichenko, N. Schneider, and N. Maximenko (2012), The emergence of striations in the North Pacific circulation, Ocean Sciences 2012 Conference, Salt Lake City, Utah.

- Divakaran, P., and G. B. Brassington (2011), Arterial ocean circulation of the southeast Indian Ocean, *Geophys. Res. Lett.*, **38**, L01802, doi:10.1029/2010GL045574.
- Dritschel, D. G., and M. E. Kessler (2008), Multiple jets as PV staircases: The Phillips effect and the resilience of eddy-transport barriers, *J. Atmos. Sci.*, doi:10.1175/2007JAS2227.1.
- Gonzalez, R. C., and R. E. Woods (2002), *Digital Image Processing*, 2nd ed., 793 pp., Prentice Hall, Englewood Cliffs, N. J.
- Harris, F. J. (1978), On the use of windows for harmonic analysis with the discrete Fourier transform, *Proceedings of the IEEE*, **66**, 51–84.
- Hausmann, U., and A. Czaja (2012), The observed signature of mesoscale eddies in sea surface temperature and the associated heat transport, *Deep-Sea Res. I*, **70**, 60–72, doi:10.1016/j.dsr.2012.08.005.
- Hogg, N. G., and W. Brechner Owens (1999), Direct measurement of the deep circulation within the Brazil Basin, *Deep-Sea Res. II*, **46**, 335–353.
- Hristova, H. G., J. Pedlosky, and M. A. Spall (2008), Radiating instability of a meridional boundary current, *J. Phys. Oceanogr.*, **38**, 2294–2307, doi:10.1175/2008JPO3853.1.
- Huang, H.-P., A. Kaplan, E. N. Curchitser, and N. A. Maximenko (2007), The degree of anisotropy for mid-ocean currents from satellite observations and an eddy-permitting model simulation, *J. Geophys. Res.*, **36**, C09005, doi:10.1029/2007JC004105.
- Ivanov, L., C. Collins, and T. Margonina (2009), System of quasi-zonal jets off California revealed from satellite altimetry, *Geophys. Res. Lett.*, **36**, L03609, doi:10.1029/2008GL036327.
- Ivanov, L., C. Collins, T. Margonina, and V. Eremeev (2011), Nonlinear Rossby waves off California, *Geophys. Res. Lett.*, **37**, L13602, doi:10.1029/2010GL043708.
- Kamenkovich, I., P. Berloff, and J. Pedlosky (2009), Role of eddy forcing in the dynamics of multiple zonal jets in a model of the North Pacific, *J. Phys. Ocean.*, **39**, 1361–1379, doi:10.1175/2008JPO4096.1.

- Martin, S. (2004), *An Introduction to Ocean Remote Sensing*, 426 pp., Cambridge University Press, Cambridge.
- Maximenko, N., B. Bang, and H. Sasaki (2005), Observational evidence of alternating zonal jets in the World Ocean, *Geophys. Res. Lett.*, *32*, L12607, doi:10.1029/2005GL022728.
- Maximenko, N., O. V. Melnichenko, P. P. Niiler, and H. Sasaki (2008), Stationary mesoscale jet-like features in the ocean, *Geophys. Res. Lett.*, *35*, L08603, doi:10.1029/2008GL033267.
- Melnichenko, O. V., N. A. Maximenko, N. Schneider, and H. Sasaki (2010), Quasi-stationary striations in basin-scale oceanic circulation: vorticity balance from observations and eddy-resolving model, *Ocean Dyn.*, *60*, 653–666.
- Pedlosky, J. (1975), On secondary baroclinic instability and the meridional scale of motion in the ocean, *J. Phys. Ocean.*, *5*, 603–607.
- Qiu, B., and S. Chen (2005), Eddy-induced heat transport in the subtropical North Pacific from Argo, TMI, and altimetry measurements, *J. Phys. Ocean.*, *35*, 458–473.
- Rhines, P. B. (1975), Waves and turbulence on a beta-plane, *J. Fluid Mech.*, *69*, 417–443.
- Rhines, P. B. (1994), Jets, *Chaos*, *4*, 313–339.
- Rio, M. H., S. Guinehut, and G. Larnicol (2011), New CNES-CLS09 global mean dynamic topography computed from the combination of GRACE data, altimetry, and in situ measurements, *J. Geophys. Res.*, C07018, doi:10.1029/2010JC006505.
- Robinson, A. R., (Ed.) , 1983: *Eddies in Marine Science*. Springer, New York, 609 pp.
- Roemmich, D., and J. Gilson (2001), Eddy transport of heat and thermocline waters in the North Pacific: a key to interannual/decadal climate variability?, *J. Phys. Ocean.*, *31*, 675–687.
- Schlag, M. G., and D. B. Chelton (2008), The influence of mesoscale eddies on the detection of quasi-zonal jets in the ocean, *Geophys. Res. Lett.*, *35*, L24602, doi:10.1029/2008GL035998.

