The influence of mesoscale ocean eddies on near-surface ocean temperature, surface stress and phytoplankton communities is investigated by collocating numerous satellite measurements along with vertical profiles of oceanic temperature and salinity to the interiors of eddies identified and tracked in altimetric sea surface height maps.

The surface currents associated with mesoscale ocean eddies impart a curl of the surface stress from the relative motion between surface air and water. This stress curl has a polarity opposite that of the vorticity of the eddy, thus attenuating the eddies by generating Ekman upwelling in the cores of anticyclones and downwelling in the cores of cyclones. Ekman pumping also arises from eddy-induced spatial variability of the sea surface temperature (SST) field that generates a wind stress curl in regions of crosswind SST gradients through a response of surface winds to SST-induced surface heating variations. SST-induced Ekman pumping is shown to be secondary to surface current-induced pumping in most regions of the World Ocean. Eddy-induced Ekman pumping resulting from the combination of surface current effects and air-sea interaction represents an order 1 perturbation of the background, basin-scale Ekman pumping velocities from the large-scale wind fields.

In western boundary currents and equatorward-flowing eastern boundary currents,
cyclonic eddies preferentially entrain water from the coastal side of the boundary current, which primes the interiors of cyclones to have phytoplankton concentrations that are elevated relative to the background. In contrast, anticyclones formed in these regions contain locally depressed phytoplankton concentrations from the offshore waters. While eddy pumping from vertical displacements of isopycnals during eddy formation can affect the biology in the interiors of cyclones during the transient stage of their development, this ecosystem response cannot be sustained because of the persistent eddy-induced Ekman downwelling throughout the rest of the eddy lifetimes. Likewise, the persistent eddy-induced Ekman upwelling in anticyclones is of little benefit because of their low phytoplankton content at the time of formation. A definitive response to eddy pumping is therefore difficult to detect from satellite observations alone.

Eddies formed in regions where anticyclones preferentially entrain water with elevated phytoplankton concentrations, such as the South Indian Ocean, or in some mid-ocean gyre regions where small-amplitude eddies form (e.g., the oligotrophic South Pacific), an ecosystem response to eddy-induced Ekman pumping is observed. Conversely, cyclones in these regions entrain water that is low in chlorophyll, resulting in negative chlorophyll anomalies that are sustained by Ekman downwelling throughout the eddy lifetimes. The phytoplankton response to eddy-induced Ekman upwelling in anticyclones is seasonal, occurring only during the winter. It is proposed that the mechanism for the lack of ecosystem response to eddy-induced Ekman upwelling during the summer is the decoupling of the mixed layer from the nutricline.

The observations presented in this dissertation provide a baseline from which coupled ocean circulation and biogeochemical models can be assessed. If coupled models are able to reproduce correctly the observed influence of mesoscale eddies on photoautotrophic communities, further insight into the mechanisms for this variability could be gained from the model output using the methodologies developed in this dissertation together with
investigation of subsurface variability in the models below the depth to which chlorophyll can be inferred from the satellite observations.
Satellite Observations of the Influence of Mesoscale Ocean Eddies on Near-Surface Temperature, Phytoplankton and Surface Stress

by

Peter Gaube

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APPROVED:

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Major Professor, representing Oceanography

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Dean of the Department College of Earth, Ocean and Atmospheric Sciences

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Dean of the Graduate School

I understand that my dissertation will become part of the permanent collection of Oregon State University libraries. My signature below authorizes release of my dissertation to any reader upon request.

__________________________
Peter Gaube, Author
ACKNOWLEDGEMENTS

Academic

I am forever indebted to Dr. Dudley Chelton. Dudley taught me how to think critically, analyze data and interpret results. Without his guidance I truly believe I would have floundered under the enormous quantity of data required to investigate the influence of eddies on near-surface temperature, phytoplankton and surface stress. His guidance allowed me to develop as a scientist. I know that I tested Dudley’s patience, but he never gave up on our research, nor on me. Dudley always tells it as it is, an important trait in someone preparing young scientist for the often grueling and highly competitive field of oceanography. I will always appreciate his honesty and sense of humor.

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“Truly great friends are hard to find, difficult to leave, and impossible to forget.

-Author G. Randolf
CONTRIBUTION OF AUTHORS

Dr. Peter G. Strutton of the University of Tasmania contributed to Chapters 2 and 3 through discussions of methodology and results.

Dr. Michael J. Behrenfeld of Oregon State University contributed to Chapters 2 and 3 through discussions of methodology and results and provided the phytoplankton carbon observations.

Dr. Larry W. O’Neill of Oregon State University contributed to chapter 1 through discussions of methodology and results and providing computer code to aid in the processing of the QuikSCAT data.
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4.1 Estimated oscillatory period (in days) of $\partial C_{phyto}/\partial t$ and $\partial CHL/\partial t$ within eddies of the 5 study regions. Averages shown in bottom row exclude those periods $\geq 45$ days.

4.2 Initial condition and values for the Franks et al. (1986) NPZ model.
1. GENERAL INTRODUCTION

1.1. Historical Background and Motivation

The first oceanographic observations of mesoscale eddies were made in the 1930’s by Columbus Iselin and colleges during quarterly Montauk-to-Bermuda cruises (Cullen, 2005). When successive sections through the Gulf Stream revealed what appeared to be multiple crossings of the Stream, Iselin suggested that these features were solitary internal waves. These peculiar rotating currents were later identified to be mesoscale eddies, but only after the advent of smoked-glass-slide bathythermographs and loran navigation systems which allowed for the collection of temperature profiles and the determination of a ship locations to within about 1 km. These two tools, developed during the second world war, were used to study the mesoscale structure of the Gulf Stream and the first intentional measurements of a mesoscale eddy were made during a 10-day occupation of a cold-core, cyclonic Gulf Stream ring (Iselin and Fuglister, 1948). It wasn’t until the mid-1960’s that the first study focused solely on the dynamics of mesoscale eddies was conducted (Fuglister, 1972), marking the beginning of the scientific investigation of mesoscale meanders and eddies.

Studies aimed at observing the influence of mesoscale eddies on marine ecosystems began late in the last century (1976-1977). Early observations of bio/physical interactions
within eddies were limited to in situ measurements made by sea-going oceanographers. These oceanographic research voyages yielded fascinating observations of the impact of eddies on phytoplankton, zooplankton, nekton and fishes that sparked considerable interest in the influence of mesoscale eddies on organisms in the ocean (an interesting review of early observations made in the Gulf Stream was published by the “Ring Group”, Backus et al., 1981). These early observations were limited to relatively short occupation of just a few eddies.

Today, eddies can be observed from satellites, providing global coverage at near weekly intervals. The sea surface height (SSH) signatures of eddies can be measured by satellite altimeters, which were first developed in the 1960s. SSH measurements from a single satellite altimeter are not sufficient, however, to resolve the $\sim 200$ km diameters of mesoscale eddies. The low-resolution maps of SSH constructed from a single satellite altimeter lead to the interpretation of westward propagating SSH anomalies as linear Rossby waves (Chelton and Schlax, 1996). Careful investigation of the observed westward propagation speed of these features revealed inconsistencies with the theoretical phase speed of Rossby waves. In the early 2000s after merging of SSH measurements from two simultaneously operating altimeters, one in a 10-day exact repeat orbit and another in a 35-day exact repeat orbit, into objectively mapped SSH fields, it became evident that these westward propagating features are nonlinear mesoscale eddies rather than Rossby waves (Chelton et al., 2007, 2011b).

With the 1978 launch of the Coastal Zone Color Scanner (CZCS), one of eight instruments flown aboard the Nimbus-7 satellite, the profound impact that eddies and meanders on marine phytoplankton became apparent (Hovis, 1980; Abbott and Zion, 1985). This early proof-of-concept mission operated intermittently for 91 months and led the way for follow-on missions, resulting in the rich archive of bio-optical observations from the visible portion of the electromagnetic spectrum available today.
It is now common knowledge that eddies are ubiquitous features of the World Ocean. Recent investigations of the sea surface height (SSH) signatures of eddies revealed that most of the eddies in the midlatitudes are nonlinear, containing a region of trapped fluid within their interiors (Chelton et al., 2011b). Moreover, westward co-propagating SSH and chlorophyll anomalies previously attributed to linear Rossby waves have been shown to be attributable to nonlinear mesoscale eddies, laying to rest the much debated influence of Rossby wave on near-surface chlorophyll (Chelton et al., 2011a).

The work presented in this dissertation explores the signatures of mesoscale eddies in the wealth of satellite observations available to the modern-day oceanographic community. By guiding our research with altimetrically derived eddy trajectories, we collocate numerous other data sets (both satellite and in situ) to the interiors of mesoscale eddies, allowing the first truly global investigation of the influence of mesoscale eddies on near-surface temperature, phytoplankton and surface stress.

1.2. Overview of Dataset and Methods

The datasets and methodologies common to the analyses in chapters 2 – 4 of this dissertation are described below. Dataset and methods not common to all of the chapters are include within each chapter.

1.2.1 Sea Surface Height, Geostrophic Currents and Eddy Tracking

The investigation of eddy-induced Ekman perturbation of sea surface temperature (SST), surface stress and phytoplankton communities requires knowledge of the amplitudes, scales and locations of mesoscale eddies. The merged SSH measurements from two simultaneous operating altimeters allows the detection of mesoscale ocean eddies that are not resolved in SSH fields constructed from a single altimeter (Chelton and Schlax, 2003). The merged SSH fields were obtained from Collecte Localis Satellites (CLS/AVISO) at
7-day intervals. While the datasets begin in October 1992, the analysis presented in chapter 2 restrict attention to the 7.5-year June 2002 - November 2009, when high-quality, cloud-free observations of SST are available (see section 2.4.1). The analysis presented in chapter 3 restricts attention to the 8-year period from 4 August 1999 through 26 December 2007 when both QuikSCAT and SeaWiFS data are available (see sections 1.2.2 and 1.2.3, respectively). The analysis presented in chapter 4 utilizes the full data record of each of the data sets investigated to maximize the number of eddy realizations that are used in each of the eddy composites.

The SSH fields used in this study are the Reference Series (REF) that is constructed by merging Topex/Posiden or Jason-1 measurements with ERS-1, -2 or ENVISAT (Ducet et al., 2000). The wavelength resolution of the merged SSH fields is abut 2° in latitude by 2° in longitude, which corresponds to Gaussian e-folding scales of about 40 km (Chelton et al., 2011b). These SSH fields were spatially high-pass filtered to remove the large-scale effects of seasonal heating and cooling by attenuating variability with wavelength scales larger than about 10° in latitude by 20° in longitude, thus isolating mesoscale variability.

As described in detail in appendix B of Chelton et al. (2011b), mesoscale eddies were identified and tracked based on closed contours of SSH. The eddy amplitude at each weekly time step along its trajectory is defined to be the difference between the SSH at the eddy centroid and the value along the eddy perimeter, delineated as the outermost closed contour of SSH that defines a compact structure. The eddy interior is defined to be the region enclosed within this SSH contour. The rotational speed $U$ for each observation of a propagating eddy is defined to be the circum-average geostrophic speed along the contour of SSH within the eddy interior around which this quantity is maximum. The horizontal radius $L_s$ of the eddy is defined to be the radius of a circle with surface area equal to that enclosed by this SSH contour along which the rotational speed is maximum.
1.2.2 Surface Winds

Vector winds and wind-driven Ekman pumping were estimated from 10 m wind measurements by the SeaWinds scatterometer onboard the QuikSCAT satellite. The QuikSCAT mission began on 19 July 1999 and ended on 23 November 2009. Scatterometers infer equivalent neutral wind velocity at 10 m relative to the moving sea surface (e.g. Ross et al., 1985; Chelton and Freilich, 2005). This relative wind can be expressed as

\[ u_{rel} = u_a - u_o, \]  

(1.1)

where \( u_a \) is the air velocity and \( u_o \) is the surface ocean velocity. The QuikSCAT measurements of \( u_{rel} \) for each orbital period were obtained from Remote Sensing Systems (http://www.ssmi.com) and gridded onto a \( 1/4^\circ \times 1/4^\circ \) grid using a loess filter with a radial half-span of 80 km. Observations within \( \sim 100 \) km of the swath edge were excluded to reduce directional ambiguities and avoid problems with spatial derivatives near the swath edges.

The surface stress \( \tau \) was estimated from the QuikSCAT equivalent neutral relative winds using the bulk formula;

\[ \tau = \rho C_D u_{rel} |u_{rel}|, \]  

(1.2)

where \( \rho \) is the air density (considered here to be constant and equal to 1.2 kg m\(^{-3}\)) and \( C_D \) is the speed-dependent drag coefficient for neutrally stable conditions. For the calculations in this study, the Large and Pond (1981) formulation for \( C_D \) was used with the low wind speed modification recommended by Trenberth et al. (1990).

The curl of the surface stress \( \nabla \times \tau \) was calculated in-swath and subsequently time
averaged along with the relative wind speed and vector components. Since the wind field $\mathbf{u}_w$ has scales larger than the order 100 km radii of the mesoscale eddies that are resolved in altimeter data from which eddies are identified for this study (Chelton et al., 2011b), the curl of the surface stress is attributable predominantly to the vorticity of the eddy surface currents $\mathbf{u}_o$. The wind stress that drives Ekman pumping within eddy interiors is thus a function of the amplitude and horizontal scale of the eddy, as well as the background wind field.

To isolate the same time scales of variability that are resolved by the objective analysis procedure used by AVISO to process the SSH fields, the various wind fields considered in this study were constructed at the same 7-day intervals as the SSH observations by temporally low-pass filtering with a half-power filter cutoff of 30 days. The wind fields were then spatially high-pass filtered to remove large-scale features that are unrelated to the mesoscale variability that is of interest in this study. After experimentation with different filter spans, half-power filter cutoffs of 6° of longitude by 6° latitude were chosen as a compromise between attenuating unwanted large-scale atmospheric variability and retaining variability at the oceanic mesoscales.

Ekman pumping was computed as

$$ W_E = \frac{1}{\rho} \frac{\nabla \times \mathbf{r}}{f} $$

(1.3)

where $\rho = 1020 \text{ kg m}^{-3}$ is the (assumed constant) surface density of sea water, $f = 2\Omega \cos \theta$ is the Coriolis parameter for latitude $\theta$ and Earth rotation rate $\Omega$ and $\zeta$ is the vorticity of the surface currents estimated geostrophically from SSH fields that were filtered using the same $6^\circ \times 6^\circ$ applied to the wind measurements.
1.2.3 Chlorophyll-a and Phytoplankton Carbon

To first order, satellite-derived near-surface chlorophyll concentration $CHL$ represents estimates of the biomass of primary producers. Phytoplankton intracellular chlorophyll concentrations are “plastic”, responding to changes in light, nutrients, taxonomy, and other environmental stressors (Laws and Bannister, 1980; Geider, 1987; Falkowski and LaRoche, 1991). This plasticity must therefore be considered when using $CHL$ as an indicator of changes in phytoplankton communities. The adjustment of $CHL$ by a cell to the ambient light field is termed photoacclimation. Because of this physiological response, $CHL$ alone is an incomplete and sometimes misleading metric for the investigation of the phytoplankton response to physical forcing. Phytoplankton carbon biomass ($C_{phyto}$) is insensitive to photoacclimation. Algorithms for estimation of $C_{phyto}$ from the backscattering of remote sensing radiances have been developed by Behrenfeld et al. (2005) and refined by Westberry et al. (2008). While these algorithms are less mature than the algorithms for $CHL$ concentration from satellite measurements of ocean color, the resulting estimates of $C_{phyto}$ provide at least a qualitative measure of biomass.

The ratio of $CHL$ to $C_{phyto}$, hereafter referred to as $r_C$, can provide insight into physiological processes occurring at the community level. This ratio is a function of the rate at which the phytoplankton community is growing (the phytoplankton growth rate $\mu$) and the mixed layer light field to which the community has acclimated. Because of the dependence of $r_C$ on the ambient light field, the computation of $\mu$ requires the normalization of observed $r_C$ by a maximum potential $r_C^{max}$ at the observed mixed layer light levels. Previous studies have used the output of a non-eddy-resolving, data assimilating global ocean model to estimate the mixed layer light field (Behrenfeld et al., 2005; Westberry et al., 2008). Because of our interest in bio/physical interactions in the interiors of mesoscale eddies, a suitable parameterization of the mixed layer light field is not available for each eddy observation. However, we estimate $\mu$ directly in a series of case study eddies.
for which MLD is estimated using Argo float profiles (see section 1.2.5).

The ratio $r_C$ also provides insight into the degree of photoacclimation that a particular photoautotrophic community is experiencing. Cells growing in conditions of low light levels have higher values of $r_C$ than cells growing in conditions of the same temperature and nutrient availability, but with high irradiance. This increase in $r_C$ ratio in low light levels is caused by the intercellular increase in light harvesting complexes that allow the phytoplankton to photosynthesize. Conversely, cells grown in a high irradiance environment have lower values of $r_C$.

Near-surface $\text{CHL}$ and $C_{\text{phyto}}$ were derived from observations of water-leaving radiance in the visible spectrum made by the SeaWiFS sensor onboard the Orbview-2 satellite. The SeaWiFS mission began on 19 September 1997 and ended on 10 September 2010. $\text{CHL}$ and particulate backscattering coefficients at 440 nm ($b_{bp}$) are calculated from raw water leaving radiances using the Garver-Siegel-Maritorena (GSM) semi-analytical ocean color algorithm (Garver and Siegel, 1997; Maritorena et al., 2002; Siegel et al., 2002). $C_{\text{phyto}}$ is estimated from $b_{bp}$ by first subtracting a background value attributed to the scattering of non-algal particles and then scaling $b_{bp}$ by a constant factor chosen to give $r_C$ values that are consistent with laboratory results and an average phytoplankton contribution to total particulate organic carbon of 30%. Further details on how $C_{\text{phyto}}$ is estimated from $b_{bp}$ and the global variability of $C_{\text{phyto}}$ are given by Behrenfeld et al. (2005).

Since clouds are opaque to electromagnetic radiation emitted in the visible spectrum, $\text{CHL}$ and $C_{\text{phyto}}$ estimates can be made only in cloud-free conditions. Data gaps during cloudy conditions must be taken into consideration when filtering the ocean color data in order to obtain reasonable estimates of $\text{CHL}$, $C_{\text{phyto}}$ and $r_C$ within the observed mesoscale eddies. The $\text{CHL}$ and $C_{\text{phyto}}$ fields were first $\log_{10}$ transformed and averaged onto the same $1/4^\circ$ latitude by $1/4^\circ$ longitude grid as the SSH and wind fields described in sections
1.2.1 and 1.2.2. The log transformation accounts for the highly skewed distributions of the untransformed data in many regions of the World Ocean (Campbell, 1995).

The time series of daily CHL and $C_{phyto}$ values at each grid point were then low-pass loess filtered and gridded at weekly intervals to attenuate variability with periods shorter than 30 days and reduce data gaps from cloud contamination. The gridded CHL and $C_{phyto}$ fields were then inverse transformed by taking the antilogarithms. Although a historical precedent exists to analyze the mesoscale anomalies of $\log_{10}$ transformed CHL (Cipollini et al., 2001; Uz et al., 2001; Killworth et al., 2004; Uz and Yoder, 2004; Siegel et al., 2007; Gutknecht et al., 2010; Siegel et al., 2011), there is merit in analyzing anomalies in the original units of $mg m^3$ for CHL and $C_{phyto}$ (see appendix D).

The ratio $r_C$ was gridded and low-pass filtered in the same manner as the CHL and $C_{phyto}$ time series, except without log transforming because the range of variability of $r_C$ is small and is more symmetrically distributed than either CHL of $C_{phyto}$.

To isolate spatial mesoscale variability, the weekly maps of CHL, $C_{phyto}$ and the $r_C$ were spatially high-pass filtered with the same $6^\circ \times 6^\circ$ half spans that were applied to the Ekman pumping fields discussed in section 1.2.2. The anomaly fields are defined by:

$$CHL = \overline{CHL} + CHL'$$

$$C = \overline{C} + C'$$

$$r_C = \overline{r_C} + r'_C,$$

where the overbar denotes the smoothed fields that are removed from the total fields to create the anomalies that are denoted with primes. To minimize filter edge effects caused by the exponential decay of CHL away from the coast, observations within a radial span of 6 grid points (approximately 150 km) from any location identified as “land” were removed.
1.2.4 Mixed Layer Depth

The upper ocean responds to wind stress and solar irradiance by the development of a surface mixed layer. The mixed layer depth (MLD) can be defined from profiles of temperature, density or both based on the depth at which the vertical gradient becomes large. The MLDs in the interiors of SIO eddies were estimated for this study from Argo float profiles. Argo floats have measured temperature and salinity profiles since the year 2000 in the upper 1000 m at 70 m intervals with closer spacing in the upper 400 m (http://www.argo.ucsd.edu). Argo profiles were collocated to the interiors of the eddies selected for this study at the closest weekly time step to the time of each Argo profile. An Argo profile was considered to be inside of an eddy if it occurred within the outermost closed contour of SSH used to defined the eddy periphery (see section 1.2.1). A density-based estimate of the MLD for each collocated Argo profile was obtained from the global data base compiled by Holte and Talley (2009) that is available at http://mixedlayer.ucsd.edu. We used their density-based MLD estimates that initializes with a variable density threshold and then models the shape of the density profile. The algorithm assembles a suite of possible MLD values and looks for groupings and patterns within the possible MLDs to select a best estimate of MLD for each profile. The seasonal cycles of MLD from collocated profiles were calculated separately for cyclones and anticyclones by regressing the time series of MLD onto the annual cycle and its first harmonic.

1.2.5 Phytoplankton growth rate $\mu$

The growth rate of the phytoplankton community ($\mu$) can be estimated at the location and time of each individual Argo float profile that occurred during the SeaWiFS data record. Following the methodology presented by Behrenfeld et al. (2005), the growth rate is estimated in units of number of cell divisions per day by;
\[
\mu = \mu_{\text{max}} \left( \frac{r_C}{r_{C}^{\text{max}}} \right) \left( 1 - e^{-3I_g} \right),
\]

(1.7)

where \( \mu_{\text{max}} \) is the maximum potential growth rate of phytoplankton community and is estimated at 2 \( \text{day}^{-1} \). The maximum possible value \( r_{C}^{\text{max}} \) for the ratio \( r_C \) was estimated as described by Behrenfeld et al. (2005) based on an empirical fit to global observations of \( r_C \) made by SeaWiFS for \( I_g \) ranging from nearly 0 to 2 \( \text{Ein} \text{ m}^2 \text{ hr}^{-1} \)

\[
r_{C}^{\text{max}} = 0.22 + (0.045 - 0.022)e^{-3I_g},
\]

(1.8)

where \( I_g \) is the median mixed layer growth irradiance experienced by the photoautotrophic community. The median is chosen here because phytoplankton are assumed to be continuously vertically mixed throughout the mixed layer, thus experiencing the median mixed layer growth irradiance. Growth irradiance was computed from individual Argo profiles following the method of Behrenfeld et al. (2005):

\[
I_g = \text{PAR} \ e^{-k_d \ \text{MLD}/2}
\]

(1.9)

where PAR is the photosynthetically active radiation and \( k_d \) is the light extinction coefficient at 490 nm as observed by SeaWiFS, both of which were obtained from http://oceancolor.gsfc.nasa.gov as daily, 9 km gridded fields. These data were bin averaged onto our 1/4\( ^\circ \) \( \times \) 1/4\( ^\circ \) grid and interpolated and smoothed in time with the same 30-day half span that was applied to \( \text{CHL}, C_{\text{phyto}} \) and the \( r_C \) fields. The multiplicative factor \( (1 - e^{-3I_g}) \) in (1.7) is a low-light adjustment to account for a decrease in \( \mu \) at low light levels where physiological adjustments in pigment concentration are insufficient to maintain constant levels of light absorption.
Satellite Observations of Mesoscale Eddy-Induced Ekman Pumping

Peter Gaube, Dudley B. Chelton and Larry O’Neill
2. SATELLITE OBSERVATIONS OF MESOSCALE EDDY-INDUCED EKMAN PUMPING

2.1. Abstract

Two mechanisms for self-induced Ekman pumping within the interiors of mesoscale ocean eddies are investigated here in chapter 2. One arises from the surface stress that occurs because of differences between surface vector wind and the surface ocean velocities. Because the wind field generally has spatial scale larger than the order 200 km diameters of mesoscale eddies, the vorticity of this relative wind field is determined primarily by the vorticity of the eddy surface currents. The surface currents therefore induce a wind stress curl with polarity opposite that of the eddy, resulting in Ekman upwelling and downwelling centered on the cores of anticyclones and cyclones, respectively. Ekman pumping can also arise from eddy-induced spatial variability of the sea surface temperature (SST) field which generates a wind stress curl in regions of crosswind SST gradients through a response of surface winds to SST-induced surface heating variations. Because eddy-induced SST variability is controlled primarily by horizontal advection by the rotational velocity within the eddies, this air-sea interaction generates a dipole structure of wind speed perturbations in the eddy interiors.

The objective of the investigation presented in this chapter is to determine the spatial structures and relative magnitudes of these two mechanisms for eddy-induced Ekman pumping. This is achieved by collocating satellite-based measurements of SST, geostrophic surface currents and surface winds to the interiors of eddies identified and tracked with an automated procedure applied to altimeter-based sea-surface height fields. On average, the surface current-induced Ekman pumping velocities are $O(10 \text{ cm day}^{-1})$. While this surface current-induced Ekman pumping depends only weakly on the wind direction, Ekman pumping associated with eddy-induced SST variations depends strongly on both the
magnitudes and structures of the SST anomalies and the ambient wind direction. Over most of the World Ocean, Ekman pumping from eddy-induced SST variations is shown to be an order of magnitude smaller than that associated with eddy-induced surface currents. SST-induced wind stress curl reinforces the magnitude of surface current-induced Ekman pumping in some regions and reduces it in others. Notable exceptions are the midlatitude zonal jet-like currents of the western boundary current extension and the Antarctic circumpolar current, where SST gradients are strong and the two mechanisms for Ekman pumping are comparable.

2.2. Introduction

Mesoscale eddies occupy approximately 25 – 30% of the surface area of the World Ocean at any time (Chaigneau et al., 2009). The surface currents associated with mesoscale ocean eddies impart a curl in the surface stress from the relative motion between surface air and water (Stern, 1965). This wind stress curl has a polarity opposite that of the vorticity of the eddy, thus attenuating the eddies by generating Ekman upwelling in the cores of anticyclonic eddies and downwelling in the cores of cyclonic eddies (Dewar and Flierl, 1987). Eddy surface current-induced Ekman upwelling has been identified as a mechanism for sustaining enhanced primary production in the cores of anticyclonic eddies in the North Atlantic (Martin and Richards, 2001; McGillicuddy et al., 2007).

In addition to the surface current-induced Ekman pumping, air-sea interaction associated with eddy-induced spatial variations of SST generates a wind stress curl and therefore Ekman pumping that are proportional to the crosswind SST gradient (Chelton et al., 2004; O’Neill et al., 2010). The feedback of this air-sea interaction on the ocean has been shown to exert a significant attenuation on the mesoscale eddy field in an idealized model of an eastern boundary current system (Jin et al., 2009).
While the two mechanisms for eddy influence on Ekman pumping have been previously identified, their relative importance has not been investigated. The objective of this study is to quantify the magnitudes of the contributions of surface current effects and air-sea interaction to the total eddy-induced Ekman pumping field and to investigate how they vary geographically over the World Ocean.

2.3. Historical Background

The influence of eddy surface currents on the surface stress has been recognized theoretically for nearly half a century (Stern, 1965). However, direct observations of this effect could not be obtained before the advent of satellite scatterometers. Scatterometers infer the vector wind at 10 m, relative to the surface ocean velocity, from the roughness of the sea surface (see, for example Chelton and Freilich, 2005). Surface stress is then obtained from the relative vector wind by the bulk aerodynamic formula (see section 1.2.2).

The first observations of the effects of eddy surface currents on the relative wind (and by inference, the surface stress) were reported by Cornillon and Park (2001) from scatterometer measurements over Gulf Stream rings (see also Park et al., 2006). Since winds usually have scales larger than the order 100-km radii of mesoscale eddies, the vorticity (curl) of the relative wind over eddies is determined primarily by the eddy surface currents. The vorticity of a rotating eddy generates an opposite vorticity in the relative wind. Anticyclonic and cyclonic eddies thus generate positive and negative wind stress curl and therefore Ekman upwelling and downwelling, respectively.

The influence of eddy-induced SST perturbations on the surface wind field was also first documented over Gulf Stream rings (Park and Cornillon, 2002) and then later further developed by (Park et al., 2006). This air-sea interaction phenomenon is well described in
the literature (see, for example, the reviews by Small et al., 2008; Chelton and Xie, 2010, and recent work by O’Neill et al. (2012)). Briefly, SST modifies the mixing and stability of the marine atmospheric boundary layer (MABL). Stabilization of the MABL over cool water decouples the surface winds from the winds aloft, thus decreasing the surface winds. In contrast, local heating over warm water destabilizes the MABL, thus enhancing vertical mixing of momentum down to the sea surface and increasing the surface winds. The effects of SST are further accentuated by the creation of an SST-induced pressure gradient force across temperature fronts. After spatially high-pass filtering to remove variability with scales larger than about 1000 km, the relationship between SST and wind speed on scales smaller than ~ 1000 km (referred to here as the mesoscale) is approximately linear (O’Neill et al., 2012) and can be expressed as

\[ W' = \alpha_{spd} T' \]  

(2.1)

where \( T' \) and \( W' \) are from the spatial high-pass filtered SST and wind speed fields. The coupling coefficient \( \alpha_{spd} \) has been shown to vary somewhat geographically and seasonally with an approximate range of variability of \( 0.3 \leq \alpha_{spd} \leq 0.6 \) (O’Neill et al., 2012).

In regions where large eddies dominate the mesoscale SSH variability, eddies have been observed to have SST signatures that closely resemble their SSH structure (e.g., the warm and cold cores in anticyclonic and cyclonic eddies, respectively, see Park and Cornillon, 2002; Park et al., 2006; Hausmann and Czaja, 2012). In regions of less energetic mesoscale eddies, the SST signature associated with eddies is better represented by an asymmetric dipole, with the sign and orientation of the leading (westward) pole being a function of eddy polarity and the background SST gradient (Hausmann and Czaja, 2012). This is analogous to the dipole structure observed from satellite measurements of near-surface chlorophyll in mesoscale eddies (Chelton et al., 2011a). The structures of
eddy-induced SST perturbations are quantified both regionally and globally in this study.

Park and Cornillon (2002) and Park et al. (2006) showed that the eddy-induced SST anomalies associated with Gulf Stream rings generate wind speed and direction perturbations. One of the objectives of this study is to extend their analysis to the global ocean and show that the well-documented linear relationship between SST and wind speed perturbations is robust over the interiors of midlatitude mesoscale eddies.

Numerous modeling studies have shown that there is a significant reduction of eddy kinetic energy (EKE) when the effects of surface current on the surface stress are accounted for in the wind stress forcing (Dewar and Flierl, 1987; Eden et al., 2009; McClean et al., 2010; Anderson et al., 2011). Since the sign of the surface current-induced surface stress curl is opposite that of the eddy, the net effect of including this surface current correction to the surface stress formulation is to attenuate the kinetic energy of the eddies (Dewar and Flierl, 1987). Eden et al. (2009) reported a 50% reduction in EKE when eddy-induced Ekman pumping was included in their model of the North Atlantic. Similarly, Anderson et al. (2011) also reported a reduction in the EKE of their model when eddy-induced Ekman pumping was included. McClean et al. (2010) showed that accounting for surface currents in the surface stress in a fully coupled ocean-atmosphere model altered the pathways, amplitudes and lifetimes of large anticyclonic eddies spawned along the Agulhas retroflection. The pathways of Agulhas anticyclones into the South Atlantic were visually more similar in the coupled model than those generated in an uncoupled ocean model when compared to those observed by altimeters. The amplitudes of the Agulhas anticyclones in an uncoupled run of the model remained too large and they propagated along a path that was displaced to the northeast of the observed trajectories.

Ekman pumping associated with eddy-induced SST influence on surface winds has thus far received much less attention from the modeling community. In a modeling study that focused on the role of this air-sea interaction in attenuating EKE, the inclusion of
this effect in a coupled model of an idealized upwelling system reduced the EKE by 25% (Jin et al., 2009). The geographic displacement of the wind stress curl anomalies from the cores of the mesoscale eddies disrupted the axisymmetric structure of the eddies, thus attenuating the mesoscale eddy field. Because the SST signatures of cyclones were stronger than those of anticyclones owing to ageostrophic effects, cyclones were attenuated more than anticyclones.

In addition to the influence of eddy-induced Ekman pumping on eddy kinematics, the vertical velocities associated with self-generated Ekman pumping can also influence biological communities in the interiors of mesoscale eddies. Martin and Richards (2001) hypothesized that eddy-induced Ekman pumping was a possible mechanism for sustaining an anomalous phytoplankton bloom that was observed in the core of a mode-water type anticyclonic eddy in the North Atlantic. They estimated eddy-induced Ekman pumping velocities from geostrophic surface currents calculated for a Gaussian-shaped SSH signature scaled to match the eddy surface current velocities measured from ship transects across the eddy (this idealized eddy has an amplitude of roughly 17 cm). For a spatially uniform 15 m s\(^{-1}\) surface wind, the upwelling was in excess of 1 m day\(^{-1}\) within the eddy core. For a more typical wind speed of 7 m s\(^{-1}\), the Ekman pumping velocity for the same example eddy would be 45 cm day\(^{-1}\).

An intensive ship survey of another mode-type anticyclone in the Sargasso Sea definitively showed the importance of eddy surface current-induced upwelling (McGillicuddy et al., 2007). Phytoplankton concentration within the core of the eddy was eight standard deviations higher than the mean background field. Sulfur hexafluoride tracer released in the eddy core upwelled at a rate of 40 cm day\(^{-1}\), which compared well with the calculated Ekman pumping velocity computed from only the eddy surface current effects averaged over the time period of the tracer experiment (McGillicuddy et al., 2007; Ledwell et al., 2008).
From an analysis of simulations using an eddy-resolving coupled biogeochemical model of the North Atlantic, Eden et al. (2009) concluded that eddy-induced Ekman pumping resulted in a 5% decrease in the integrated net primary production in the North Atlantic. In light of the observation of increased production by McGillicuddy et al. (2007), the overall decreases in production in the model simulation is surprising. The decrease is likely attributable to an overall decrease of about 50% in the EKE compared with model runs that did not include the surface current effect on the surface stress (Eden et al., 2009) because of the attenuation effects summarized above. When limiting the analysis to the interiors of mesoscale eddies in a coupled ecosystem model, Anderson et al. (2011) found that inclusion of eddy-induced Ekman pumping significantly affected not only the phytoplankton growth rates, but also phytoplankton community compositions. Phytoplankton biomass increased significantly in the cores of mode-type anticyclones, in which the near-surface isopycnals are shoaled over a lens of water below which isopycnals are depressed. The biomass and growth rates of diatoms increased in regular anticyclones as well. As in all other modeling studies, eddy-induced Ekman pumping from surface current effects reduced the amplitudes of all eddies in the Anderson et al. (2011) simulation. Moreover, the upward displacement of shallow isopycnals decreased in the cores of cyclones and increased in the cores of mode-type anticyclones. For reasons that were not fully explained, the inclusion of eddy-induced Ekman pumping increased the relative abundance of mode-type eddies (Anderson et al., 2011).

Modeling studies and limited observational studies have thus shown that eddy-induced Ekman pumping is important to both the physics and the biology in mesoscale eddies. With the exception of the modeling study by Jin et al. (2009) that considered only the effects of eddy-induced SST perturbations on Ekman pumping, the focus to date has been almost entirely on the surface current contribution to eddy Ekman pumping. The objective of this study is to investigate the relative importance of these two effects from
The satellite data sets used in this analysis and the methods used to isolate the surface current and SST effects on eddy-induced Ekman pumping are described in sections 1.2.1, 1.2.2 and 2.4. In preparation for the analysis of SST-induced Ekman pumping in mesoscale ocean eddies, the geographic structure of eddy-induced SST anomalies was determined and the coupling of the surface wind to these anomalies is presented in section 2.5.1. A series of idealized Gaussian-shaped anticyclonic eddies are considered in section 2.5.2 to demonstrate the nature of Ekman pumping due to eddy surface currents and SST anomalies. Global maps of the magnitude of Ekman pumping due to eddy surface current and SST effects are presented in section 2.5.3 and 2.5.4. A detailed examination of the geographic structure of eddy-induced Ekman pumping within the interiors of tens of thousands of midlatitude eddies is presented in section 2.5.5. To investigate the relative contribution of SST and surface current effects to observed eddy-induced Ekman pumping velocities, eight study regions are analyzed in section 2.6. The regional variability is discussed in section 2.7 in the context of the global results and we conclude in section 2.8 that the surface current contribution to eddy Ekman pumping is more important than eddy-induced SST effects over most of the World Ocean. However the SST influence is not negligible and in fact dominates surface current effects in much of the Southern Ocean and the midlatitude extensions of some western boundary currents.

### 2.4. Methods

The two contributions to Ekman pumping within the interiors of mesoscale ocean eddies are estimated for this investigation from satellite observations of surface winds, SST and SSH over a 7.5 year period June 2002 - November 2009, during which observations are available for all three variables. The June 2002 starting date is dictated by the launch of
AMSR-E onboard the Aqua satellite from which all-weather microwave measurements of SST are available to create the optimally interpolated SST product utilized in this study. The November 2009 end date corresponds to the end of the QuikSCAT data record.

Details of the sea surface height fields, data processing and eddy tracking are given in section 1.2.1. A total of 12,144 anticyclonic and 12,609 cyclonic eddies with lifetimes of 12 weeks or longer were tracked during the 7.5 year study time period.

The processing of QuikSCAT data and filtering applied to isolate variability at the oceanic mesoscale are described in section 1.2.2.

2.4.1 NOAA Optimally Interpolated SST Observations

The SST fields used in this study are the optimally interpolated SST product produced by the NOAA National Climatic Data Center (Reynolds et al., 2007), that combines microwave and infrared satellite observations with in-situ measurements of SST to obtain daily, global fields on a 1/4° latitude by longitude grid (data available at ftp://eclipse.ncdc.noaa.gov/pub/OI-daily-v2/NetCDF).

The quality of these SST fields depends upon the availability of the microwave measurements available globally from the Advanced Microwave Scanning Radiometer on the Earth Observing System Aqua satellite (AMSR-E). The AMSR-E data record began on 18 June 2002 and ended on 4 October 2011. To isolate variability on oceanic mesoscales, the daily fields of SST were temporally and spatially filtered in the same manner as the QuikSCAT wind observations (section 1.2.2).