- Scott, R. B., B. K. Arbic, C. L. Holland, A. Sen, and B. Qiu (2008), Zonal versus meridional velocity variance in satellite observations and realistic and idealized ocean circulation models, *Ocean Modell.*, 23, 102–112, doi:10.1016/j.ocemod.2008.04.009.
- van Seville, E., I. Kamenkovich, and J. K. Willis (2011), Quasi-zonal jets in 3D Argo data of the northeast Atlantic, *Geophys. Res. Lett.*, 38, L02606, doi:10.1029/2010GL046267.
- Wang, J., M. A. Spall, G. R. Flierl, and P. Malanotte-Rizzoli (2012), A new mechanism for the generation of quasi-zonal jets in the ocean, *Geophys. Res. Lett.*, 39, L10601, doi: 10.1029/2012GL051861.

Figure 2.1. Zonal geostrophic velocity estimated from MDOT (1992–2002; color) with locations of normalized SST front probability overlaid (2002–2011; black). Normalization was performed by subtracting from each pixel the local mean and dividing by the standard deviation within this same window. We used a $4.1^\circ \times 4.1^\circ$ window centered on each pixel and a threshold of 1.2 when displaying normalized front probability.

Figure 2.2. Effective magnitude-squared response of the high pass-filtering methods used in this study. The filter is nearly isotropic in latitude and longitude in units of deg. Wavelengths in units of km assume $1 \text{ deg} = 111.2 \text{ km}$ and, thus, correspond to meridional distances. The shaded region highlights wavelengths of 300–500 km. For convenience, this filter response is compared with that used in an earlier study [Buckingham and Cornillon, 2013].

Figure 2.3. (a) Mean equatorward SST gradient, (b) SST front probability and (c) (locally) normalized SST front probability computed from microwave SST (June 2002–May 2011). Data in (a) were spatially high pass-filtered prior to estimating meridional gradients; data in (b) were obtained by applying the SIED to swath SST and computing SST front probability as the number of front detections divided by the number of clear pixels; normalization in (c) was performed as described in Figure 2.1.

Figure 2.4. Histogram of mean equatorward SST gradient (June 2002–May 2011) at locations where SST front probability exceeds 0.12 (cf. Figure 2.3b). The bulk of the histogram is found right of zero, as demonstrated by the skewness (*skew*) of the distribution.

Figure 2.5. (a) Probability of equatorward SST gradient as a function of time-of-year in the subtropical South Pacific (2002–2011); (b) probability of equatorward SST gradient for each subtropical basin (Table 2.1) with the probability of large positive and negative gradients in the South Pacific highlighted; (c) probability of equatorward gradient falling below $-0.0048^{\circ}\text{C km}^{-1}$ (α_1) and exceeding $+0.0048^{\circ}\text{C km}^{-1}$ (α_2). Equatorward gradients were computed from spatially high-filtered SST and do not contain large-scale background gradients.

Figure 2.6. Spatially-filtered SST (color; seven-day mean centered on June 14, 2006) with spatially-filtered ADT (contours; valid June 14, 2006) in the South Pacific. Positive HGEs are highlighted in black and the centroids of identified frontal features are shown in yellow. Contours range between ± 4 cm.