2.4.2 Collocation of Observations within Eddies Interiors

The SST and wind fields filtered as described above were collocated within the interiors of the eddies identified from their SSH signatures. Composites of SST, wind speed and Ekman pumping were then constructed for tens of thousands of midlatitude eddies, defined as having originated between 15° and 45° latitude in both hemispheres (figure
2.1a). On average, midlatitude cyclones are somewhat larger in amplitude than anticyclones (average amplitudes of 8 cm and 7 cm, respectively, figure 2.1b) with the larger amplitude eddies (amplitude $\geq 10$ cm) being predominantly cyclonic. This is consistent with the global analysis presented in Chelton et al. (2011b). Histograms of the horizontal scales $L_S$ of these midlatitude eddies are shown in figure 2.1c. Approximately 75% of the eddies have $L_S$ in the range 62 to 128 km. The combination of the average amplitude of midlatitude cyclones being larger than the amplitudes of anticyclones with the radial scale of midlatitude eddies being nearly the same for either polarity results in rotational speeds ($U$) that are on average faster in cyclones (figure 2.1d).

The compositing was done on a common grid by scaling the distances from the eddy centroid to each $1/4^\circ$ grid point in the eddy interior by the instantaneous eddy radius $L_S$, defined as the SSH contour around which the tangential geostrophic velocity within the eddy interior is maximum. Each normalized grid location was then interpolated onto a high-resolution grid for radial distances ranging from 0 to $2L_S$.

It will become apparent in section 2.5.1 that the eddy induced influence on SST consists predominantly of horizontal advection of the background SST gradient by the azimuthal velocity within the eddy interior. As shown schematically in figure 2.2, this results in dipole structures of SST anomalies, with sign depending on the direction of the background SST gradients and the rotational direction of the eddy. The composites of mesoscale SST anomalies were computed in a translating and rotated frame of reference determined by the orientation of the large-scale SST gradient computed from 200-day smoothing of the SST fields. When this ambient SST gradient vector had a nonzero northward component, the SST anomalies were rotated to orient the large-scale SST gradient vector at a polar angle of $90^\circ$. When the ambient SST gradient vector had a nonzero southward component, the SST anomalies were rotated to orient the large-scale SST gradient vector at a polar angle of $-90^\circ$. 
The importance of composite averaging the SST anomalies in a rotated coordinate system defined by the gradient of the ambient SST field is clear from figure 2.2, in which composites calculated from non-rotated observations would blur the dipole structure of the eddy-induced SST perturbations. Because wind speed responds approximately linearly to mesoscale SST anomalies in accordance with previous studies of air-sea interaction over SST frontal regions (e.g. O’Neill et al., 2012), wind speed anomalies were composited in the same coordinate system as the underlying SST anomalies.

It will become apparent in section 2.5.2 that the detailed structure of the surface current contribution to eddy-induced Ekman pumping depends on the direction of the large-scale background winds. Composites of Ekman pumping for all midlatitude eddies were therefore computed in a coordinate system that is rotated to orient the average background wind direction over the eddy to a polar angle of 0°. The background wind is defined from the weekly QuikSCAT relative wind vectors smoothed to attenuate mesoscale variability with wavelength scales shorter than 6° × 6°.

For the eight study regions investigated in section 2.6, composites within eddy interiors include only the times when the eddy centroid is located within the bounds of each region. After an eddy propagates outside of the study region, observations associated with that eddy are no longer included in the composites.

Throughout this study, composite medians are considered rather than composite averages because they are less sensitive to occasional outliers in the anomaly fields.

2.5. Results and Discussion

2.5.1 Eddy-induced SST Perturbations of Surface Wind Speed

The dipole structure alluded to in section 2.4.2 (figure 2.2) is readily apparent, but strongly asymmetric. For eddies rotating clockwise in an SST gradient with a nonzero
northward component ($\partial SST/\partial y > 0$), the leading (westward) pole has a negative SST anomaly (figure 2.3a). As a westward propagating clockwise-rotating eddy propagates through a northward SST gradient, the northward velocity on the western side of the eddy advect cool waters from the southwestern quadrant to the northwestern quadrant, resulting in a negative SST anomaly in the northwestern quadrant. The clockwise-rotating surface currents on the trailing edge act to advect relatively warm water from the northeastern quadrant to the southeastern quadrant, resulting in a positive SST anomaly in the southeastern quadrant. The opposite is true for clockwise-rotating eddies propagating in regions with a background SST gradient having a nonzero southward component ($\partial SST/\partial y < 0$); in this case, a positive SST anomaly in the northwestern quadrant is associated with the leading edge of the eddy and a negative SST anomaly in the southwest quadrant is associated with the trailing edge (figure 2.3c).

For counterclockwise-rotating eddies in a northward SST gradient ($\partial SST/\partial y > 0$), the currents associated with the leading edge of the westward propagating eddy advect warm water from the north, thereby creating a positive SST anomaly in the southwest quadrant of the eddy. On the trailing side, the counterclockwise-rotating currents bring cool water from the southeastern quadrant into the northeastern quadrant, resulting in a negative SST anomaly in the northeastern quadrant (figure 2.3b). The opposite is observed for counterclockwise-rotating eddies in a southward SST gradient ($\partial SST/\partial y < 0$), figure 2.3d).

The geographical patterns of anomalous SST within the eddy interiors in the mid-latitudes are thus seen to be indicative of horizontal advection of SST by the rotational velocities of the eddies. The dipole patterns of SST anomalies result from a combination of the rotational sense of the eddies and the direction of the background SST gradient. A similar horizontal advection of near-surface chlorophyll concentrations ($CHL$) by the rotational velocities of the eddies has also been observed to be the dominant mechanism.
by which eddies influence CHL in the midlatitudes (Chelton et al., 2011a). However, the asymmetry of the dipole of the SST composites is greater than that observed in the chlorophyll composites. This asymmetry is measured by the ratio $r$ of the magnitude of the pole in the left half of the composites to the magnitude of the pole in the right half of the composites and is shown in the title of each panel shown in figure 2.3. Values for $r$ in the SST and windspeed anomalies range between $\sim 2 - 2.5$. In contrast, the asymmetry of the dipoles of near-surface chlorophyll range $\sim 1.3 - 1.7$ (Chelton et al., 2011a). It is shown in appendix C that this larger asymmetry is consistent with the influence of vertical motions on SST anomalies in the cores of the eddies.

Composites of wind speed anomalies in the same rotated coordinate system as the SST composites (figures 2.3e-h) have almost identically the same structure as the SST composites (figures 2.3a-d). The coupling coefficient calculated between SST and wind speed anomalies of all long-lived midlatitude eddies is 0.48 $m$ $s^{-1}$ per °C (figure 2.4). This falls within the range of coupling coefficients previously reported for SST frontal regions (O’Neill et al., 2010; O’Neill, 2012; ?). It is thus seen that the air-sea interaction studied extensively in SST frontal regions also occurs over eddy-induced SST anomalies.

### 2.5.2 Ekman Pumping in Idealized Gaussian-shaped Eddies

The magnitude and geographic structure of the Ekman pumping resulting from eddy surface currents can be assessed from consideration of the SSH structure of an idealized Gaussian shaped mesoscale ocean eddy with axisymmetric structure, amplitude $A$ and e-folding scale $L_e$,

$$\eta(r) = A \ e^{-r^2/L_e^2}$$

(2.2)

where $r$ is the radial distance from the eddy centroid. A series of three anticyclonic
Gaussian eddies with differing amplitudes are used to estimate the magnitude and spatial structure of Ekman pumping from surface currents. The idealized eddies have e-folding radii $L_e$ of 100 km rotating under a uniform zonal wind of $7 \text{ m s}^{-1}$ at 30°N. These values were chosen because they are typical of observed eddies and wind conditions. The largest of the idealized eddies has an amplitude of 22 cm that represents the upper 95th percentile of midlatitude eddies. Eddies with amplitude this large are found in regions of high eddy kinetic energy, such as western boundary currents and their midlatitude extensions (Chelton et al., 2011b). Medium and small eddies with amplitudes of 10 cm and 3 cm, respectively are chosen to represent eddies formed elsewhere. These amplitudes represent the 75th and 25th percentile of all long-lived midlatitude eddies.

The geostrophic current vorticity, SSH and the resultant eddy current-induced Ekman pumping of the idealized anticyclones are shown in row a of figure 2.5 for the case of latitudinally uniform westerly winds. In the largest of the three idealized eddies, Ekman pumping velocities exceed $23 \text{ cm day}^{-1}$ at the eddy core with secondary side lobes of downwelling occur over the northern and southern flanks of the eddy, i.e lateral to the zonal wind direction (figure 2.5 row a, second column). The eddy surface current-induced Ekman upwelling in the core of the medium sized anticyclone is $9.5 \text{ cm day}^{-1}$, and in the smallest amplitude anticyclone, it approaches $2.7 \text{ cm day}^{-1}$ at the eddy core.

Several important characteristics can be deduced from these idealized examples of eddy surface current-induced Ekman pumping. Because the surface current induced wind stress curl has a polarity opposite that of the eddy, current-induced Ekman pumping always generates upwelling in the cores of anticyclones and downwelling in cyclones. It is also evident that the magnitude of surface current-induced Ekman pumping is a nonlinear function of the surface current speed because of the squared dependence (1.2) of the surface stress on the relative wind. The magnitude of current-induced Ekman pumping at the centers of eddies is invariant with wind direction. However the center lobe of strong Ekman
pumping over the cores of the eddies is elongated in the direction of the wind. Likewise, the two weaker side lobes of Ekman pumping with opposing sign are also elongated in the direction of the wind. The two structures shown in row a of figure 2.5 rotate according to the wind direction.

The SST influence does not depend on the Gaussian structure. To assess the relative importance of the influence of surface currents and SST on the total Ekman pumping, we consider an SST structure constructed to match those observed over the average midlatitude northern hemisphere anticyclone propagating in regions with northward SST gradient (i.e figure 2.3c). The magnitudes of the SST anomalies shown in in figure 2.5b have been scaled to have a maximum amplitude of 1.3, 0.6 and 0.25 °C, which represent the same 95th, 75th and 25th percentile of the magnitude of the SST anomalies of long-lived midlatitude eddies averaged within a distance of $L_s$ from the eddy centroids. SSH contours are overlaid to emphasize that the SST anomaly is not located at the eddy core, but instead is an asymmetric dipole with the warmest SST on the leading (westward) edge of the eddy having a larger magnitude than that on the trailing edge.

To estimate the Ekman pumping velocity induced by air-sea interactions associated with eddy-induced SST anomalies, the same 7 m s$^{-1}$ zonal wind is blown over these idealized Gaussian eddies with the prescribed SST anomaly. The crosswind SST gradient field influences the degree to which SST anomalies induce a curl in the wind stress (Chelton et al., 2004; O’Neill et al., 2012). The SST-induced wind stress curl anomalies are computed according to the empirical relation $\nabla \times \tau'_{SST} = \alpha^{strcrl}_c \left( \frac{\partial T}{\partial n} \right)' + \alpha^{strcrl}_d \left( \frac{\partial T}{\partial s} \right)'$ assuming nominal values of both the coupling coefficient of the wind stress curl to the crosswind SST gradient of $\alpha^{strcrl}_c = 0.01$ N m$^{-2}$ per °C and a coupling coefficient of the wind stress curl to the downwind SST gradient of $\alpha^{strcrl}_d = 0.002$ N m$^{-2}$ per °C (see appendix B for details of this calculation).

The magnitude of the SST-induced Ekman upwelling for the largest SST anomaly
(1.3°C) is 14.7 cm day\(^{-1}\) (figure 2.5, row b, column 2). In the idealized scenario presented here, the westerly background wind that blows across the prescribed SST anomalies generates Ekman upwelling to the north of the eddy centroid and Ekman downwelling at the eddy centroid, extending to the south. The magnitude of the upwelling and downwelling are asymmetric, with a maximum downwelling of 12.8 cm day\(^{-1}\) because of spatial asymmetry of the eddy-induced SST anomalies. This geographical structure of SST-induced Ekman pumping is the same for all three prescribed SST anomalies (figure 2.5, row b), except that the magnitude of the upwelling and downwelling decreases linearly with decreasing SST anomaly. The structure of the SST-induced Ekman pumping in figure 2.5b is very different than that of current-induced Ekman pumping shown in figure 2.5a.

To investigate the dependence of SST-induced Ekman pumping on the direction of the background wind, the Ekman pumping for the same three prescribed SST anomalies is shown in figure 2.5c, for the case of southerly winds. The structure of the SST-induced Ekman pumping is qualitatively similar to that driven by the westerly winds except rotated 90°. As mentioned above, the crosswind SST gradient just to the south of the eddy centroid drives Ekman downwelling in excess of 12.8 cm day\(^{-1}\) in the case of westerly winds (figure 2.5b). For southerly winds, however, the crosswind SST gradient to the east of the eddy centroid is slightly larger in magnitude resulting in an Ekman downwelling velocity of 14 cm day\(^{-1}\).

It is thus apparent that the magnitude and spatial structure of SST-induced Ekman pumping is a function of both the magnitude and geographic structure of the SST anomalies and the direction and magnitude of the background wind.

Initially, one can conclude that SST-induced Ekman pumping over eddies is small when compared to the pumping resulting from eddy surface currents. However, in regions of intense SST gradients, the magnitude of SST-induced Ekman pumping can approach, and even exceed, that of the eddy surface current-induced Ekman pumping. From con-
sideration of eight diverse regions of the World Ocean in sections 2.6., it is shown that SST contributions to eddy-induced Ekman pumping are generally small compared with surface current effects.

### 2.5.3 Surface Current-induced Ekman Pumping

Eddy surface current-induced Ekman pumping was calculated from geostrophic velocities within the eddy interiors as described in appendix A. A global map of the average magnitude of this current-induced Ekman pumping is shown in figure 2.6a. Approximately 75% of the 7.5 year averaged eddy current-induced Ekman pumping values fall within the range of 1.4 to 8.6 cm day$^{-1}$ with a 99th percentile of about 23 cm day$^{-1}$. Current-induced Ekman pumping increases towards the equator because of the inverse dependence of Ekman pumping on the Coriolis parameter $f$. In the midlatitudes, current-induced Ekman pumping is highest in regions of energetic mesoscale eddies, such as in the vicinity of boundary currents. Large current-induced Ekman pumping generally occurs where the eddy amplitudes are large (see figure 10 of Chelton et al., 2011b).

### 2.5.4 SST-induced Ekman Pumping

Crosswind SST gradient associated with eddy-induced SST perturbations induce a curl in the wind stress field as summarized in section 2.2.. The analysis in section 2.5.2 (see figure 2.5) showed that the Ekman pumping velocities from eddy-induced SST perturbations are usually smaller than those from the geostrophic eddy surface currents. In regions where the wind blows along strong and persistent SST gradients, however, the wind stress curl from lateral variations of wind speed across the SST front can generate Ekman pumping velocities that are comparable with, and at times larger than, the Ekman pumping from eddy surface currents (figure 2.6b). These regions can be identified from the ratio of SST-induced Ekman pumping to current-induced Ekman pumping shown in figure 2.6c. The magnitude of SST-induced Ekman pumping varies regionally, with
the largest pumping values occurring in the regions of strongest SST-gradients shown in figure 2.6d. This correspondence is shown in figures 2.6c and d by the black contours of $W_{SST}/W_{\nabla \times u_g} = 1$. Almost all of the regions where the mean magnitude of the SST gradient exceeds $2^\circ$ per 100 km are enclosed by this contour (figure 2.6d).

2.5.5 Eddy-Induced Ekman Pumping in the Midlatitudes

To investigate the dependence of the structure and magnitude of eddy-induced Ekman pumping on wind direction, composite medians of observed total eddy-induced Ekman pumping were constructed from QuikSCAT-based wind stress curl fields over all midlatitude eddies in a coordinate system that was rotated to align the background wind direction at a polar angle of $0^\circ$. The average upwelling velocities in the interior of anticyclones exceed $2.5 \text{ cm day}^{-1}$ with flanges of weak downwelling oriented perpendicular to the wind at the flanges of the eddies (figure 2.7a). The inverse is observed in cyclonic eddies, for which maximum downwelling exceeds $3 \text{ cm day}^{-1}$ with flanges of weak upwelling aligned perpendicular to the wind (figure 2.7c). The asymmetry of maximum upwelling and downwelling in anticyclones and cyclones is consistent with a disproportionately large number of large-amplitude midlatitude eddies being cyclones (figure 2.1b).

The Ekman pumping from surface current effects computed from the SSH fields and QuikSCAT wind speeds and directions in the same rotated coordinate systems is shown in figures 2.7b and d. The structures of the current-induced Ekman pumping are nearly identical to those obtained for idealized Gaussian eddies in figure 2.5a. Moreover, the structures of the current-induced Ekman pumping in figures 2.7b and d are also nearly identical to those in figure 2.7a and b that were computed from QuikSCAT wind stress curl fields, which include both SST and surface current-induced Ekman pumping. This clearly indicates that Ekman pumping is predominantly controlled by surface currents in midlatitude eddies. The very slight counterclockwise rotation of the major axes of the Ekman pumping in the cores of eddies and the side lobes of downwelling in anticyclones
and upwelling in cyclones are likely the result of relatively subtle SST effects on eddy-induced Ekman pumping.

While surface current effects usually dominated the total Ekman pumping in mesoscale eddies, the influence of SST becomes apparent regionally in the next section.

2.6. Regional variability of eddy-induced Ekman Pumping

The maps of the average magnitudes of eddy surface current-induced Ekman pumping and the pumping from air-sea interaction associated with eddy-induced SST anomalies in figure 2.6 show that the dominant mechanisms vary regionally. In this section the eddy-induced Ekman pumping is investigated in detail in eight diverse regions of the World Ocean. The trajectories of eddies in these eight regions are shown in figure 2.8a. The Agulhas Return Current, South East Atlantic and Kuroshio Extension regions were chosen because of a historical precedent for studies of air-sea interaction over fronts in these regions. The Caribbean Sea was chosen because it is a region of strong current-induced Ekman pumping but weak background SST gradients (figures 2.6a and d). The Central Pacific around the Hawaiian Islands, the South Pacific and the South Indian Ocean where chosen because of the potential role that eddy-induced Ekman pumping can have on ecosystems trapped within the eddy cores in these low nutrient, or oligotrophic, regions.

2.6.1 The Southern Indian Ocean

The Leeuwin Current that flows poleward along the western coast of Australia generates large-amplitude eddies that propagate westward with a slight meridional deflection (poleward for cyclones, equatorward for anticyclones (Morrow et al., 2004)). Many of the eddies in the south Indian Ocean (SIO) originate in the Leeuwin Current and can be tracked as far west as 50°E in the merged SSH fields (figure 2.8a). The average
amplitudes of eddies in the SIO is 8.3 cm, which is the 73rd percentile of midlatitude eddies with life-times greater than or equal to 12 weeks. The average radial scale of SIO eddies is 92 km with anticyclones having slightly larger horizontal scales than cyclones and correspondingly slower rotational speeds (figure 2.8b). There were 355 cyclonic and 345 anticyclonic eddies tracked during the 7.5-year time period considered here (figure 2.8a). There were a total of 8,773 and 9,804 weekly eddy realizations of anticyclones and cyclones, respectively.

Composite medians of Ekman pumping computed from QuikSCAT winds in the cores of SIO eddies exceed 6 cm day$^{-1}$ (figure 2.9i and j). Whereas the core of the eddy-induced Ekman upwelling in anticyclones is centered over the eddy centroid (figure 2.9i), the core of the downwelling in cyclones is displaced about $L_S/4$ (roughly 25 km) to the east (figure 2.9j). The same offset of approximately $L_S/4$ is found in the geostrophic current-induced Ekman pumping computed from altimetric SSH fields (figure 2.9c). The offset of the SSH contours and composite geostrophic vorticity visible in figure 2.9d indicates that the composite Ekman pumping inferred from QuikSCAT data should also be offset (figure 2.9j). The geographic structure of the eddy-induced SST anomalies (figure 2.9g) further shifts the core of total Ekman downwelling slightly to the northeast in cyclones.

The average wind direction in the SIO region is from the southeast. The degree of persistence of the wind direction can be characterized by the directional steadiness, defined to be the ratio of the magnitude of the vector averaged weekly wind observations to the scalar averaged weekly wind speed. The directional steadiness of the winds in the SIO is nearly 1 over most of the domain (figure 2.10). In the far southwestern reaches of the domain, however, the directional steadiness decreases to as low as 0.4. In addition, the direction of the winds shift from southerly in the eastern part of the domain to nearly easterly in the western portion of the region. These changes in the wind direction result in contours of composite medians of Ekman pumping that are nearly circular in the interiors
of the eddies (figures 2.9i and j) rather than elongated as would be the case in a region with high directional steadiness (figure 2.5a).

The geographic structure of the observed Ekman pumping velocities appears to be dominated by the current-induced Ekman pumping and not the SST-induced Ekman pumping, as evident from comparing figures 2.9b and f for anticyclones and figures 2.9c and g for cyclones. This is consistent with the dominance of surface current-induced Ekman pumping in the SIO region (figure 2.6c).

2.6.2 The Hawaiian Ridge

The Hawaiian Ridge (HAW) is a region of active eddy variability with a local maxima in eddy amplitude in the central North Pacific (Chelton et al., 2011b). Eddies of the HAW have an average amplitude of 6.7 cm, which is the 63rd percentile of midlatitude long-lived eddies. The average spatial scales of these eddies is 112 km which is the 76th percentile of midlatitude long-lived eddies. There were 263 cyclones and 259 anticyclones consisting of 4,785 and 5,278 individual weekly eddy realizations, respectively during the 7.5-year period considered here. The trajectories of these eddies are shown in figure 2.8a.

The winds in the HAW region are steady and nearly easterly (figure 2.10), which result in Ekman pumping anomalies in the interiors of HAW eddies that are elongated in the direction of the wind with weak side lobes of upwelling (downwelling) at the northern and southern peripheries of cyclones (anticyclones). The composite median Ekman upwelling exceeds 6 cm day$^{-1}$ in the cores of anticyclones while downwelling in cyclones excess of 5 cm day$^{-1}$ (figures 2.11i and j, respectively). The altimetrically determined geostrophic currents associated with HAW eddies (figure 2.11a and d) generate surface current-induced Ekman pumping with structures very similar to the Ekman pumping inferred from QuikSCAT winds (compare figures 2.11b-c and i-j). Note in particular the southward offset of the cores of downwelling relative to the eddy centroids in cyclones (figures 2.11c and j).
In anticyclones, SST anomalies create very little Ekman upwelling ($\approx 0.2 \text{ cm d}^{-1}$, figure 2.11f). The nature of the SST anomalies of these anticyclones is such that the Ekman upwelling from both SST and surface currents (figure 2.11b and f) are spatially similar and reinforce each other. In HAW cyclones, the largest SST-induced downwelling velocities are displaced south of the surface current-induced Ekman downwelling maxima (figure 2.11g and c), which shifts the core of maximum total downwelling further to the south of the eddy centroid. Note, however, that the SST-induced Ekman pumping is much weaker than the current-induced pumping. The observed Ekman pumping is therefore well represented by the geostrophic current-induced pumping (figures 2.11b and c).

2.6.3 The Southeast Atlantic

Eddies spawned in the Southeast Atlantic (SEA) are highly energetic with average amplitudes of 14.2 cm, which is the 89th percentile. The average radial scale of SEA eddies is 84 km, which is the 44th percentile. There were 161 cyclonic and 187 anticyclonic long-lived eddies consisting of of 2,512 and 2,976 eddy realizations, respectively, during the 7.5-year period considered here. Many of the anticyclonic SEA eddies propagate for long distances across the South Atlantic (Chelton et al., 2011b). As noted in section 2.4.2, however, in this study the eddies are considered only within the bounds of the study region as shown in figure 2.8a.

Eddy-induced Ekman pumping velocities exceed $-14 \text{ cm day}^{-1}$ in SEA cyclones and 13 cm day$^{-1}$ in SEA anticyclones (figures 2.12i and j). As with all anticyclones, maximum current-induced Ekman pumping occurs at the eddy centroid with velocities in excess of 11 cm day$^{-1}$ (figure 2.12b). The slightly smaller SEA cyclones have current-induced downwelling velocities in excess of 8 cm day$^{-1}$.

In contrast to the SIO and HAW regions considered above, SEA eddies have large SST anomalies (figures 2.12e and h) which generate a much stronger SST-induced Ekman pumping compared with most other regions investigated here. SEA anticyclones have SST
anomalies in excess of 1 °C, which generate Ekman pumping velocities of 8 cm day$^{-1}$ which are comparable to 11 cm day$^{-1}$ Ekman pumping generated by surface currents (figures 2.12f and b). SEA cyclones have slightly smaller SST anomalies than SEA anticyclones with SST-induced Ekman pumping of only −5 cm day$^{-1}$ (figure 2.12g).

The combination of large-amplitude eddies (and hence strong rotational surface currents) and strong SST anomalies in SEA anticyclones generates observed Ekman pumping anomalies with a geographic structure that is influenced by both mechanisms. The SST anomalies of anticyclones generate a maximum SST-induced Ekman upwelling approximately 1.5 $L_s$ to the south of the eddy centroid (figure 2.12f). The observed Ekman upwelling (figure 2.12i) is maximum to the south of the eddy centroid, approximately half way between the maximum of SST-induced upwelling and the current-induced upwelling in figures 2.12f and b.

SEA cyclones also have observed Ekman pumping structures that are influenced by both mechanisms (figures 2.12c,g and j). However, the SST-induced Ekman downwelling combines constructively with the surface current-induced Ekman downwelling to create the observed eddy-induced Ekman downwelling with core displaced to the south and with side lobes of upwelling to the north and south in cyclones (figure 2.12j).

In contrast to the constructive superposition of SST and surface current-induced Ekman pumping in SEA cyclones, there is destructive superposition of the two mechanisms in anticyclones which explains why SEA anticyclones have eddy-induced Ekman upwelling velocities that are slightly smaller in magnitude than the downwelling in the SEA cyclones, despite their generally large sea surface height amplitudes.

2.6.4 The Caribbean Sea

The eddy field of the Caribbean Sea (CAR) is characterized by mesoscale eddies with average amplitudes of 7.8 cm which is the 70th percentile. The average radial scale of the CAR eddies is 126 km which is the 86th percentile. There were 131 cyclonic and
11 anticyclonic long-lived eddies consisting of 11,700 and 1,439 realizations, respectively, during the 7.5-year study period.

The Ekman pumping signatures of CAR eddies observed in the composites of QuikSCAT wind stress curl (figures 2.13i and j) are very similar to the Ekman pumping of the geostrophic surface currents (figures 2.13b and c). The Ekman pumping velocities are on average $9 \text{ cm d}^{-1}$ in anticyclones and $-6 \text{ cm d}^{-1}$ in cyclones. The geographic structure of the Ekman pumping is nearly axisymmetric, centered slightly to the south of the eddy centroid in cyclones (figures 2.13j and c) and slightly to the north in anticyclones (figures 2.13i and b).

In contrast to other regions consider here, the SST-induced Ekman pumping velocities (figures 2.13f and g) are negligible, more than 2 orders of magnitude smaller than the surface current-induced Ekman pumping.

2.6.5 The South Pacific

The eddies in the central South Pacific (SP) have average amplitude of 4.3 cm, which is the 40th percentile. The average radial eddy scales are 96 km (59th percentile). There were 11,196 anticyclonic and 8,156 cyclonic long-lived eddies consisting of 29,673 and 27,015 eddy realizations, respectively, during the 7.5-year period considered here. Although these-small amplitude eddies generate relatively small Ekman pumping velocities when compared to the large-amplitude eddies found in the other regions, this region is one of the oligotrophic subtropical gyres with extremely low primary production (Behrenfeld et al., 2005) which represent a large portion of the midlatitude World Ocean.

The SST gradient magnitudes are also small in the SP (figure 2.6d). Effectively all of the eddy-induced Ekman pumping in these eddies is thus generated by eddy surface currents, as shown in figure 2.14. The average Ekman pumping velocities of SP eddies are only $2 \text{ cm d}^{-1}$ because of the relatively small amplitudes of the eddies (figures 2.14 i and j). The eddy surface currents generate Ekman pumping velocities of nearly $2 \text{ cm day}^{-1}$.
(figures 2.14b and c). The SST-induced Ekman pumping slightly reinforces the geostrophic current-induced Ekman pumping (figures 2.14 f and g).

While the observed Ekman pumping velocities (figures 2.14 i and j) are equal in magnitude to the Ekman pumping from current effects; their structures differ somewhat from those expected from surface current effects alone. We are not able to explain this discrepancy. It could be because of the small signal-to-noise ratio of the QuikSCAT measurements of surface winds because of the weak surface velocities of the eddies.

2.6.6 The North Side of the Kuroshio Extension

The Kuroshio Extension in the western North Pacific can be partitioned into two regions in close proximity where different mechanisms dominate the eddy-induced Ekman pumping. The very strong SST gradient on the north side of the Kuroshio Extension (KEN) (figure 2.6b) which are expected to have a strong influence on the geographic structure of eddy-induced Ekman pumping because of the SST-induced wind stress curl. In contrast, the south side of the Kuroshio Extension (KES) is characterized by weaker SST gradients so that current-induced Ekman pumping velocities are expected to exert a stronger influence on Ekman pumping.

Eddies in the KEN have average amplitudes of 11 cm which is the 82nd percentile. Anticyclones make up a larger portion of large-amplitude eddies (figure 2.8g). The average radial scale of the eddies in the KEN are 89 km which is the 50th percentile. There were 232 cyclonic and 280 anticyclonic long-lived eddies consisting of 5,837 and 7,104 realizations, respectively over the 7.5-year period considered here.

The surface current-induced Ekman pumping of anticyclones in the KEN is centered on the eddy centroid with upwelling in excess of 3.5 cm day$^{-1}$ (figure 2.15b). The SSH signatures of cyclones in the KEN are shifted slightly to the south of the eddy centroid (figure 2.15d) resulting in current-induced Ekman downwelling velocities of 3 cm day$^{-1}$ that are also centered to the south of the eddy centroid (figure 2.15c). This asymmetry
between maximum upwelling in anticyclones and downwelling in cyclones is a result of a disproportionately larger number of large eddies being anticyclonic along this northern flank of the Kuroshio Extension (figure 2.8g).

The composite SST anomalies for KEN eddies exceed 0.6 °C in both cyclones and anticyclones (figures 2.15e and h). These large SST anomalies are a result of the strong and persistent SST gradient in the KEN region (figure 2.6b). The associated SST-induced Ekman pumping velocities approach 1 cm day^{-1} (figures 2.15f and g). The observed eddy-induced Ekman upwelling in cyclones has a geographic structure that results from the superposition of SST and surface current effects on Ekman pumping. Maximum Ekman upwelling is shifted to the north of the eddy centroid with a secondary region of strong Ekman downwelling centered in the southwestern quadrant of the eddy composites (figure 2.15i).

In cyclones, maximum eddy-induced Ekman downwelling is shifted to the north of the eddy centroid because of SST effects (figure 2.15j). SST-induced Ekman pumping is apparently also responsible for the secondary region of Ekman upwelling located to the south of the centroids of cyclones (figure 2.15j).

2.6.7 The South Side of the Kuroshio Extension

Eddies on the KES have average amplitudes of 13.8 cm, which is the 88th percentile. A disproportionately large number of the large-amplitude eddies (≥ 20 cm) are cyclonic on the southern flank of the Kuroshio Extension. The average amplitudes KES eddies are 15.5 cm and 10.7 cm for cyclones and anticyclones, respectively. The average spatial scale of the KES eddies is 102 km, which is the 66th percentile. There were 284 cyclonic and 245 anticyclonic long-lived eddies consisting of 6,271 and 5,485 realizations, respectively over the 7.5-year study period considered here.

The surface current-induced Ekman pumping of KES eddies (figures 2.16b and c) is generally similar to that in the KEN eddies, with the exception of larger downwelling
velocities in KES cyclones because of their larger amplitudes (figures 2.16d and 2.15d).

The SST anomalies within the interiors of KES eddies (figures 2.16f and g) are much smaller than in the KEN, resulting in negligibly small SST-induced Ekman pumping velocities of less than 0.1 cm day$^{-1}$ (figures 2.16f and g). The observed eddy-induced Ekman pumping (figures 2.16i and j) is generally similar to the surface current contributions (figures 2.16b and c). However there are some differences between the pumping observed by QuikSCAT and that due to only surface current effects. There appears to be a shift in the alignment of the major axis of Ekman upwelling in the interiors of anticyclones, when compared to the upwelling from currents alone (figure 2.16b and i). We are not able to explain these features, but they are likely attributable to the fact that some the mesoscale features in this regions are actually meanders rather than isolated eddies. In addition, there are flanges of Ekman pumping of opposite signs from the Ekman pumping in the eddy cores that are difficult to explain by the small SST-induced Ekman pumping.

### 2.6.8 The Agulhas Return Current

The mesoscale eddies in the Agulhas Return Current (ARC) have large average amplitudes of 12.4 cm which is the 85th percentile. The average scale is 83 km which is the 43rd percentile. The SST gradients are strong in the ARC (figure 2.6d) and are therefore expected to have eddy-induced Ekman pumping signatures that are strongly influenced by the SST-induced wind stress curl. The winds over the ARC are predominately westerly and exceed 10 m s$^{-1}$ (figure 2.10). There were 655 anticyclone and 755 cyclones consisting of 15,169 and 16,981 realizations, respectively over the 7.5-year time period considered here.

The SST-induced Ekman pumping in the ARC region consists of a dipole with somewhat stronger upwelling (downwelling) on the south sides of the cores of anticyclones (cyclones) (figures 2.17f and g). The surface current-induced Ekman pumping in ARC eddies has the classic structure, with upwelling (downwelling) centered on the cores of
anticyclones (cyclones) (figures 2.17b and c). The zonal elongation of these features with secondary sidelobes of opposing Ekman pumping to the north and south are very similar in structure to the Ekman pumping from an idealized Gaussian eddy shown in figure 2.5a. As a result of the strong SST gradients associated with the large zonally aligned temperature fronts of the ARC, Ekman pumping from SST anomalies has a major influence on the total observed Ekman pumping in figures 2.17i and j. The maximum total Ekman upwelling in ARC anticyclones (figure 2.17i) is displaced to the south of the maximum from surface current effects (figure 2.17g), but to the north of the maximum from SST effects (figure 2.17if). Moreover, the stronger pole of downwelling in figure 2.17f is displaced to the north in figure 2.17i. Comparable displacements are evident from figures 2.17c, g and j for cyclones.

2.7. Conclusions

The investigation of regional variability of the relative influences of SST and surface current effects on eddy-induced Ekman pumping in section 2.5. revealed that SST-induced Ekman pumping is usually secondary to current-induced pumping. In regions away from strong SST gradients, SST effects on wind stress curl act to displace the regions of maximum eddy-induced Ekman pumping somewhat from the eddy centroid. In regions of strong SST gradients, which occur in meandering zonal-currents such as the Agulhas Return Current and the Kuroshio Extension, the opposite occurs; eddy surface currents perturb the spatial structure of SST-induced Ekman pumping within eddies.

Through isolating eddies formed in different regions of the World Ocean, we observed differences in the relative influence of SST and surface current driven wind stress curl perturbation in creating the observed Ekman pumping anomalies within the interiors of eddies. In general, eddy surface currents sustain upwelling in the interiors of anticy-
clonic eddies and downwelling in the cores of cyclonic eddies through their effect on the surface stress. The magnitude of this upwelling depends on the amplitude and associated rotating geostrophic current velocity in the mesoscale eddies and on the magnitude of the background wind field. In most regions, the modification of current-induced Ekman pumping by SST-induced wind stress curl is secondary. In some regions, however, the observed eddy-induced Ekman pumping field most closely resembles the SST-induced portion of the Ekman pumping field. This is most clearly apparent in the ARC, which we attribute to the strong spatial SST gradients and the strong and steady winds observed in the ARC region.

Through the collocation of relative wind vectors, surface wind stress curl and SST to the eddy interiors defined by SSH, we have shown that the Ekman pumping observed in the interior of mesoscale eddies results from a combination of the influence of eddy surface currents on the surface stress and the influence of SST on the surface stress field. Isolation of the Ekman pumping from the two mechanisms shows that their relative impact on the total eddy-induced Ekman pumping field varies regionally.

The vorticity of surface currents in mesoscale eddies generate upwelling in the cores of anticyclones and downwelling in the cores of cyclones. These vertical motions have been estimated here by combining the observed geostrophic surface currents from the SSH anomalies of eddies with background wind fields obtained from QuikSCAT as described in appendix A. Eddy-induced SST anomalies also act to generate Ekman pumping through air-sea interaction in which SST modifies the overlying wind field. In regions of strong SST gradients, eddy-induced SST anomalies can exceed 0.5 °C, with instantaneous anomalies often exceeding 1 °C. The geographical structure of eddy-induced SST anomalies generate a curl of the wind stress that in turn generates Ekman pumping. This SST-induced Ekman pumping is a function of both the geographical structure of the SST signatures of the eddies and the wind direction. Current-induced Ekman pumping always generates upwelling in
the cores of anticyclones and downwelling in the cores of cyclones. The superposition of SST-induced Ekman pumping can augment the current-induced Ekman pumping either constructively or destructively, depending on the magnitudes of the SST anomalies and magnitude and direction of the ambient wind. The magnitude of the SST contribution to Ekman pumping is thus highest in regions of strong SST gradients (figure 2.6c and d). Because the SST-induced wind perturbations are non-negligible in these regions, their contribution to Ekman pumping must be included in modeling studies to adequately resolve the influence of eddies and fronts on biogeochemical cycling.

The \( \sim 30 \text{ cm day}^{-1} \) eddy-induced Ekman pumping velocities in large eddies is an order 1 perturbation of the background, basin-scale Ekman pumping velocities from the large-scale wind fields (Risien and Chelton, 2008). The importance of eddy-induced Ekman pumping to near-surface biogeochemical cycling has been noted from field observations (Martin and Richards, 2001; McGillicuddy et al., 2007) and coupled bio/physical ocean models (Eden et al., 2009; Anderson et al., 2011). While the typical Ekman pumping velocities of \( < 10 \text{ cm day}^{-1} \) are smaller in most regions than the average basin-scale Ekman pumping, they are non-negligible because of their persistence over the long lifetimes of the mesoscale eddies. Since nearly all of the observed mesoscale eddies outside of the tropics are nonlinear, with rotational velocities exceeding the rate of propagation of these features, fluid is trapped within the cores of eddies (Chelton et al., 2011b; Early et al., 2011). The persistent eddy-induced Ekman upwelling in anticyclones can act to inject new nutrients into the cores of trapped fluid in the eddies, thus stimulating new production that can modulate open-ocean carbon cycling. Eddy-Induced Ekman downwelling in the cores of cyclonic eddies can continually transport phytoplankton downward and thus out of the euphotic zone, therefore suppressing primary production. Further investigation of the impact of eddy-induced Ekman pumping on marine ecosystems trapped within eddy cores is required in order to assess the importance of these vertical motions to global
primary production.