Figure 2.7. High gradient events (HGEs) in the South Pacific, 2002–2011: (a) positions of positive (red) and negative (blue) HGEs; (b) signed histogram of HGEs (color) with mean equatorward SST gradient (2002–2011) at $0.0002^{\circ}\text{C km}^{-1}$ overlaid (contours); signed histogram computed as the number of positive HGEs minus number of negative HGEs found within $0.5^{\circ}\times 0.5^{\circ}$ bins; (c) time-longitude diagram of HGEs at 35°S ; solid black line corresponds to an object with constant westward translation speed of 1.68 km d^{-1} and lifetime of 2 years; gray lines are positions of anticyclonic eddies having lifetimes ≥ 16 wks and obtained from the *Chelton et al.* [2011a] eddy database.

Figure 2.8. Westward propagation speeds of HGEs and mesoscale eddies in the subtropics. HGE speeds were estimated from the Radon transform of time-longitude histograms of HGEs. Speeds in the North Atlantic are not shown. Eddy speeds correspond to the average speeds of mesoscale eddies observed and tracked by *Chelton et al.* [2011a] and having lifetimes ≥ 16 wks.

Figure 2.9. Mean northward SST gradient, SST anomaly and SSH anomaly obtained when averaging within a coordinate system centered on positive HGEs (June 2002–May 2011). Subtropical basins include (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Oceans (Table 2.1). Black dots denote extrema and solid vertical lines highlight average x -positions of the eddy-like pair. Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.

Figure 2.10. As in Figure 2.10 but obtained when averaging along-track altimeter data (SLA) in the South Pacific (June 2002–May 2011). Altimeter measurements are those from the Jason-1 spacecraft and were not spatially filtered prior to averaging. This figure should be compared with Figure 2.9b, third column.

Figure 2.11. Mean temperature anomaly from Argo profiles (June 2002–May 2011) at depths of 20, 200 and 1000 m when averaged within a coordinate system centered on positive HGEs. To the left, we illustrate positions of profiles at 1000-m depth. Basins include the (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Ocean (Table 2.1). Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.

Figure 2.12. Persistence of (a) mean equatorward SST gradient (June 2002–May 2011) and (b) mean zonal geostrophic velocity, $\bar{u}$, estimated from SSH (January 1993–December 2011). Persistence is estimated from a regression of standard deviation over averaging periods $T = 1/13, 1, 2, 3, 4$ years. See text for a definition of persistence. Bathymetric features shallower than 2500 m are highlighted in black.

Figure 2.13. SST gradients enhanced and reduced by mesoscale eddies in the Southern Hemisphere. We ignore spatial offsets of SST relative to SSH, assume a background Pole-to-Equator temperature gradient and assume the front is dominated by the northward gradient of temperature anomalies themselves. (a) Geostrophic currents associated with positive HGEs: an anticyclonic (a cyclonic) anomaly located North (South) of the front will advect warmer (cooler) waters toward the front-axis. The net effect of advection is to enhance the existing SST gradient. (b) Geostrophic currents associated with negative HGEs: a cyclonic (an anticyclonic) anomaly located North (South) of the front will advect warmer (cooler) water toward the front axis. The net effect of advection is to reduce the existing SST gradient.

Table 2.1. Definitions of subtropical regions examined in this study.

Region	Latitude	Longitude
North Pacific (NP)	20°–45°N	180°–230°E
South Pacific (SP)	20°–50°S	200°–280°E
North Atlantic (NA)	30°–55°N	325°–350°E
South Atlantic (SA)	20°–35°S	35°W–10°E
Indian (I)	15°–35°S	55°–110°E

Table 2.2. The magnitude and position of extrema in average SST anomaly. Units of magnitude are °C. Units of position are km and are given in terms of distances from the center of the front (positive, eastward and equatorward). Here, in contrast with plots in Figure 2.9, we define the center of the front as the extremum in average equatorward SST gradient.