Modeling efforts are needed to elucidate the complex nature of the interactions between Ekman pumping and nutrient cycling in the interior eddies. These studies must include the corrections for the influence of both currents and SST anomalies on the wind stress curl field. The portion of the eddy-induced Ekman pumping from surface current effect can be easily included in ocean circulation models by computing the surface stress from the relative vector wind defined by (1.1) that includes the influence of surface ocean currents. This has been incorporated in several recent models (e.g. Eden et al., 2009; Mc-Clean et al., 2010; Anderson et al., 2011). Parameterizing the effects of air-sea interaction on the wind stress curl field to include the well-documented impact of SST perturbation on the wind field is more complicated. This has been shown to be important from an idealized coupled model by (Jin et al., 2009). By including the influence of both surface eddy currents and eddy-induced SST perturbation on the wind stress curl, future modeling studies could determine where the attenuation from eddy-induced Ekman pumping plays an important role in eddy dynamics and in providing nutrients into the euphotic zone of the interiors of mesoscale eddies.
FIGURE 2.1: (a) Trajectories of midlatitude eddies between 15° and 45° latitude in both hemispheres. Red and blue tracks correspond to cyclones and anticyclones, respectively. (b) Cumulative probability function (top) and histogram (middle) of eddy amplitudes, cyclones in blue and anticyclones in red, and the ratio of cyclonic to anticyclonic histograms (bottom). (c) Same as (b) but for the eddy radial scale $L_S$. (c) Same as (a) except for the rotational speed $U$ of the eddy.
FIGURE 2.2: Schematic diagram of eddy-driven horizontal advection of SST for clockwise and counterclockwise-rotating eddies (top and bottom, respectively) propagating westward in regions where the SST gradient is (a) southward and (b) southeastward. An otherwise smooth contour of SST (dashed lines) is distorted by the rotational velocity field within the eddy, as shown by the solid lines. Advection of SST within the large-scale background SST gradient results in the positive and negative SST anomalies shown by the red and blue regions, respectively. The dependence of the locations of these SST anomalies on the direction of the large-scale background SST gradient that is evident from comparison of (a) and (b) was accounted for by compositing eddy observations in a coordinate system rotated for each eddy (see section 2.4.2).
FIGURE 2.3: Composite medians of SST (a-d) and wind speed (e-h) anomalies in mid-latitude eddies computed from the Reynolds OIv2 SST analyses and QuikSCAT wind speeds. Eddies are segregated according to the meridional direction of the background SST gradient (either northward or southward) and the rotational sense of the eddies (either clockwise or counterclockwise). Composite medians are rotated to align the background SST gradient to either a polar angle of ±90° (see section 2.4.2). The magnitude of the asymmetry between the primary and secondary pole of the anomalies is labeled as the value \( r \) in each panel. The x and y coordinates of the composite medians are normalized by the eddy scale \( L_a \).
FIGURE 2.4: The coupling coefficient calculated from the observations of SST and wind speed anomalies within radii $L_s$ of the centroids of midlatitude eddies. Each point is the mean of the wind speed anomalies bin averaged by the SST anomalies. 2-D histogram of the distribution of filtered SST and wind speed. Contour interval is 10% starting at 1%. The innermost contour thus corresponds to 71%. Linear fit estimated from binned least squares regression (dashed line) to observations that fell within the 11% contour of the 2-d histogram is shown by the dashed line. The 95% confidence interval of each bin averaged is calculated using the standard formula $\pm \sigma_W q_{t}(0.025, N^* - 1)/\sqrt{N^*}$, where $\sigma_W$ is the standard deviation of the wind speed estimates in each bin, $N^*$ is a conservative estimate of the effective number of degrees of freedom and $q_{t}(0.025, N^* - 1)$ is the 2.5% percentage point of the Student’s t random variable with $N^* - 1$ degrees of freedom. We estimated $N^*$ as the number of seasons of data (i.e., four times the number of years), which we believe is a very conservative lower bound of the actual value. The value of the slope of the line fit by least squares regression is labeled on the figure.
FIGURE 2.5: Ekman pumping velocities from surface current and SST effects for a series of idealized anticyclonic Gaussian eddies with e-folding scales of 100 km at 30°N propagating through a southward SST gradient in a uniform 7 m s\(^{-1}\) zonal wind. The x and y axes are normalized to the eddy e-folding scale. The schematics are representative of three eddy SSH anomalies. Column 2 corresponds to large eddies with an amplitude of 20 cm, and a SST anomaly of 1.3 °C, which is the 95th percentile of midlatitude eddies. Column 3 corresponds to medium sized eddies with an amplitude of 10 cm and an SST anomaly of 0.6 °C which is the 75th percentile. Column 4 corresponds to small eddies with an amplitudes of 5 cm and an SST anomaly of 0.25 °C, representing the 25th percentile. Row (a), first column, geostrophic vorticity overlaid with contours of SSH (contour interval 2 cm) for the large eddy. The three columns to the right of the eddy geostrophic vorticity schematic are the Ekman pumping resulting from the geostrophic surface currents of eddies of 3 different sizes. Row (b), first column, SST anomalies. The panels to the right of the SST schematic are the Ekman pumping velocities resulting from the SST-induced wind stress perturbations for SST anomalies of three different magnitudes. Row (c), same as row (b) but for a southerly background wind.
FIGURE 2.6: Global $1/4^\circ \times 1/4^\circ$ maps of averages over the 7.5-year data record: (a) The magnitude of current-induced Ekman pumping and (b) the magnitude of SST-induced Ekman pumping. (c) Median Ratio of current-induced to SST-induced Ekman pumping, shown on a $\log_{10}$ scale. (d) The magnitude of the SST gradient. The contours overlaid in both panels c and d correspond to the ratio $W_{SST}/W_{\nabla \times u_g} = 1.$
FIGURE 2.7: Composite medians of midlatitude eddy Ekman pumping rotated to orient the wind vector towards the polar angle of 0° for anticyclones (left) and cyclones (right): (a) Computed from QuikSCAT winds; and (b) computed from geostrophic surface velocities determined from the altimetric SSH fields (see appendix A for details). Contour interval is 0.5 cm day$^{-1}$ and the number of eddy realizations used in each of the composites is labeled as $N$ in each panel title. The x and y coordinates of the composite medians are normalized by the eddy scale $L_s$. 
FIGURE 2.8: (a) The trajectories of cyclonic (blue) and anticyclonic (red) mesoscale eddies originating in each of the study regions. Histograms of eddy amplitude (upper), scale $L_S$ (middle) and eddy rotational speed $U$ (lower) for eddies of the: (b) South Indian Ocean; (c) Hawaiian Ridge; (d) Southeastern Atlantic; (e) Caribbean Sea; (f) South Pacific; (g) north flank of the Kuroshio Extension; (h) south flank of the Kuroshio Extension and (i) the Agulhas Return Current.
FIGURE 2.9: Composite medians of eddies originating in the south Indian Ocean (SIO), defined as $20^\circ S - 35^\circ S$ and $80^\circ E - 120^\circ E$. The x and y coordinates of the composite medians are normalized by the eddy scale as defined in the text. Composite median geostrophic vorticity (color) overlaid with contours of SSH for anticyclones (a) and cyclones (d). Composites of Ekman pumping from the geostrophic surface currents for anticyclones (b) and cyclones (c). Composites of SST anomalies overlaid with contours of SSH anomalies for anticyclones (e) and cyclones (h). Composites of SST-induced Ekman pumping for anticyclones (f) and cyclones (g). Observed Ekman pumping anomalies for anticyclones (i) and cyclones (j). Note the different colorbar scales for panels b-c, i-j and f-g.
FIGURE 2.10: Directional steadiness of the wind, defined to be the magnitude of the vector-average wind divided by the scalar average wind speed, overlaid with vectors of the mean wind, both computed from the QuikSCAT data. Vectors are plotted on a $3.75^\circ$ latitude by $6.25^\circ$ longitude grid.
Hawaii

FIGURE 2.11: Same as figure 6, except for the Hawaiian Ridge region defined as a box with vertices at 15°N – 25°N and 180°E – 220°E. Note the different colorbar scales for panels b-c, i-j and f-g.
South East Atlantic

FIGURE 2.12: Same as figure 6, except for the South East Atlantic region defined as $45^\circ S - 35^\circ S$ and $0^\circ E - 20^\circ E$. Note the different colorbar scales for panels b-c, i-j and f-g.
FIGURE 2.13: Same as figure 6, except for the Caribbean region, defined as $10^\circ N - 20^\circ N$ and $275^\circ E - 310^\circ E$. Note the different colorbar scales for panels b-c, i-j and f-g.
South Pacific

FIGURE 2.14: Same as figure 6, except for the South Pacific region, defined as $15^\circ S - 45^\circ S$ and $190^\circ E - 270^\circ E$. Note the different colorbar scales for panels b-c, i-j and f-g.
Kuroshio North

FIGURE 2.15: Same as figure 6, except for the north side of the Kuroshio, defined as $35^\circ N - 45^\circ N$ and $140^\circ E - 180^\circ E$. Note the different colorbar scales for panels b-c, i-j and f-g.
Kuroshio South

FIGURE 2.16: Same as figure 6, except for the south side of the Kuroshio, defined as 25° N – 35° N and 140° E – 180° E. Note the different colorbar scales for panels b-c, i-j and f-g.
FIGURE 2.17: Same as figure 6, except for the Agulhas Return Current region, defined as $40^\circ S - 50^\circ S$ and $20^\circ E - 120^\circ E$. Note the different colorbar scales for panels b-c, i-j and f-g.
Phytoplankton Response to Eddy-Induced Ekman Pumping in the South Indian Ocean

Peter Gaube, Dudley B. Chelton, Pete G. Strutton and Michael J Behrenfeld
3. PHYTOPLANKTON RESPONSE TO EDDY-INDUCED EKMAN PUMPING IN THE SOUTH INDIAN OCEAN

3.1. Abstract

Nonlinear mesoscale eddies can influence biogeochemical cycles in the upper ocean through vertical and horizontal advection within the cores of the eddies. The relative importance of these two processes depends on the polarity of an eddy (cyclones versus anticyclones) and the initial biological conditions of the fluid trapped in the core of the eddy at the time of formation. Eddies originating in the South Indian Ocean are unique in that anticyclones contain elevated levels of chlorophyll-a, enhanced primary production and phytoplankton communities generally associated with nutrient-replete environments. From analysis of 8 years of satellite measurements of sea surface height, chlorophyll, phytoplankton carbon and surface winds, we present evidence that the eddy-induced Ekman upwelling in these anticyclones is sufficient to sustain the enhanced phytoplankton anomalies observed within the eddy cores. The primary mechanism for this eddy-induced Ekman upwelling is the curl of the surface stress resulting from the difference between the surface vector winds and the rotating surface currents within interiors of the anticyclones. The biological response to this eddy-induced Ekman upwelling is evident only during the Austral winter. This seasonal dependence appears to be attributable to wintertime deepening of the mixed layer that facilitates the injection of nutrients from the deep water into the euphotic zone. In contrast to anticyclones, the chlorophyll-a content of cyclones is anomalously low throughout the year. This is consistent with the persistent surface current-induced downwelling that occurs within cyclones.
3.2. Introduction

Mesoscale eddies (spatial scales of order $10 - 100 \ km$) are ubiquitous features of the World Ocean (Chelton et al., 2011b). Eddies are known to play an important role in the meridional transfer of heat and salt (Qiu and Chen, 2005; Roemmich and Gilson, 2001; Chaigneau et al., 2011) and in the nutrient enrichment of oligotrophic pelagic ecosystems (McGillicuddy and Robinson, 1997; Oschlies, 2002; McGillicuddy et al., 2007). There are three primary mechanisms by which mesoscale eddies can influence marine ecosystems: 1) Eddy pumping drives upwelling during the formation of cyclonic eddies, resulting in the injection of new nutrients into the euphotic zone in the cores of cyclones. In anticyclones, eddy pumping generates downwelling during eddy formation acting to transport organisms and nutrient out of the euphotic zone in interiors of anticyclones. 2) The rotational velocities of eddies advect ecosystems around their peripheries resulting in the meridional displacement of isocontours of biological variables such as chlorophyll (Chelton et al., 2011a). 3) Eddy-induced Ekman pumping arising from eddy surface currents and/or air-sea interaction generates upwelling in the interiors of anticyclones and downwelling in the interiors of cyclones throughout the lifetime of an eddy (see chapter 2). The relative importance of the various eddy influences on marine ecosystems varies geographically (see chapter 4).

The goal of this study is to investigate the eddy influence on chlorophyll concentrations ($CHL$) in the cores of South Indian Ocean (SIO) mesoscale eddies. The SIO is singled out because of the uniqueness of the eddies in this region. In particular, anticyclones that form off the west coast of Australia initially have very high CHL content. It will be shown that this predisposes the $CHL$ in anticyclones to be influenced primarily by eddy self-induced Ekman pumping.

The primary emphasis in this study is on $CHL$ estimated from satellite measure-
ments of ocean color (Garver and Siegel, 1997; Maritorena et al., 2002; Siegel et al., 2002). Although \textit{CHL} is representative of phytoplankton biomass to first order, it is known that \textit{CHL} alone is an incomplete metric for the assessment of biomass and primary production. We therefore also investigate the effects of eddies on phytoplankton carbon concentration \((C_{phyto})\), a direct estimate of phytoplankton biomass that can also be measured from satellites (Behrenfeld et al., 2005), as well as anomalies of the the ratio \(r_C\) of \textit{CHL} to \(C_{phyto}\). The ratio \(r_C\) is a function of multiple factors including the growth rate of phytoplankton and the light field to which phytoplankton have acclimated (see section 1.2.3). For this study, we interpret \(r_C\) as a measure of phytoplankton growth rates, which is justified by investigating the effect of the ambient light field on the observed \(r_C\) ratios and directly estimating the phytoplankton growth rate \((\mu)\) based on a series of individual Argo float profiles that occurred in the interiors of SIO eddies.

Recent investigations of the physical characteristics of mesoscale eddies reveal that nearly all of the eddies resolved by the merged sea-surface height (SSH) fields are non-linear (Chelton et al., 2011b). The SIO that is the focus of this study is a region of large eddy kinetic energy (Killworth et al., 2004; Feng et al., 2005; Chelton et al., 2011b). Waite et al. (2007) showed that large-amplitude anticyclonic eddies in the SIO are associated with anomalously high primary production, when compared to cyclones and the regions surrounding eddies. Fortuitously, this is also a region with favorable clear-sky conditions, allowing relatively good spatial and temporal coverage of satellite ocean color measurements.

The SIO is also a unique region biologically. The eastern boundary current in the SIO is the Leeuwin Current, which flows poleward and thus differs from the eastern boundary currents in other ocean basins. Anticyclonic eddies formed from the pinching off of meanders of poleward-flowing boundary currents preferentially entrain continental shelf water which seeds the interiors of anticyclones with productive (referred to as eutrophic)
ecosystems (Pearce and Griffiths, 1991). On the other hand, cyclones that form in the poleward-flowing Leeuwin current entrain water from offshore during formation, resulting in trapped ecosystems that are low in nutrients and phytoplankton (referred to as oligotrophic). Observations confirm this distinction between anticyclones and cyclones in the eastern SIO (Thompson et al., 2007; Waite et al., 2007).

Another distinction between anticyclones and cyclones in the SIO is that mixing (as quantified by measurements of different phytoplankton characteristics that are known to acclimate to local light fields at different rates) is up to 20 times faster and deeper in anticyclones compared with cyclones (Thompson et al., 2007). Moreover, these observations show that vertical mixing is greater in the cores of anticyclones than outside of the vortex (Thompson et al., 2007), suggesting that some physical mechanism is acting within the cores of the anticyclones to enhance mixing by modulating the vertical transport of phytoplankton and nutrients.

A cyclonic Leeuwin Current eddy surveyed in 2003 was observed to be surface-convergent, as evidenced by the fact that a drifter released away from the eddy center was retrieved two weeks later in the exact center of the cyclone (Waite et al., 2007). The accumulation of surface drifters in the interiors of Leeuwin Current cyclones is a common occurrence (Griffin et al., 2001). The observed surface convergence is expected from the effects of eddy-induced Ekman downwelling. The observations of surface convergence, and hence downwelling, suggest that CHL should be reduced within the euphotic zone, as was observed by Waite et al. (2007).

The few ship-based observations of Leeuwin Current eddies provide a context from which satellite observations of phytoplankton communities in thousands of SIO eddies can be interpreted. We will show that the elevated levels of CHL observed in the interiors of anticyclones that form in the Leeuwin Current persist throughout the lifetimes of the eddies. For these phytoplankton communities to be sustained at concentrations above the
background levels, there must be a replenishment of nutrients within the eddy interiors. Previous studies of eddies in the North Atlantic Ocean by Martin and Richards (2001) and McGillicuddy et al. (2007, 2008) suggest a mechanism for the necessary nutrient supply. As originally pointed out by Stern (1965) and Dewar and Flierl (1987), the surface currents in rotating eddies generate a curl of the surface stress with polarity opposite that of the vorticity of the eddy. Anticyclones thus generate cyclonic curl of the surface stress and hence upwelling through Ekman pumping.

This surface current-induced Ekman pumping can upwell nutrients from below the euphotic zone in anticyclones. In the case of cyclones, the surface current-induced Ekman downwelling can advect organisms downward and out of the euphotic zone. Vertical velocities associated with surface current-induced Ekman pumping within mesoscale eddies can exceed 1 m day$^{-1}$ in large wind events (Martin and Richards, 2001), but are more typically $\sim 0.1m$ day$^{-1}$ (Gaube et al., 2012)

Surface current-induced Ekman upwelling has been verified from a detailed survey of an anomalous phytoplankton bloom in the core of a North Atlantic mode-water anticyclone (McGillicuddy et al., 2007; Ledwell et al., 2008). The vertical velocity calculated from the upward flux of a tracer released at a depth of $\sim 100$ m near the core of the anticyclones compared well with the eddy-induced Ekman upwelling velocity calculated from scatterometer-derived wind stress curl values.

The previous studies by Martin and Richards (2001) and McGillicuddy et al. (2007) postulated that surface current-induced upwelling is especially effective in mode-type anticyclonic eddies, in which the near-surface isopycnals are shoaled over a lens of water below which isopycnals are depressed. The shallowness of the isopycnals above the mode-water lens allows upwelling velocities to stimulate new production in the euphotic zone. Results presented here for hundreds of eddies in the SIO show that a biological response to surface current-induced upwelling in anticyclones is not limited to mode-type eddies. The shal-
lowness of isopycnals in mode-type eddies is evidently not critical for Ekman upwelling to sustain \(\text{CHL}\) in anticyclones.

Another mechanism for eddy-induced Ekman pumping arises from eddy perturbations of the SST field. SST influence on the overlying wind field has been studied extensively in SST frontal regions (see the reviews by Small et al. (2008) and Chelton and Xie (2010)). Park and Cornillon (2002) and Park et al. (2006) have shown from satellite measurements of winds and SST over Gulf Stream rings that this same air-sea interaction also occurs over mesoscale eddies in association with eddy-induced perturbations of the SST field. In a companion paper, we show that this SST influence on winds over eddies occurs globally (Gaube et al., 2012).

In the SIO region that is the focus of this study, it is shown in section 3.4. that the total eddy-induced Ekman pumping from the combined effects of surface currents and air-sea interaction over SST anomalies is typically 10 cm day\(^{-1}\). These velocities are small compared with coastal upwelling (Huyer, 1983) or submesoscale upwelling (Mahadevan et al., 2008) which are order 1 m day\(^{-1}\). However, eddy-induced Ekman pumping velocities are localized to the eddy core and are persistent, though with time-varying intensity, throughout the lifetime of an eddy. The integrated effects of the small but persistent eddy-induced Ekman upwelling in anticyclones can therefore significantly influence primary production in the closed ecosystem that are trapped in the eddy core. A prerequisite for this, however, is that the anticyclones must have high phytoplankton concentrations at the time of formation. As noted above, this is the case for SIO eddies.

In a recent modeling study of the North Atlantic Ocean, Eden et al. (2009) argued that the net impact of the injection of nutrients into the euphotic zone by eddy-induced Ekman pumping is minimal. When eddy-induced Ekman pumping was included in a coupled bio/physical model, they found that the integrated net primary production decreased by 5%. They also found that the eddy kinetic energy decreased by up to 50%. Because
the net effect of eddy-induced Ekman pumping is to attenuate eddies (Dewar and Flierl, 1987), this decrease in eddy kinetic energy comes as no surprise.