Region	min	max	$(x, y)_{min}$	$(x, y)_{max}$
NP	0.40	0.42	(0, -72)	(0, 81)
SP	0.39	0.39	(0, -71)	(0, 81)
NA	0.41	0.38	(0, -71)	(0, 81)
SA	0.43	0.37	(0, -82)	(0, 72)
I	0.44	0.41	(0, -82)	(0, 72)

Table 2.3. As in Table 2.2 but for average SSH anomaly. Units of magnitude are cm.

Region	min	max	$(x, y)_{min}$	$(x, y)_{max}$
NP	3.1	3.3	(51, -71)	(41, 102)
SP	3.7	4.0	(41, -72)	(40, 102)
NA	5.1	5.1	(30, -61)	(30, 92)
SA	4.5	4.0	(61, -82)	(51, 122)
I	5.4	4.6	(62, -91)	(61, 122)

Table 2.4. Fraction of variance in mean equatorward SST gradient due to HGEs. Averaging periods are one, four and nine years starting in June of the displayed year. In the case of multi-year averages, averaging periods commence in June of the first year and end in June of the second year. See text for a definition of HGEs.

Year(s)	NP	SP	NA	SA	I
2002	0.29	0.31	0.21	0.22	0.19
2003	0.28	0.36	0.24	0.32	0.19
2004	0.32	0.34	0.24	0.38	0.20
2005	0.31	0.32	0.27	0.27	0.20
2006	0.33	0.32	0.21	0.34	0.23
2007	0.28	0.33	0.27	0.29	0.23
2008	0.28	0.31	0.21	0.35	0.20
2009	0.29	0.29	0.20	0.25	0.23
2010	0.26	0.28	0.21	0.35	0.27
2002–2005	0.29	0.35	0.27	0.26	0.21
2006–2009	0.28	0.31	0.21	0.35	0.23
2002–2010	0.27	0.33	0.25	0.32	0.23

APPENDIX A

Statistics of neighboring mesoscale eddies obtained from a database of tracked eddies

One of the concepts raised in the Discussion in Chapter 2 is that mesoscale eddies may interact. Here, we present evidence gleaned from the database of tracked eddies used in Chapter 1 that suggests mesoscale eddies of opposite polarity are closer to each other, on average, than are eddies of the same polarity. This suggests some level of interaction between eddies.

Figure A.1 displays the two-dimensional histograms of the positions of eddies closest to anticyclones. Here, we examine eddies with lifetimes ≥ 4 weeks and located within the South Pacific (20° – 50° S, 200° – 280° E) during 1992–2011. Additionally, we have defined the distance between eddies as the distance between eddy centers. Unique eddies that have been identified and tracked are counted only once per bin. In Figure A.2, we display the histograms of the positions of eddies closest to cyclones.

Several characteristics are evident in these figures. First, opposite polarity eddies are observed within closest proximity to one another more often than are eddies of the same polarity. This arrangement occurs 6–9 times more often than does eddies of opposite polarity (the exact ratio is 7.52, although ratios of maxima in Figures A.1 and A.2 have values of 6.3–8.8). Second, cyclones are observed east of anticyclones more often than west. Additionally, anticyclones are observed west of cyclones more often than east. These trends are consistent and suggest a zonal orientation¹ of opposite-polarity eddies. Third, mean and most probable distances between eddy centers is approximately 200 km for opposite-polarity eddies and 230–250 km for same-polarity eddies. (There is also the suggestion of an east-west orientation between opposite-polarity eddies but its significance is unknown due to low sample number.)

Given results of Chapters 1 and 2 (i.e., that mesoscale eddies are intimately related to striations and bands), the distance between eddies of opposite polarity may have relevance to zonal bandsthe separation between bands is approximately 200 km and the distance between eddies of

¹By orientation of two eddies, we mean the orientation of a line drawn between the centers of the eddies.

opposite polarity is nearly the same, leading us to believe that eddies of opposite polarity and that have meridional orientation contribute most to the observed patterns.

Figure A.1. Histograms of the positions of neighboring eddies in the South Pacific when examining (a) cyclones closest to anticyclones and (b) anticyclones closest to anticyclones. Distances are defined as distances from eddy centers.

Figure A.2. As in Figure A.1 but for (a) cyclones closest to cyclones and (b) anticyclones closest to cyclones.

APPENDIX B

SST, SSH and Argo data averaged within a coordinate system centered on negative HGEs

We document the relationship of microwave SST and SSH associated with negative high gradient events (HGEs). Negative HGEs are defined as equatorward temperature gradients less than $-0.0048^{\circ}\text{C km}^{-1}$ above the background temperature gradient and whose total connected area exceeds 8 pixels (i.e., 0.5 degrees squared). See Chapter 2.

Averages of SST gradient, SST anomaly and SSH anomaly obtained in five subtropical basins (Table 2.1) are shown in Figure B.1. In addition, we display in Figure B.2 the result of averaging temperatures from Argo profiles in this same coordinate system. While results are comparable to those found for positive HGEs, slight differences exist. In particular, the axis of the dipole pair in SSH appears to be rotated 10° counter-clockwise (clockwise) from north in the Northern (Southern) Hemisphere. While it is difficult to interpret such differences, simple advective models of mesoscale eddies propagating at angles relative to a background temperature gradient suggest that propagation directions of mesoscale eddies plays a role (F. Schlosser, 2013; personal communication).

Figure B.1. Mean northward SST gradient, SST anomaly and SSH anomaly obtained when averaging within a coordinate system centered on negative HGEs (June 2002–May 2011). Sub-tropical basins include (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Oceans (Table 2.1). Black dots denote extrema and solid vertical lines highlight average x -positions of the eddy-like pair. Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.

Figure B.2. Mean temperature anomaly from Argo profiles (June 2002–May 2011) at depths of 20, 200 and 1000 m when averaged within a coordinate system centered on negative HGEs. To the left, we illustrate positions of profiles at 1000-m depth. Basins include the (a) North Pacific, (b) South Pacific, (c) North Atlantic, (d) South Atlantic and (e) Indian Ocean (Table 2.1). Abscissae (x) and ordinates (y) are distances in eastward and northward directions, respectively.

BIBLIOGRAPHY

- Afanasyev, Y., S. O'Leary, P. Rhines, and E. Lindahl (2012), On the origin of jets in the ocean, *Geophysical & Astrophysical Fluid Dynamics*, 106(2), 113–137.
- Baldwin, M. P., P. Rhines, H. Huang, and M. McIntyre (2007), Atmospheres: The jet-stream conundrum, *Science*, 315(5811), 467–468.
- Bendat, J. S., and A. G. Piersol (2011), *Random Data: Analysis and Measurement Procedures*, 3rd ed., John Wiley and Sons, New York.
- Bennett, A. F., and W. B. White (1986), Eddy heat flux in the subtropical North Pacific, *Journal of Physical Oceanography*, 16(4), 728–740.
- Berloff, P., I. Kamenkovich, and J. Pedlosky (2009), A mechanism of formation of multiple zonal jets in the oceans, *Journal of Fluid Mechanics*, 628, 395.
- Berloff, P., S. Karabasov, J. T. Farrar, and I. Kamenkovich (2011), On latency of multiple zonal jets in the oceans, *Journal of Fluid Mechanics*, 686, 534–567.
- Bernstein, R., and W. WB (1982), Meridional eddy heat flux in the Kuroshio Extension Current, *Journal of Physical Oceanography*, 12, 154–159.
- Beron-Vera, F. J., M. J. Olascoaga, and G. J. Goni (2010), Surface ocean mixing inferred from different multisatellite altimetry measurements, *Journal of Physical Oceanography*, 40(11), 2466–2480.
- Buckingham, C., and P. Cornillon (2010), Observed quasi-zonal banded structures in SST frontal probability: why do we see them?, in *Ocean Sciences 2010 Conference*, AGU/TOS/ASLO, Portland, Oregon.
- Buckingham, C., and P. Cornillon (2013), The contribution of eddies to striations in absolute dynamic topography, *Journal of Geophysical Research: Oceans*, 118, 1–14.