A more recent coupled bio-physical modeling study by Anderson et al. (2011) found that then inclusion of eddy-induced Ekman pumping has a significant effect on phytoplankton growth rates in the interiors of mesoscale eddies as well as on the phytoplankton community composition within the eddies. A particularly noteworthy result of this modeling study is that the biomass of the photoautotrophic community was higher in the cores of mode-type anticyclones. The biomass and growth rates of diatoms also increased in non-mode type anticyclones when surface current-induced Ekman pumping was included in the model (Anderson et al., 2011).

The influence of eddies on phytoplankton communities is investigated in this study by collocating satellite observations of eddy-induced Ekman pumping, $CHL$ and $C_{phyto}$ to the interiors of mesoscale eddies as identified and tracked from their sea-surface height (SSH) signatures. The data analyzed in this study are summarized in sections 1.2.1, 1.2.2 and 3.3 along with details of the filtering applied to each variable in order to isolate the eddy signals and collocate them to the eddy interiors. Eddy effects are investigated by analyzing composites constructed from hundreds to thousands of individual eddy realizations. The results are presented in section 3.4, beginning with an overview of SIO eddies and the background $CHL$ and $C_{phyto}$ fields and followed by a description of the effects of eddy-induced Ekman pumping on phytoplankton communities within SIO eddies. Section 3.4 also includes a description of the seasonal covariability between mixed layer depth and phytoplankton in the cores of the SIO eddies. A mechanism for an observed seasonal response of phytoplankton to eddy-induced Ekman upwelling is shown in section 3.5 along with an assessment of the ability of eddy-induced Ekman upwelling to sustain observed primary production rates in the core of a representative anticyclone. A summary of the results of this study and conclusion are presented in section 3.6.
3.3. Methods

An overview of the methods used to identify and track mesoscale eddies in altimetric SSH maps is given in section 1.2.1. In this study we analyze eddies tracked in SSH maps starting in September 1997 through September 2010, the time period covered by the SeaWiFS data record. A total of 10,191 anticyclones and 10,534 cyclones with lifetimes longer than 12 weeks were identified in the 19-year data record considered here.

Measurements of eddy-induced Ekman pumping were made for every eddy identified during the 10-year QuikSCAT record beginning in July 1999 and ending in November 2009. The processing of QuikSCAT data and filtering applied to isolated the mesoscale are described in section 1.2.2.

Near-surface chlorophyll (CHL) and phytoplankton carbon (Cphyto) were estimated from water leaving radiance measured by SeaWiFS as described in section 1.2.3.

3.3.1 Collocation of Bio-optical Observations with Eddy Interiors

This investigation of mesoscale biological-physical interactions is based on eddies with lifetimes of 12 weeks and longer that have been identified and tracked based on their signatures in sea surface height (SSH), as described in section 1.2.1. The satellite-based estimates of Ekman pumping, CHL, Cphyto and $r_C$ within the eddies were constructed by collocating the various satellite datasets to the interiors of each eddy realization. Collocated values within each eddy interior are referenced to the eddy centroid and interpolated onto a grid with distance from the centroid normalized by the eddy radius scale $L_a$. This normalization allows composites to be constructed from thousands of eddy observations on a common grid defined by the horizontal size of each individual eddy. Throughout the study we use composite medians rather than then the more common composite means because the median is less sensitive to occasional outliers of the anomaly fields.
3.3.2 Argo Float Profiles

Details of how mixed layer depth (MLD) as measured by Argo floats are collocated to eddy interiors is given in section 1.2.4. For the nearly 8-year overlap of the altimeter and Argo float data record, a total of 1593 and 1321 profiles were collocated to SIO cyclones and anticyclones, respectively (figure 3.1).

3.3.3 Phytoplankton growth rate $\mu$

For our investigation of the influence of SIO eddies on oceanic biology, we do not attempt to calculate the phytoplankton growth rate $\mu$ directly for each eddy. Instead, we analyze anomalies of the $r_C$ ratio to gain insight into the variability of phytoplankton biomass and growth rate and why these two quantities differ. We justify this approach by highlighting a series of case study eddies where we are able to estimate $\mu$ directly and show that the $r_C'$ is representative of $\mu$ within eddies. In addition, we also show that the effects of the acclimation of phytoplankton communities to the mixed layer light field are secondary to the effects of eddy-induced nutrient fluxes within eddies (see section 3.5).

Satellite-based estimates of $r_C$ have been used by Behrenfeld et al. (2005) to construct a maximum potential ratio $r_C^{\text{max}}$ as a function of mixed layer growth irradiance ($I_g$, the median light level in the mixed layer). At low light levels of $I_g \leq 0.5\ Ein\ m^2\ hr^{-1}$, the ratio $r_C$ increases because of photoacclimation, which can lead to increased CHL observed at the ocean surface, even without a change in the local nutrient field. To address the role of photoacclimation in observed eddy-induced CHL anomalies in chapter 3., we compared the observed mixed layer growth irradiance (which depends on MLD, irradiance and CHL) to $r_C^{\text{max}}$ derived by Behrenfeld et al. (2005). MLD was estimated from Argo float profiles and photosynthetically active radiation was obtained from SeaWiFS, as described in section 1.2.4.

Details of how $\mu$ is computed are given in section 1.2.5. In total, 1274 and 1036 MLD estimates were made for cyclones and anticyclones respectively during the 8 year
SeaWiFS data record analyzed here. In total, we made 628 and 684 direct estimates of \( \mu \) within the interiors of cyclonic and anticyclonic SIO eddies, respectively. The number of \( \mu \) estimates differs from the total number of Argo profiles (see section 1.2.4) because of missing CHL and \( C_{phyto} \) as a result of clouds.

### 3.3.4 Climatological Nitrate and Density Observations

Climatological monthly averages of potential density \( \sigma_\theta \) and nitrate concentrations \([NO_3^-]\) were obtained on a \( 1^\circ \times 1^\circ \) grid at standard levels from the World Ocean Atlas, 2005 (WOA05), an objectively analyzed climatology of temperature, salinity, oxygen and nitrate (Locarnini et al., 2006; Antonov et al., 2006; Garcia et al., 2006). To investigate seasonal variability, the monthly values from the WOA05 were averaged over winter (May through October) and summer (November through April). The long averages were necessary because of the sparse distribution of nutrient observations. To further mitigate the effects of sparseness of the nutrient observations, these seasonal maps were meridionally averaged between latitudes of \( 30^\circ \) \( S \) and \( 20^\circ \) \( S \) over the study domain to obtain longitude-depth sections.

### 3.4 Results

The trajectories of the mesoscale ocean eddies investigated in this study are shown in figure 3.2a. These include all eddies originating within the white contour, with anticyclones and cyclones shown as red and blue, respectively. As discussed below, the white contour corresponds to the region of positive cross correlation between CHL' and Ekman pumping. These SIO eddies have an average amplitude of 7.6 cm with no significant difference in the amplitudes of cyclones and anticyclones. However, eddies with amplitudes larger than 25 cm are almost exclusively anticyclonic (figure 3.2b). The average radial scale \( L_s \) is 107 km with cyclones being slightly smaller than anticyclones (figure 3.2c), resulting in
slightly higher rotational velocities in cyclones (figure 3.2d).

The SIO is one of the most cloud-free regions of the World Ocean, with the eastern portion of the domain exceeding 70% of the maximum possible coverage of daily SeaWiFS level-3 *CHL* (figure 3.3a) allowing for the collection of high-quality *CHL* and *C* _phyto_ measurements. Eddies spawned by the poleward-flowing Leeuwin Current along the west coast of Australia propagate westward with small equatorward deflections of anticyclones and poleward deflections of cyclones (Morrow et al., 2004; Chelton et al., 2007, 2011b). After leaving the productive regions of the Leeuwin Current, these eddies propagate into the oligotrophic SIO (figures 3.2a and 3.3b). Of particular interest are the eddies that form from the pinching off of anticyclonic meanders that contain coastal waters with high *CHL*. These anticyclones propagate to the west-northwest into a region of low *CHL* concentrations (figures 3.2a and 3.3b) within which there is strong positive cross correlation between chlorophyll anomalies *CHL'* and Ekman pumping velocity in the interiors of the eddies (figure 3.3d).

To investigate the relationship between eddy-induced Ekman pumping and *CHL'*, composite medians within radii of 2*L* _s_ of the eddy centroid, were computed in normalized radius coordinates (see section 1.2.1) from all the the eddies within a smoothed contour of the 0.2 cross correlation coefficient in figure 3.3d. The 0.2 cross correlation threshold was chosen to avoid eddies that propagate in the region of strong meridional gradient of *CHL* to the south of the study region (figure 3.3b) where eddy *CHL* variability is dominated by the rotational advection of the background *CHL* gradient around the eddy periphery (Chelton et al., 2011a). There were 358 cyclonic and 328 anticyclonic eddies with lifetimes of 12 weeks and longer within the region of ≥ 0.2 cross correlation during the 7 year period considered in this study (January 2001 through December 2007). This time period represents when we have overlapping data record of from SeaWiFS, QuikSCAT and the merged altimetry. Each weekly observation is referred to here as an eddy realization.
In total, there were 7604 realizations of cyclones and 8031 realizations of anticyclones.

As discussed in the introduction, surface currents generate Ekman upwelling and downwelling in the surface mixed layers of anticyclones and cyclones, respectively. These surface current effects are modulated by Ekman pumping from the SST influence on surface winds, also discussed in the introduction, which has more complicated spatial structure. This air-sea interaction contribution to the total eddy-induced Ekman pumping is usually secondary to surface current effects (Gaube et al., 2012). The Ekman upwelling values averaged over the interiors (i.e. within a radius of $L_s$) of the 8031 realizations of anticyclones in the study region have a mean value of 7 cm day$^{-1}$ with a standard deviation of 4 cm day$^{-1}$ and an upper 75th percentile of 12.7 cm day$^{-1}$. The mean and standard deviation of Ekman downwelling velocities in the cores of the 7604 realizations of cyclones are both 5 cm day$^{-1}$ and the upper 75th percentile is 9 cm day$^{-1}$. The larger Ekman pumping speeds in anticyclones are likely because the eddies with the largest rotational speeds $U$ are predominately anticyclones (figure 3.2,g).

The observed eddy-induced Ekman pumping velocities are small when compared with the values of tens of meters per day estimated for submesoscale upwelling that occurs along the peripheries of mesoscale eddies (Lévy et al., 2001, 2012). However, unlike the upwelling associated with submesoscale variability that is generally transient, the upwelling in the cores of mesoscale anticyclones is persistent throughout the lifetime of the eddy, though with temporally varying magnitude depending on the strength of the background winds. The spatial structure of eddy-induced Ekman pumping is such that the magnitudes of the vertical velocities are maximum in the cores of eddies of either polarity (contour in figure 3.4). The eddy cores contain fluid that is trapped by the rotational velocities of the eddy (McWilliams and Flierl, 1979; Early et al., 2011). The trapped core is thus isolated from the surrounding fluid, so that even small but persistent upwelling of nutrients from deep water into the euphotic zone is important to the ecosystem trapped
within the fluid. The $CHL'$ response to the influx of new nutrients from upwelling in anticyclones should thus be positive. Correspondingly, the $CHL'$ response to downwelling in cyclones is expected to be negative because both nutrients and phytoplankton are transported downward and out of the euphotic zone.

Composite medians of $CHL'$ and eddy-induced Ekman pumping were constructed for each calendar month (figure 3.4). The strong seasonal cycle of the $CHL'$ in figure 3.4 is striking (also see figure 3.5a). There is remarkable agreement between the spatial distributions of Ekman pumping and $CHL'$ during the winter period May though October. Positive $CHL'$ anomalies coincide almost exactly with contours of Ekman upwelling in anticyclones. Likewise, negative $CHL'$ anomalies coincide almost exactly with contours of Ekman downwelling in cyclones.

In contrast to the high wintertime correlation between eddy-induced Ekman pumping and $CHL'$, eddy-induced Ekman pumping has little influence on $CHL'$ in anticyclones from November through April. This is evident from the contours in figure 3.4, which show that eddy-induced Ekman pumping is actually stronger during this period than in the wintertime. The dipole structure of composite medians of eddy $CHL'$ in anticyclones during November through April is very similar to the dipole structures observed in global composites of $CHL'$ in midlatitude eddies of both polarities (Chelton et al., 2011a; Siegel et al., 2011). These structures arise from horizontal advection of the surrounding large-scale $CHL$ field by the rotational velocities in eddies.

Wintertime composite medians (May through October) of $CHL'$ and eddy-induced Ekman pumping are shown in figure 3.6a. Composite medians of $C'$ for the same eddies (figure 3.6b) are distinctly different from the $CHL'$. Anticyclones are characterized as having positive $C'$ along the southwestern edge of the westward-propagating eddies with negative $C'$ offset to the northeast of the eddy centroids. This structure is similar to the above-noted dipole signature of rotational advection. The geographic pattern of $C'$ is
consistent with the background southward gradient of $C_{phyto}$ found in this region (figure 3.3c).

In contrast to the anticyclones, the wintertime $C_{phyto}'$ in cyclones in the SIO are effectively zero, meaning that cyclones have only a weak effect on the background $C_{phyto}$ field. This is because the cyclones that form in the Leeuwin Current propagate west-southwest into the region of weak background gradients of $C_{phyto}$ in the southern part of the study region (see figure 3.3c).

As discussed in section 1.2.3, anomalies of the ratio $r_C$ are used here as a proxy for the phytoplankton growth rate, $\mu^1$. Positive and negative $r_C'$ are observed in the cores of anticyclones and cyclones, respectively (figure 3.6c). These $r_C'$ ratios are restricted to the inner cores of the eddies within radii of less than about $L_s$ from the eddy centroid. The maximum values of $r_C'$ in anticyclones are collocated with the maximum eddy-induced Ekman pumping. To the extent that this ratio is a good proxy for $\mu$, this suggests that the growth rates of the photoautotrophic communities trapped within SIO anticyclones are sustained by the Ekman pumping within the cores of the eddies. The same pattern, except reversed, is observed in cyclones, where the negative values of $r_C'$ are collocated with the maximum Ekman downwelling. This suggest that continuous surface convergence of fluid within the mixed layer towards the eddy center transports phytoplankton and nutrients downward and hence out of the euphotic zone, resulting in negative values of $r_C'$.

The seasonal appearance of the dipole structures in the $CHL'$ composites from November through April suggests the possibility that these structures might be the result of seasonal variability of the background $CHL$ field. To investigate the influence of the background $CHL$ gradient on the observed composites of $CHL'$ and $r_C'$, we further smoothed the $CHL$ fields to attenuate variability with periods shorter than 200 days and smaller than $6^\circ \times 6^\circ$. The 200 day filter span was used to remove all temporal eddy vari-

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$^1$In section 3.5, we estimate $\mu$ explicitly within the interiors of cyclones and anticyclones and show that higher growth rates are more often found in the interiors of anticyclones.
ability from the CHL so that the resultant CHL gradient was free of any eddy signatures. The magnitude of the northward component of the CHL gradient in anticyclones has a bimodal seasonal cycle with peaks in both May and December (figure 3.5b). In cyclones, the seasonal cycle of CHL gradient has a maximum in late spring and a minimum in early winter (figure 3.5b). These variations are, to some degree, in phase with what we would expect from the CHL' composites (figure 3.4). In particular, the dipole structure in eddies of both polarities is clearest during the summertime when the northward background CHL gradient is strongest (figures 3.4 and 3.5a). However, the maximum of the meridional gradient of the background CHL in May in the anticyclones occurs when the CHL distribution has a monopole structure that is consistent with a response to eddy-induced Ekman pumping. Thus, while the strength of the ambient CHL gradient may be a factor in determining when the dipole structure is apparent, it is evidently not the only factor controlling the CHL' distribution in SIO eddies.

To investigate whether the seasonal variations of the CHL distribution within eddies and their relationship to Ekman pumping arise from processes occurring below the sea surface, estimates of MLD from the individual Argo float profiles were collocated to the interiors of mesoscale eddies as described in section 1.2.4. The observed seasonal variability of CHL' (figure 3.5a) is approximately in phase with the seasonal variability of the MLD within the interior of cyclones and approximately out of phase within the interiors of anticyclones (figure 3.5c). MLD is determined from the stratification of the upper ocean, which is a function of solar heating and wind mixing. The wintertime maximum MLD (May, June and July), reaches a depth of 70 m in anticyclones but only 30 m in cyclones. From observations of SIO anticyclones, Thompson et al. (2007) suggested that the deeper MLD in the cores of anticyclonic eddies is related to warm upper-ocean temperature anomalies. Williams (1988) showed that the surface cooling of the warm-core anticyclonic eddies from air-sea interaction generates increased convection that enhances the deepening
of the mixed layer. Shallower MLDs in cyclonic eddies are attributed, at least in part, to
the absence of convective deepening over the cold cores of the cyclones in this regions.

The similarity of the geographic structures of the maximum $CHL'$ and eddy-induced
Ekman pumping during winter and the coincidence with the winter maximum of MLD
suggest that the changes in MLD may play a role in the observed seasonal variations of
$CHL'$ in eddy interiors. This can be assessed from the seasonal variability of nutrients
(nitrate, $NO_3^-$) and stratification constructed from the WOA05 as described in section
3.3.4. Zonal sections of meridionally averaged concentration of nitrate [$NO_3^-$] and potential
density $\sigma_\theta$ across the study domain for winter and summer are shown in figure 3.7. During
the Austral winter, the surface layer is well mixed down to approximately 100 m with
detectable levels of $NO_3^-$ found throughout the mixed layer. During the summer, however,
the water column is strongly stratified, with almost no $NO_3^-$ in the surface mixed layer
(figure 3.7).

The winter/summer differences in stratification and nutrient concentrations in the
upper 100 m (which corresponds approximately to the euphotic zone) are in agreement
with the observed $CHL'$ and $r'_C$. The highest upper-ocean $NO_3^-$ values and deepest MLD
occur when $CHL'$ and $r'_C$ are strongest in anticyclones. This suggests that the seasonal
appearance and disappearance of the coupling between $CHL'$ to eddy-induced Ekman
pumping could be a result of changes in stratification and the depth of the nutricline.

3.5. Discussion

The analyses presented in section 3.4. showed that Ekman pumping is strong over
both anticyclones and cyclonic eddies. During the wintertime, anomalously high $CHL$
within anticyclones is coincident with Ekman upwelling and anomalously low $CHL$ within
cyclones is coincident with Ekman downwelling. These responses of $CHL'$ to eddy-induced
Ekman pumping are seasonal in nature, disappearing during the summertime, despite the fact that eddy-induced Ekman pumping remains strong throughout the year. The possible mechanisms suggested for the observed seasonal variability of the biological response to Ekman pumping are further discussed in this section. We also assess whether the eddy-induced Ekman upwelling in anticyclones is sufficient to sustain the primary production rates that have been measured in a ship-based survey of an SIO eddy.

Eddies in the SIO are larger in amplitude and more nonlinear than in any other eastern boundary current system (Chelton et al., 2011b). As discussed in the introduction, the SIO is unusual in that anticyclones are associated with positive CHL'. The source of the high CHL in the cores of these anticyclones is the entrainment of eutrophic shelf water as the eddies form from meanders of the poleward-flowing Leeuwin Current along the eastern boundary of the SIO (Pearce and Griffiths, 1991; Moore et al., 2007; Dietze et al., 2009). The eddies selected for this study form primarily in the Leeuwin Current. Although some of the eddies are observed to originate far offshore of the Leeuwin Current (figure 3.2a), their high CHL' suggest that they actually originated in the Leeuwin Current. Eddies identified by the automated eddy tacking procedure can disappear and reappear weeks or even months later, depending on the evolving structures of the eddy from one time step to the next. This was verified for some case study eddies by studying animations of SSH in the SIO and comparing the tracks derived from the automated procedure to eddies tracked visually.

To investigate whether the composites of CHL' in figure 3.4 are not biased by the inclusion of eddies formed in the central SIO, we constructed composite medians of CHL' by month from only those eddies that formed east of 108° E between 21° S and 35° S. These eddies were considered to have originated from meanders of the Leeuwin Current. They account for about half of the eddies shown in figure 3.2a. The resulting composites are visually very similar to those shown in figure 3.4 (see appendix E). The influence of
the strong zonal CHL gradient across the Leeuwin Current results in a slight offset of the wintertime CHL′ maximum from the largest Ekman pumping velocities in the cores of anticyclones and a rotation of the dipole structures during the summertime. Otherwise, the composites of just Leeuwin Current eddies show the same winter-only coupling between CHL′ and eddy-induced Ekman pumping.

As summarized in section 3.4, eddy-induced Ekman pumping is coupled to CHL′ only in the wintertime. During late spring and summer, CHL′ is greatly reduced in the interiors of anticyclones despite strong eddy-induced Ekman upwelling (figure 3.4). Insolation is strongest during this time of year, which increases the stratification in the surface layer of the ocean. The association of weakened CHL′ and r′C with increased stratification suggest two possible connections: 1) the increased stratification causes a decrease in primary production, and 2) the phytoplankton community adjusts its photosynthetic pigment concentration to acclimate to the increased growth irradiance (photoacclimation). These two hypotheses are investigated in this section.

Consider first the hypothesis that seasonal variability of CHL in anticyclones is the result of photoacclimation of the photoautotrophic community. As the light available for photosynthesis is reduced in wintertime and the mixed layer deepens, the phytoplankton community is expected to respond by increasing intercellular CHL. This process of photoacclimation can increase observed CHL without an accumulation of biomass. This is accounted for in the estimation of μ (1.7) by inclusion of the nutrient and temperature dependent CHL to Cphyto ratio parameter rCmax. As shown in figure 3.8, rCmax decreases exponentially to a constant with increasing Ig. This result of photoacclimation is most pronounced for Ig ≤ 0.5 Ein m² hr⁻¹. When Ig ≥ 1Ein m² hr⁻¹, rCmax is nearly invariant. Estimates of Ig derived from 1.9 based on MLD from Argo profiles in the cores of SIO eddies show that 18% of the Ig values fall into the range Ig ≤ 1 Ein m² hr⁻¹ and 8% of Ig fall into the Ig ≤ 0.5 Ein m² hr⁻¹, (figure 3.8). The latter suggests that
photoacclimation is likely not the primary driver of the observed \( CHL' \) signatures of SIO anticyclones.

Even though photoacclimation does not appear to be the sole mechanism resulting in the observed elevated \( CHL' \) in the cores of anticyclones during the winter, a significant percentage (10%) of the observed \( I_g \) values that fall into the intermediate range of \( 0.5 \leq I_g \leq 1 \text{Ein m}^2\text{hr}^{-1} \), where increases in \( r_C \) can be attributed to photoacclimation. We therefore investigate the influence of SIO eddies on phytoplankton community growth rate during the summer and winter by estimating \( \mu \) directly, based on estimates of \( r_C \), MLD and \( K_d \), as described in section 1.2.5.

During the summer, when SIO anticyclone are observed to advect the ambient \( CHL \) field around their periphery, the average \( \mu \) within eddies during the summer is 0.5 day\(^{-1}\) and there is no significant difference between \( \mu \) in cyclones and anticyclones (figure 3.9e). During winter, when \( CHL' \) and \( r'_C \) are hypothesized to be modulated by eddy-induced Ekman pumping (figure 3.6), the average growth rate within SIO eddies is 0.61 day\(^{-1}\) and the largest growth rate values (\( \mu \geq 1.3 \text{day}^{-1} \)) occur almost exclusively in the interiors of anticyclones (figure 3.9f). It is noteworthy that only 25 estimates of \( \mu \) in excess of 1.3 day\(^{-1}\) were made during the Austral winter in the interiors of SIO anticyclones. The observed wintertime \( \mu \geq 1.3 \text{day}^{-1} \) might therefore not be significantly different in cyclones and anticyclones, in a statistical sense. However, for the observations analyzed in this study, phytoplankton growth rates within the interiors of anticyclones are higher in anticyclones than within cyclones during the Austral winter.

From the investigation of \( I_g \) and \( \mu \) within eddy interiors summarized above, we conclude that the role of photoacclimation in generating \( CHL' \) and \( r'_C \) anomalies in the cores of anticyclones cannot be neglected. However, more than 92% of the \( I_g \) estimates were in excess of 0.5 Ein m\(^2\) hr\(^{-1}\), a threshold \( I_g \) value above which \( r_C^{\text{max}} \) is nearly independent of \( I_g \). Direct estimates of \( \mu \) suggest that phytoplankton growth rates are
elevated in the interiors of SIO anticyclones during the Austral winter compared with cyclones. In light of these observations, it seems clear that photoacclimation is not the sole mechanism for the wintertime appearance of elevated $\text{CHL}'$ and $r_C'$ in the cores of SIO anticyclones. One or more additional mechanisms must therefore play a role in the observed wintertime ecosystem response to eddy-induced Ekman pumping in SIO anticyclones.

The second hypothesis considered here is that the observed seasonality of $\text{CHL}'$ and $r_C'$ is related to seasonal coupling and decoupling between the mixed layer and the nutricline. The vertical extent over which Ekman dynamics dominate over inviscid fluid motions (i.e., geostrophy) is a function of stratification through the dependence of turbulence on stratification (Ekman, 1905; Schudlich and Price, 1998). The source water for Ekman pumping is at the base of the mixed layer. Since the nutricline is generally deeper than the base of the mixed layer, wintertime deepening of the MLD could influence the effectiveness of eddy-induced Ekman pumping in supplying nutrients into the cores of mesoscale eddies and thus influencing $\text{CHL}$ and the $r_C$ ratio through wintertime coupling of the mixed layer and the nutricline.

The seasonality of coupling and decoupling between the mixed layer and the nutricline in anticyclones is shown schematically in figure 3.10. The positive SSH signature of anticyclones is mirrored as a deepening of subsurface isopycnals in the eddy interiors. The MLD, which is defined by the subsurface maximum of the density gradient (see section 1.2.4) therefore also deepens in the interiors of anticyclones. The MLD is deeper in winter than in summer because of enhanced wind mixing and reduced solar heating. With sufficient deepening, the base of the mixed layer intersects the nutricline, thus injecting nutrients into the euphotic zone. During the summer, increased solar radiation causes the mixed layer to shoal and decouple from the nutricline: Ekman pumping at the base of the mixed layer can no longer bring new nutrients into the euphotic zone, resulting in the
cessation of the elevated $CHL$ at the core of anticyclonic eddies.

The observations of climatological nutrient and stratification data shown in figure 3.7 support this hypothesized seasonal wintertime coupling and summertime decoupling between the mixed layer and the nutricline. The disappearance of anomalous $CHL$ and $r_C$ ratio in the cores of eddies during the summertime, despite continued strong eddy-induced Ekman pumping, may thus be attributed to the shallowing of the mixed layer.

Other mechanisms described in the literature that could sustain phytoplankton populations in anticyclonic eddies are not able to reproduce the observed $CHL$ and $r_C$ ratio maxima in the cores of anticyclonic eddies. For example, Mahadevan et al. (2008) proposed that nonlinear Ekman pumping, resulting from the inclusion of momentum advection in the formulation of Ekman pumping, could explain the enhanced primary production rates observed in the interior of a mode-water type eddy in the North Atlantic. In contrast to the observed wintertime monopole structure centered on the cores of SIO anticyclones, nonlinear Ekman pumping results in a dipole of upwelling and downwelling across the interiors of mesoscale ocean eddies. McGillicuddy et al. (2008) argued that a parcel of water transported into the euphotic zone in the nonlinear Ekman upwelling region of an eddy would not spend sufficient time in the euphotic zone to explain the anomalously high primary production values observed within the core of the mode-type eddy before being transported below the euphotic zone in a region of nonlinear Ekman downwelling. In addition, nonlinear Ekman pumping would enhance $CHL$ and the $r_C$ ratio along the eddy periphery, rather than in the cores of the eddies as is observed in the SIO eddies during winter.

In contrast to the replacement of the wintertime monopole of positive $CHL'$ in anticyclones with the dipole structure symptomatic of horizontal advection in the summertime, cyclonic eddies in the SIO are associated with negative $CHL'$ throughout the year (figure 3.4). The magnitude of the $CHL'$ in SIO cyclones varies seasonally with the
strongest negative anomalies occurring during the Austral winter. We believe that this puzzling characteristic of the seasonal variability in cyclones is an artifact of the filtering applied here in an effort to isolate eddy-related CHL variability. During the Austral winter, the CHL' values in anticyclones increase relative to the background CHL field (figure 3.5d). In contrast, CHL in the interiors of cyclones is always lower than the background and the difference between the background CHL and that inside of cyclones is nearly constant throughout the year (figure 3.5d). Because cyclones are generally adjacent to one or more anticyclones, the (real) seasonal increase in CHL within anticyclones, relative to the background, results in an apparent seasonal decrease in CHL as a result of the spatial high-pass filtering.

To illustrate this artifact of filtering, consider a pair of counter-rotating eddies propagating through the SIO. During the Austral summer, the anticyclonic eddy advects the ambient CHL around it’s periphery resulting in the observed dipole structure of CHL’ (see figure 3.4). The average value of horizontally stirred CHL within the anticyclonic eddy is effectively the same as the background CHL (figure 3.5d). In the cyclonic eddy, eddy-induced Ekman downwelling pushes CHL and nutrients out of the euphotic zone throughout the year, causing the CHL values within the cyclone to be lower than the background (figure 3.5d). During the Austral winter, however, CHL within the anticyclonic eddy increases relative to the background while the CHL within the cyclonic eddy remains lower than the background CHL (figure 3.5d). When these fields are spatially high-pass filtered as described in section 1.2.3 to create the CHL' analyzed in this study, the wintertime increase in CHL within the anticycle increases the overall average CHL surrounding the cyclones, thus resulting in an artificial increase in the magnitude of the negative CHL' within cyclones.

To test this hypothesis we conducted a simple simulation by removing the CHL values within the interiors of anticyclonic eddies prior to spatially high-pass filtering.
After removing the smoothed \( CHL \) field free of the seasonal variability of \( CHL' \) within anticyclones, the resultant monthly \( CHL' \) composites of cyclones vary only slightly in magnitude from month to month and do not have the clear seasonal cycle seen in figures 3.4 and 3.5a. We thus conclude that the apparent seasonal change in the magnitude of \( CHL' \) in cyclones is an artifact of the increase in the magnitude of \( CHL' \) in anticyclones. The mirrored nature of the seasonal cycles of \( CHL' \) in cyclones and anticyclones in figure 3.5a is consistent with this interpretation.

It is perplexing that an accumulation of biomass (i.e., an increase of \( C_{phyto} \)) is not observed in association with the increase of \( CHL \) in the cores of anticyclones during the wintertime (figure 3.6b). In oligotrophic regions such as the SIO, a decoupling of \( C_{phyto} \) from \( CHL \) can occur from photoacclimation (Behrenfeld et al., 2005). As discussed above, however, photoacclimation does not appear to be the primary mechanism controlling seasonal variations of \( CHL \) and \( r_C \) in anticyclones. This paradox may be an indication of the effects of grazing of the phytoplankton by zooplankton trapped in the cores of the anticyclones.

Phytoplankton in oligotrophic ecosystems are generally tightly coupled to their grazers (Burkill et al., 1993; Froneman and Perissinotto, 1996). Paterson et al. (2007) investigated the grazing of phytoplankton by zooplankton in a pair of Leeuwin Current eddies of opposite polarities. The anticyclonic eddy contained significantly more zooplankton biomass than the cyclone. The zooplankton community was dominated by dinoflagellates that primarily graze upon diatoms. Estimates of the phytoplankton mortality from grazing by zooplankton in both the cyclone and anticyclone suggested that grazing rates were sufficient to account for all the primary production measured within the eddies. A tight coupling between phytoplankton biomass and the grazer community could explain why an accumulation of biomass is not observed, even when high growth rates give rise to increased \( CHL' \) and \( r'_C \) (figure 3.6c). While this presents an interesting hypotheses
that could help to explain the observed lack of accumulation of phytoplankton biomass ($C_{phyto}$), the postulated role of zooplankton in controlling phytoplankton biomass cannot be tested from the satellite data analyzed in this study. Additional in situ measurements will be necessary to test this hypothesis rigorously.

The degree to which eddy-induced Ekman pumping can contribute to new production in SIO anticyclones can be assessed from consideration of a representative eddy-induced Ekman upwelling velocity of $W_E = 0.1 \text{ m day}^{-1}$ in conjunction with published values for the concentration of dissolved inorganic nitrogen ([DIN]) measured in the core of an SIO anticyclone. For the this calculation, we consider [DIN] as representative of the nitrate concentration ([NO$_3^-$]). The influx of NO$_3^-$ into the euphotic zone from below the nutricline for eddy-induced Ekman upwelling of $W_E$ can be calculated as:

$$F_{NO_3^-} = W_E[NO_3^-] - K_v \frac{\partial[NO_3^-]}{\partial z}$$

(3.1)

where [NO$_3^-$] is the mean NO$_3^-$ concentration just below the nutricline (10 mmol m$^{-3}$), $K_v$ is the vertical eddy diffusivity taken to be $10^{-4} \text{ m}^2 \text{ s}^{-1}$, and $\partial[NO_3^-]/\partial z$ is the vertical gradient of [NO$_3^-$] in the nutricline (0.025 mmol m$^{-4}$). These values of [NO$_3^-$] and $\partial[NO_3^-]/\partial z$ are estimated from observed vertical profiles of DIN taken during the 2003 multi-disciplinary survey conducted on the R.V. Southern Surveyor (Greenwood et al., 2007). For a 100 m depth of the euphotic zone and a pumping velocity of $W_E = 0.1 \text{ m d}^{-1}$ we estimate the flux of NO$_3^-$ per unit area into the euphotic zone to be 0.1 mmol NO$_3^-$ m$^{-2}$ d$^{-1}$. To assess whether this Ekman upwelling induced flux of nitrate is sufficient to sustain the observed primary production values observed in the anticyclones surveyed by Waite et al. (2007), we need to estimate the portion of primary production due to the reduction of NO$_3^-$, which is often referred to as new production. New production was not directly estimated in the 2003 survey of the Leeuwin Current eddies. Primary production values in units of carbon can, however, be converted to nitrogen assuming all
of the nutrients obey the Redfield ratio. Waite et al. (2007) reported average primary production in the interior of an SIO anticyclone to be $1 \text{ mg C m}^{-3} \text{ d}^{-1}$, which converts to $0.429 \text{ mmol N m}^{-2} \text{ d}^{-1}$ of primary production per unit area assuming a euphotic zone depth of 100 m. This represents the total production in units of nitrogen, which can be converted to new production using a published f-ratio (the fraction of primary production fueled by the reduction of $NO_3^-$). Waite et al. (2007) reported an f-ratio of 0.35 in the core of a Leeuwin Current anticyclone, which yields an estimate of $0.15 \text{ mmol } NO_3^- m^{-2} d^{-1}$ of new production per unit area. This is close to the estimate of $0.1 \text{ mmol } NO_3^- m^{-2} d^{-1}$ obtained above from (3.1). The difference is within the uncertainty of the measurements and the assumption that macronutrients in the cores of anticyclones vary according to the Redfield ratio. It thus appears that the flux of $NO_3^-$ into the euphotic zone solely from Ekman upwelling could sustain the observed new production rates within the cores of anticyclonic SIO eddies.

3.6. Summary and Conclusions

Satellite observations have shown that mesoscale ocean eddies in the SIO are highly nonlinear (Chelton et al., 2011b). Fluid is therefore trapped within the cores of these eddies, thus limiting mixing with the surrounding water (Early et al., 2011). The cores of the eddies also coincide with the regions of the highest surface current-induced wind stress curl, resulting in Ekman pumping velocities that approach order ten centimeters per day. Evidence was presented that the Ekman upwelling in anticyclonic SIO eddies sustains enhanced CHL concentrations when compared to the oligotrophic waters that surround them. Anomalies of CHL and the CHL to $C_{phyto}$ ratio $r_C$ in the interiors of the anticyclones exhibit a strong seasonality with the highest CHL and $r_C$ values occurring during the Austral winter.
Based on climatologies constructed from historical measurements of temperature, salinity and nitrate, we have speculated that blooms of CHL occur only during the winter because this is the only time of the year that the mixed layer is sufficiently deep to reach the nutricline. The shallowing of the Ekman layer during the Austral summer decouples the euphotic zone from the nutricline, thus reducing the summer time CHL response to Ekman upwelling. While in situ observations are not sufficient to test this hypothesis rigorously, we showed that this seasonal coupling and decoupling of the mixed layer and the nutricline is more consistent than other possible mechanisms with the observed seasonal variability of eddy-induced CHL anomalies in the SIO. Our hypotheses could be tested with coupled ecosystem models.
FIGURE 3.1: (a) Location of individual Argo float profiles that occurred within anticyclonic (a) and cyclonic (b) eddies in the SIO. Argo floats were collocated to the interiors of eddies using the outermost closed contour of SSH that enclosed the eddy centroid. The x and y location of each Argo profile relative to the eddy centroid has been normalized by the radius scale $L_s$ of the eddy realization nearest in time to the profile.
FIGURE 3.2: (a) Trajectories of the eddies investigated here. Cyclones shown in blue and anticyclones in red. The start location of each eddy is demarcated with a black circle. The contour of 0.2 smoothed cross correlation between CHL' and eddy-induced Ekman pumping velocity (figure 3.3d) is shown in white. This contour is used to select eddy realizations from which composite medians and the statistics shown in panels (b-d) of this figure are constructed. See section 3.4. for justifications of the use of this contour and section 3.5. and appendix E for a discussion of difference in the results of the analysis of these eddies and eddies originating only in the Leeuwin Current. (b) Histograms of eddy amplitude with cyclones shown in blue and anticyclones in red. (c) The ratio of the number of cyclones to anticyclones as a function of eddy amplitude (b). (d) and (e) are the same as (b) and (c) except for the eddy radius scale $L_s$. (f) and (g) are the same as (b) and (c) except for the rotational speed $U$. 
FIGURE 3.3: Maps for the South Indian Ocean for: (a) Percent coverage of daily SeaWiFS CHL observations reported as percent of maximum possible number of observations. (b) Mean CHL from SeaWiFS calculated over the 8 year study period. (c) Mean $C_{phyto}$, estimated from backscattering at 440 nm as measured by SeaWiFS. (d) Cross-correlation coefficient between $CHL'$ and eddy-induced Ekman pumping velocity within eddy interiors.
FIGURE 3.4: Monthly composite medians of $CHL'$ anomalies overlaid with contours of eddy-induced Ekman pumping velocity. Contour interval of Ekman pumping velocity is $1 \text{ cm day}^{-1}$, negative velocities are shown as dashed curves. The left two columns are of anticyclones and cyclones, respectively, for January through June progressing forward in time from top to bottom. The right two panels are the same as the left two, except for July through December. The title of each composite median has the number of eddy realizations used in the composite.
FIGURE 3.5: (a) Seasonal variability of $CHL'$ from SeaWiFS within eddy interiors for anticyclones and cyclones shown as solid and broken curves, respectively. The seasonal cycle is calculated by regression of the annual cycle and it’s first harmonic. (b) Seasonal variability of the meridional component of the background $CHL$ gradient from SeaWiFS. (c) The seasonal cycle of MLD estimated from Argo float profiles. (d) The seasonal cycle of $CHL$ within eddy interiors. The grey curve in (d) is the seasonal cycle of $CHL$ averaged over the study domain. The green overlay on all four panels represents the wintertime when $CHL'$ anomalies are collocated with maximum eddy-induced Ekman upwelling in the interiors of SIO anticyclones (see figures 3.4 and 3.6).
South Indian Ocean wintertime composite medians