- Cayula, J.-F., and P. Cornillon (1992), Edge detection algorithm for SST images, *Journal of Atmospheric and Oceanic Technology*, 9(1), 67–80.
- Centurioni, L., J. Ohlmann, and P. Niiler (2008), Permanent meanders in the California Current System, *Journal of Physical Oceanography*, 38(8), 1690–1710.
- Chelton, D. B., and M. G. Schlax (1996), Global observations of oceanic Rossby waves, *Science*, 272(5259), 234–238.
- Chelton, D. B., and F. J. Wentz (2005), Global microwave satellite observations of sea surface temperature for numerical weather prediction and climate research, *Bulletin of the American Meteorological Society*, 86(8), 1097–1115.
- Chelton, D. B., and S.-P. Xie (2010), Coupled ocean-atmosphere interaction at oceanic mesoscales, *Oceanography*, 23(4), 52–69.
- Chelton, D. B., J. C. Ries, B. J. Haines, L.-L. Fu, and P. S. Callahan (2001), Satellite altimetry, *International Geophysics*, pp. 1–131.
- Chelton, D. B., M. G. Schlax, R. M. Samelson, and R. A. de Szoeke (2007), Global observations of large oceanic eddies, *Geophysical Research Letters*, 34(15).
- Chelton, D. B., P. Gaube, M. G. Schlax, J. J. Early, and R. M. Samelson (2011a), The influence of nonlinear mesoscale eddies on near-surface oceanic chlorophyll, *Science*, 334(6054), 328–332.
- Chelton, D. B., M. G. Schlax, and R. M. Samelson (2011b), Global observations of nonlinear mesoscale eddies, *Progress in Oceanography*, 91(2), 167–216.
- Cheney, R., L. Miller, R. Agreen, N. Doyle, and J. Lillibridge (1994), TOPEX/POSEIDON: the 2-cm solution, *Journal of Geophysical Research: Oceans*, 99(C12), 24,555–24,563.
- Cho, J. Y., and L. M. Polvani (1996), The emergence of jets and vortices in freely evolving, shallow-water turbulence on a sphere, *Physics of Fluids*, 8, 1531–1552.

- Cravatte, S., W. S. Kessler, and F. Marin (2012), Intermediate zonal jets in the tropical Pacific Ocean observed by Argo floats, *Journal of Physical Oceanography*, 42(9), 1475–1485.
- Deans, S. R. (1993), *The Radon Transform and Some of its Applications*, 2nd ed., Krieger, Malabar, Florida.
- Di Lorenzo, E., A. Belmadani, H. Luo, O. Melnichenko, N. Schneider, and N. Maximenko (2012), The emergence of striations in the North Pacific circulation, in *Ocean Sciences 2012 Conference*, AGU/TOS/ASLO, Salt Lake City, Utah.
- Divakaran, P. (2011), Mesoscale circulation and variability of the Indian Ocean, Ph.D. thesis, School of Earth Sc., Univ. of Melbourne, Melbourne, Australia.
- Divakaran, P., and G. B. Brassington (2011), Arterial ocean circulation of the southeast Indian Ocean, *Geophysical Research Letters*, 38(1).
- Dritschel, D., and M. McIntyre (2008), Multiple jets as PV staircases: the Phillips effect and the resilience of eddy-transport barriers, *Journal of the Atmospheric Sciences*, 65(3), 855–874.
- Ducet, N., P.-Y. Le Traon, and G. Reverdin (2000), Global high-resolution mapping of ocean circulation from TOPEX/Poseidon and ERS-1 and -2, *Journal of Geophysical Research: Oceans*, 105(C8), 477–498.
- Firing, E. (1987), Deep zonal currents in the central equatorial Pacific, *Journal of Marine Research*, 45(4), 791–812.
- Gonzalez, R. C., and E. Richard (2002), *Digital Image Processing*, 2nd ed., Prentice Hall, Englewood Cliffs, New York.
- Gouriou, Y., T. Delcroix, and G. Eldin (2006), Upper and intermediate circulation in the western equatorial Pacific Ocean in October 1999 and April 2000, *Geophysical Research Letters*, 33(10), 1–4.
- Harris, F. J. (1978), On the use of windows for harmonic analysis with the discrete Fourier transform, *Proceedings of the IEEE*, 66(1), 51–83.

- Hausmann, U., and A. Czaja (2012), The observed signature of mesoscale eddies in sea surface temperature and the associated heat transport, *Deep Sea Research I*, 70, 60–72.
- Hogg, N. G., and W. B. Owens (1999), Direct measurement of the deep circulation within the Brazil Basin, *Deep Sea Research II*, 46(1), 335–354.
- Holton, J. R. (1979), *An Introduction to Dynamic Meteorology*, Elsevier Academic Press, San Diego, California.
- Hristova, H. G., J. Pedlosky, and M. A. Spall (2008), Radiating instability of a meridional boundary current, *Journal of Physical Oceanography*, 38(10), 2294–2307.
- Huang, H.-P., A. Kaplan, E. N. Curchitser, and N. A. Maximenko (2007), The degree of anisotropy for mid-ocean currents from satellite observations and an eddy-permitting model simulation, *Journal of Geophysical Research: Oceans*, 112(C9), 1–12.
- Ivanov, L., C. Collins, and T. Margolina (2009), System of quasi-zonal jets off California revealed from satellite altimetry, *Geophysical Research Letters*, 36(3), 1–6.
- Ivanov, L., C. Collins, T. Margolina, and V. Eremeev (2010), Nonlinear Rossby waves off California, *Geophysical Research Letters*, 37(13), 1–6.
- Kamenkovich, I., P. Berloff, and J. Pedlosky (2009), Role of eddy forcing in the dynamics of multiple zonal jets in a model of the North Atlantic, *Journal of Physical Oceanography*, 39(6), 1361–1379.
- Le Traon, P., Y. Faugère, F. Hernandez, J. Dorandeu, F. Mertz, and M. Ablain (2003), Can we merge GEOSAT follow-on with TOPEX/Poseidon and ERS-2 for an improved description of the ocean circulation?, *Journal of Atmospheric and Oceanic Technology*, 20(6), 889–895.
- Maltrud, M., and G. Vallis (1991), Energy spectra and coherent structures in forced two-dimensional and beta-plane turbulence, *Journal of Fluid Mechanics*, 228, 321–342.
- Martin, S. (2004), *An Introduction to Ocean Remote Sensing*, Cambridge University Press, Cambridge.

- Maximenko, N. A., B. Bang, and H. Sasaki (2005), Observational evidence of alternating zonal jets in the world ocean, *Geophysical research letters*, 32(12).
- Maximenko, N. A., O. V. Melnichenko, P. P. Niiler, and H. Sasaki (2008), Stationary mesoscale jet-like features in the ocean, *Geophysical Research Letters*, 35(8).
- Maximenko, N. A., O. V. Melnichenko, S.-P. Xie, and N. Schneider (2010), Atmospheric signatures of oceanic striations, in *Altimetry for Oceans and Hydrology - OSTST Meeting*, Lisbon, Portugal.
- Melnichenko, O. V., N. A. Maximenko, N. Schneider, and H. Sasaki (2010), Quasi-stationary striations in basin-scale oceanic circulation: vorticity balance from observations and eddy-resolving model, *Ocean Dynamics*, 60(3), 653–666.
- Nakano, H., and H. Hasumi (2005), A series of zonal jets embedded in the broad zonal flows in the Pacific obtained in eddy-permitting ocean general circulation models, *Journal of Physical Oceanography*, 35(4), 474–488.
- Nowlin Jr, W., and M. Clifford (1982), The kinematic and thermohaline zonation of the Antarctic Circumpolar Current at Drake Passage, *J. Mar. Res.*, 40, 481–507.
- Orsi, A. H., T. Whitworth, and W. D. Nowlin (1995), On the meridional extent and fronts of the Antarctic Circumpolar Current, *Deep Sea Research I*, 42(5), 641–673.
- Panetta, R. L. (1993), Zonal jets in wide baroclinically unstable regions: persistence and scale selection, *Journal of the Atmospheric Sciences*, 50(14), 2073–2106.
- Park, K., P. Cornillon, and D. L. Codiga (2006), Modification of surface winds near ocean fronts: effects of Gulf Stream rings on scatterometer (QuikSCAT, NSCAT) wind observations, *Journal of Geophysical Research: Oceans*, 111(C3).
- Pascual, A., Y. Faugère, G. Larnicol, and P.-Y. Le Traon (2006), Improved description of the ocean mesoscale variability by combining four satellite altimeters, *Geophysical Research Letters*, 33(2).

- Pedlosky, J. (1975), On secondary baroclinic instability and the meridional scale of motion in the ocean, *Journal of Physical Oceanography*, 5, 603–607.
- Qiu, B., and S. Chen (2005), Eddy-induced heat transport in the subtropical North Pacific from Argo, TMI, and altimetry measurements, *Journal of Physical Oceanography*, 35(4), 458–473.
- Rhines, P. B. (1975), Waves and turbulence on a beta-plane, *J. Fluid Mech*, 69(3), 417–443.
- Rhines, P. B. (1994), Jets, *Chaos: An Interdisciplinary Journal of Nonlinear Science*, 4(2), 313–339.
- Rio, M., S. Guinehut, and G. Larnicol (2011), New CNES-CLS09 global mean dynamic topography computed from the combination of GRACE data, altimetry, and in situ measurements, *Journal of Geophysical Research: Oceans*, 116(C7).
- Robinson, A. (1983), *Eddies in Marine Science*, Springer, New York.
- Roemmich, D., and J. Gilson (2001), Eddy transport of heat and thermocline waters in the North Pacific: A key to interannual/decadal climate variability?, *Journal of Physical Oceanography*, 31(3), 675–687.
- Schlax, M. G., and D. B. Chelton (2008), The influence of mesoscale eddies on the detection of quasi-zonal jets in the ocean, *Geophysical Research Letters*, 35(24).
- Scott, R. B., B. K. Arbic, C. L. Holland, A. Sen, and B. Qiu (2008), Zonal versus meridional velocity variance in satellite observations and realistic and idealized ocean circulation models, *Ocean Modelling*, 23(3), 102–112.
- Small, R., S. de Szoëke, S. Xie, L. O’Neill, H. Seo, Q. Song, P. Cornillon, M. Spall, and S. Minobe (2008), Air–sea interaction over ocean fronts and eddies, *Dynamics of Atmospheres and Oceans*, 45(3), 274–319.
- Sokolov, S., and S. R. Rintoul (2007), Multiple jets of the Antarctic Circumpolar Current south of Australia, *Journal of Physical Oceanography*, 37(5), 1394–1412.

- Treguier, A., and R. Panetta (1994), Multiple zonal jets in a quasigeostrophic model of the Antarctic Circumpolar Current, *Journal of Physical Oceanography*, 24(11), 2263–2277.
- van Seville, E., I. Kamenkovich, and J. K. Willis (2011), Quasi-zonal jets in 3-D Argo data of the northeast Atlantic, *Geophysical Research Letters*, 38(2).
- Wang, J., M. A. Spall, G. R. Flierl, and P. Malanotte-Rizzoli (2012), A new mechanism for the generation of quasi-zonal jets in the ocean, *Geophysical Research Letters*, 39(10).
- Williams, G. P. (1978), Planetary circulations: 1. Barotropic representation of Jovian and terrestrial turbulence, *Journal of the Atmospheric Sciences*, 35(8), 1399–1426.
- Wunsch, C., and D. Stammer (1998), Satellite altimetry, the marine geoid, and the oceanic general circulation, *Annual Review of Earth and Planetary Sciences*, 26(1), 219–253.