Figure 3.6: Composite medians for SIO eddies during the winter (May through October): (a) CHL anomalies in anticyclones (left) and cyclones (right) (b) $C_{phyto}'$ anomalies and (c) $r_C'$ anomalies. The contours overlaid on each panel are the composite median Ekman pumping velocities with a contour interval of 1 cm day$^{-1}$; negative velocities are shown as dashed lines.
FIGURE 3.7: Longitudinal variations of seasonal averages of potential-density $\sigma_\theta$ (color) and nitrate concentration $[NO_3^-]$ (contours) from the World Ocean Atlas 2005: (a) Winter (May - October) and (b) summer (November - April). A contour interval of 0.1 mmol m$^{-3}$ and range 0.2 – 1.0 mmol m$^{-3}$ were chosen so that only nutrients in approximately the first 100m of the water column are shown.
FIGURE 3.8: Maximum potential $\text{CHL}:C_{\text{phyto}}$ ratio, $r_C^{\text{max}}$ as defined by equation (1.8) as a function of median mixed layer growth irradiance ($I_g$) (black curve) and the cumulative probability density function of observed $I_g$ as defined by equation (1.9) within the cores of SIO eddies (gray line).
FIGURE 3.9: Statistical distributions of phytoplankton growth rate $\mu$ in summer (left) and winter (right) computed within the interiors of cyclones (blue curves) and anticyclones (red curves). Histogram are shown in (a) and (b) and the corresponding cumulative histogram are shown in (c) and (d). The ratio of the numbers of cyclones to anticyclones are shown in (e) and (f).
FIGURE 3.10: Schematic representation of the winter coupling (left) and subsequent summer decoupling (right) of the mixed layer from the nutricline within anticyclonic eddies. During the winter, MLD is deep and impinges on the nutricline allowing new nutrients to be upwelled into the interiors of anticyclones, resulting in the observed elevated winter time $CHL'$. During the summer, MLD shallows, decoupling the mixed layer from the nutricline and limiting the $CHL'$ to the base of the mixed layer where it cannot be observed from satellites.
Geographical Variability of the Influence of Mesoscale Ocean Eddies on Phytoplankton Communities

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4. GEOGRAPHICAL VARIABILITY OF THE INFLUENCE OF MESOSCALE OCEAN EDDIES ON PHYTOPLANKTON COMMUNITIES

4.1. Abstract

Recent observations have revealed that nearly all midlatitude mesoscale eddies are nonlinear, with rotational velocities that exceed their rate of propagation, hence containing a region of trapped fluid within their interiors. In this study we show that ecosystems and associated nutrient concentrations entrained into the eddy interiors at the time and location of eddy formation predetermine whether elevated concentrations of phytoplankton are observed in cyclones or anticyclones. These trapped ecosystems are subjected to physical processes throughout the lifetimes of the eddies. During eddy formation, isopycnals are displaced upward in cyclonic eddies and downward in anticyclonic eddies. This results in an episodic influx of nutrients into cyclones and a flushing of nutrient out of the euphotic zone in anticyclones. After this transient phase of development, vertical fluxes of nutrient are controlled primarily by the rotating surface currents that impart a curl of the surface stress with a vorticity opposite that of the eddy. This results in Ekman upwelling in the interiors of anticyclones and downwelling in cyclones. Eddy-Induced Ekman pumping thus sustains a slow, but persistent, influx of new nutrients into the cores of anticyclones and out of the cores of cyclones. The vertical transport of nutrients and phytoplankton acts in conjunction with the azimuthal advection of background biogeochemical gradients by eddy surface currents to create the observed biological signatures of eddies.

In this chapter, we show that the relationship between the observed biological signatures of eddies and their polarity varies regionally. Five diverse regions of the World Ocean are investigated, each having biological signatures in the eddies attributable to one or more of the above-mentioned mechanisms. Inferences gleaned from these regional
investigations are used to guide a statistical analysis of all midlatitude eddies.

Eddies that form in moderately productive regions of the World Ocean are observed to have sea surface height and biological signatures that are negatively correlated, resulting in elevated or suppressed bio-optical properties in cyclonic and anticyclonic eddies, respectively. Conversely, eddies originating in oligotrophic regions of low primary productivity are observed to have surface height expressions that are positively correlated with their chlorophyll signatures, resulting in enhanced chlorophyll concentrations in the interiors of anticyclones and low chlorophyll in cyclones.

A common theme that emerges from the regional analyses is that the concentrations of phytoplankton communities within eddy interiors oscillate at an average period of about once per month. Predator-prey interactions between the phytoplankton communities and their grazers are proposed as a possible mechanism for these oscillations. A simple time-dependent ecosystem model is used to explore the validity of this hypothesis.

4.2. Introduction

Mesoscale eddies with spatial scales of $\sim 10 - 100 \text{ km}$ are ubiquitous features of the World Ocean. These eddies are known to play an important role in the meridional transfer of heat (Qiu and Chen, 2005; Roemmich and Gilson, 2001), salt (Chaigneau et al., 2011) and the nutrient enrichment of oligotrophic pelagic ecosystems (McGillicuddy and Robinson, 1997; Oschlies, 2002; McGillicuddy et al., 2007). Outside of the tropics, nearly all of the eddies tracked in sixteen years of merged sea-surface height (SSH) fields are nonlinear (Chelton et al., 2011b). Nonlinearity implies a region of trapped fluid within eddies which can advect ecosystems away from the region of eddy formation. The rotational currents of eddies can also act to advect ecosystems around eddy peripheries (Chelton et al., 2011a; Siegel et al., 2011). The surface currents of eddies also impart a curl in the surface stress
that is opposite to that of the vorticity of the eddy, resulting in Ekman upwelling in anticyclones and downwelling in cyclones (Martin and Richards, 2001; McGillicuddy et al., 2007; Gaube et al., 2012). Finally, isopycnals within the interiors of eddies are shoaled in cyclones and deepened in anticyclones during eddy formation, resulting in an ephemeral vertical flux of nutrients within their cores.

The relative importance of each of the aforementioned mechanisms can be quantified regionally. In this study, we analyze a suite of remotely sensed data to investigate the ecosystem responses to the various physical forcing in five diverse regions of the World Ocean.

This study is organized as follows: Section 4.3 introduces the four primary mechanisms by which eddies influence phytoplankton. Section 4.4 provides an overview of the observations used in this study and references back to section 1.2., which describes the filtering applied to the data to isolate variability at the mesoscale. The procedure for collocating the various satellite observations to the interiors of mesoscale eddies to create eddy composites and time series of observations within eddies is described in section 4.4.1. A simple time-dependent ecosystem model used to explore predator-prey interactions within eddies is described in section 4.4.2. Results are presented in section 4.5 beginning with a description of midlatitude eddy composites (section 4.5.1) and followed by the investigation of eddies in five regions of the World Ocean. These regions are separated by their primary production levels; regions that sustain moderate levels of primary production (referred to hereafter as mesotrophic regions, section 4.5.2) and low levels of primary production (referred to hereafter as oligotrophic regions, section 4.5.3). A global analysis of the influence of eddies on phytoplankton carbon and chlorophyll is presented in section 4.5.3. The oscillatory temporal variability of phytoplankton biomass within eddies is explored using a simple time dependent ecosystem model in section 4.5.4. Our conclusions drawn from the observations are presented in section 4.6, along with recom-
mendations for future modeling studies to elucidate the mechanisms responsible for the observed variability.

4.3. Historical Background

4.3.1 The Entrainment of Phytoplankton During Eddy Formation

Ecosystems trapped within eddies during formation “prime” the eddy interior towards either elevated or suppressed phytoplankton concentration. Take for example an eddy that entrains eutrophic coast water during formation. The ecosystem entrained into the eddy interiors has a specific concentration of nutrients. These nutrients can be in multiple forms, biologically available dissolved nutrients or those assimilated into organisms which can be recycled at a later date. These nutrients are trapped within the eddy as it propagates away from where it originated. This trapping of fluid within eddies was eloquently shown in recent idealized modeling study of mesoscale ocean. The fluid entrained within an eddy during formation is retained during the entirety of its lifetime, with only minor leakage caused by the slow decay of the eddy and loss of trapped fluid within the eddy core (Early et al., 2011). Previous laboratory experiments have shown that the source of the fluid trapped within an eddy is a function of the eddy polarity (Pearce and Griffiths, 1991). The preferential entrainment of water with elevated phytoplankton concentration into either cyclonic or atycliclnc eddies is determined by the direction of the meandering current from which the eddies form and the direction of the ambient phytoplankton concentration gradient.

In the poleward-flowing western boundary currents, such as the Gulf Stream, Kuroshio Current and Brazil Current, cyclonic meanders entrain water from the shoreward side of the current into their interiors which is generally higher in phytoplankton concentration than the water offshore. These cyclonic meanders subsequently pinch off
and form isolated cyclonic eddies that contain a parcel of nearshore water trapped within their interiors. Conversely, anticyclonic meanders of poleward-flowing western boundary currents entrain water from offshore of the current, pinching off into anticyclonic eddies that contain water generally lower in phytoplankton concentration than cyclones formed from the same boundary currents.

In equatorward-flowing eastern boundary currents such as the California Current, Peru-Chile Current and Benguela Current, cyclonic eddies also preferentially entrain coastal water during formation. This coastal water generally has phytoplankton concentrations that are elevated compared to the water found offshore of the current. Anticyclones formed in these classic eastern boundary currents preferentially entrain water from offshore of the current, generally devoid of nutrients and phytoplankton.

There is an exception to these “classic” boundary current systems. The anomalously poleward-flowing Leeuwin Current along the west coast of Australia generates anticyclones that have elevated phytoplankton concentrations (Pearce and Griffiths, 1991; Moore et al., 2007). This results from anticyclonic meanders of the Leeuwin Current preferentially entraining water from the continental shelf. The subsequent pinching off of these meanders forms isolated anticyclonic mesoscale eddies. Conversely, cyclones formed in the Leeuwin Current preferentially entrain water from the offshore side of the current, priming the interiors of the cyclones to have locally depressed phytoplankton concentrations.

Undulating meanders of vertically sheared open-ocean currents can similarly generate eddies that entrain the ambient phytoplankton concentration and the nutrients assimilated within these organisms. To illustrate this process, consider meanders of an eastward zonal current in a southward ($\partial CHL/\partial y < 0$) ambient $CHL$ gradient, as can be seen in the image of $\log_{10}(CHL)$ for the South Pacific (figure 4.1). Within northward undulations, high $CHL$ (also high nutrient) water is advected to the north. As these meanders becomes unstable, cyclonic eddies pinch off, entraining the high $CHL$ water. The opposite is seen
in southward undulations within which low CHL (low nutrient) water is advected to the south. These southward undulations become unstable and pinch off forming anticyclonic eddies that will contain water low in CHL.

The above scenario can be seen in the image of CHL' anomalies shown in figure 4.1b where cyclonic eddies (identified by negative SSH anomalies) contain positive CHL' anomalies and anticyclonic eddies (identified by positive SSH anomalies) are associated with negative CHL' anomalies.

It will be shown in this study that the entrainment of the ambient phytoplankton concentration into eddies during formation predetermines whether eddies of a particular polarity are associated with locally elevated or depressed phytoplankton concentrations.

4.3.2 The Horizontal Advection of Phytoplankton Around Eddies

The azimuthal advection of large-scale biophysical gradients around the peripheries of eddies has been observed to be the dominant mechanism by which eddies influence surface CHL in the midlatitudes (Chelton et al., 2011a). This can be thought of as the “stirring” of background gradient by the rotational velocities of eddies, which results in a dipole structure in composites of anomaly fields. The sign, magnitude and orientation of these dipoles is a function of the direction and magnitude of the gradient of the ambient field being stirred by the eddy, the speed of the eddy currents and the rotational sense of the eddies. For example, a counterclockwise-rotating eddy propagating through a northward CHL gradient ($\partial CHL/\partial y > 0$) will result in a positive anomaly located in the southwest quadrant of the eddy and a negative anomaly in the northeast quadrant (figure 4.2a). In the same background field a clockwise-rotating eddy will result in a negative anomaly in the northwest quadrant and a positive anomaly in the southeast quadrant (figure 4.2b).
4.3.3 Eddy Pumping

Early observations of eddies formed by the pinching-off of meanders of the Gulf Stream found that ecosystems within the interiors of newly formed cyclonic eddies had elevated biomass at the base of the euphotic zone (Backus et al., 1981). However, these anomalies decayed soon after ring formation (Wiebe et al., 1976). Observations of a cyclonic eddy formed in the lee of the Hawaiian Islands similarly attributed elevated nitrate concentration and primary production to the uplifting of isopycnals within the core of the cyclone and the subsequent pumping of nutrients into the euphotic zone, a process that the authors referred to as “eddy pumping” (Falkowski et al., 1991). Within this particular Hawaiian cyclone, both the maximum chlorophyll and the maximum change in variable fluorescence (a measure of the efficiency of photosynthesis) occurred at a depth of 50 m, presumably close to the base of the euphotic zone.

Eddy pumping occurs during formation of cyclones when isopycnals are displaced upwards through the process of baroclinic adjustment, resulting in the introduction of new nutrients from below the nutricline into the euphotic zone that stimulates enhanced phytoplankton growth rates which could result in elevated phytoplankton biomass (figure 4.2e). Baroclinic adjustment is the vertical displacement of isopycnals to allow the rotational velocity of eddies to approach zero in the deep ocean. Isopycnals are displaced downwards during the formation of anticyclones, which flushes nutrients and organisms downward from the euphotic zone, resulting in negative anomalies of photoautotrophic biomass and growth rate (figure 4.2f).

The vertical velocities generated by eddy pumping can be estimated by comparing published observations of isopycnal displacements within eddies to the time it takes eddies to grow and reach a steady state. Siegel et al. (1999) reported that the root mean squared isopycnal displacement resulting from the passage of a Gulf Stream cyclone through the BATS (Bermuda Atlantic Time series Study) moorings was \( \approx 50 \) m. We can estimate
the time-rate of eddy growth from our altimetric SSH observations. By averaging the
amplitude of numerous eddies as a function of eddy age, time series of eddy amplitude can
be constructed. For Gulf Stream eddies, defined to have originated between $35^\circ N - 45^\circ N$
and $290^\circ E - 325^\circ E$, the initial growth rate gives an estimate of about 4 weeks for the
time-scale of eddy formation (figure 4.3a). The order of magnitude estimate of the vertical
velocities associated with eddy pumping are thus $O \left(1 \text{ m day}^{-1}\right)$ for large eddies in the
North Atlantic. These eddy pumping velocities are expected to scale approximately lin-
early with eddy amplitude. Since Gulf Stream eddies in the North Atlantic are quite large
in amplitude ($\sim 20 \text{ cm}$) when compared with other midlatitude eddies (Chelton et al.,
2011b), an eddy pumping velocity of $1 \text{ m day}^{-1}$ likely represents an upper-bound esti-
mate. As an interesting side-note, the time-scale of eddy formation appears to be fairly
consistent throughout the World Ocean (figure 4.3). Most of the variability of eddy pump-
ing vertical velocities will therefore be a result of variability in isopycnal displacement,
which is approximately inversely proportional to the eddy amplitude.

Eddy pumping of the opposite sign occurs during the decay of eddies from the relax-
ation of subsurface isopycnals (upward in anticyclones and downward in cyclones)(Franks
et al., 1986). However, the decay of eddy amplitude is slow, hence eddy pumping velocities
during eddy decay would be small, compared to those occurring during eddy formation.
Therefore eddy pumping associated with eddy decay is not likely nearly as important as
eddy pumping during formation, thus we do not explore whether eddy pumping during
eddy decay triggers an observable ecosystem response.

4.3.4 Eddy-induced Ekman Pumping

The surface currents associated with mesoscale ocean eddies impart a curl of the
surface stress from the relative motion between surface air and water (Stern, 1965). This
surface stress curl has a polarity opposite to that of the vorticity of the eddy, thus attenu-
ating the eddies by generating Ekman upwelling in the cores of anticyclonic eddies and
downwelling in the cores of cyclonic eddies (Dewar and Flierl, 1987). In addition to the surface current-induced Ekman pumping, air-sea interaction associated with eddy-induced spatial variations of SST generates a wind stress curl and therefore Ekman pumping that are proportional to the crosswind SST gradient (Chelton et al., 2004; O’Neill et al., 2010). Eddy-induced Ekman pumping upwells nutrients from below the euphotic zone into the interiors of anticyclones (figure 4.2d) and downwells organisms out of the euphotic zone in cyclonic eddies (figure 4.2c). Vertical velocities associated with eddy-induced Ekman pumping are on the order $10 \text{ cm day}^{-1}$ globally Gaube et al. (2012), but can approach values of $\sim 1 \text{ m day}^{-1}$ in large eddies at high wind speeds Martin and Richards (2001).

Eddy-induced Ekman pumping was hypothesized by Martin and Richards (2001) as a possible mechanism for sustaining an anomalous phytoplankton bloom that was observed in the core of a mode-water type anticyclonic eddy in the North Atlantic Ocean. They estimated eddy-induced Ekman pumping velocities from geostrophic surface currents calculated for a Gaussian-shaped SSH signature with an amplitude of about $17 \text{ cm}$ scaled to match the eddy surface current velocities measured from ship transects across the eddy. For a spatially uniform $15 \text{ m s}^{-1}$ surface wind (the maximum wind speed they observed during their survey and subsequently used to estimate Ekman pumping), the upwelling was in excess of $1 \text{ m day}^{-1}$ within the eddy core. For a more typical wind speed of $7 \text{ m s}^{-1}$, the Ekman pumping velocity for the same eddy would be $40 \text{ cm day}^{-1}$.

An intensive ship survey of another mode-type anticyclone in the Sargasso Sea showed definitively the importance of eddy surface current-induced upwelling (McGillicuddy et al., 2007). Phytoplankton concentration within the core of the eddy was eight standard deviations higher than the mean background field. Sulfur hexafluoride tracer released in the eddy core upwelled at a rate of $40 \text{ cm day}^{-1}$, which compared well with the calculated Ekman pumping velocity averaged over the time period of the tracer experiment computed from only the eddy surface current effects (McGillicuddy et al.,
Eddy-induced Ekman pumping has also been observed from satellites to influence the surface chlorophyll concentration ($CHL$) of ecosystems trapped within anticyclonic eddies of the South Indian Ocean (see chapter 3). A $CHL$ response to eddy-induced Ekman pumping is observable from satellite measurements of near-surface $CHL$ in anticyclones formed in the poleward-flowing Leeuwin Current because the surface $CHL$ is representative of the $CHL$ throughout the euphotic zone (Feng et al., 2007).

Although the vertical velocities associated with eddy-induced Ekman pumping are small compared to those resulting from eddy pumping, the associated vertical nutrient fluxes are transient, only occurring during the formation of an eddy. Since the magnitude of the ecosystem response to this pumping presumably scales with the eddy amplitude, an observable response likely occurs only in regions of large amplitude eddies. Moreover, because the biological response to eddy pumping is generally largest near the bottom of the euphotic zone (Backus et al., 1981; Falkowski et al., 1991; McGillicuddy et al., 2007; Thompson et al., 2007), satellite measurements of surface biological properties might not be able to detect an ecosystem response to eddy pumping (Siegel et al., 2011).

### 4.4. Methods

An overview of the methods used to identify and track mesoscale eddies in altimetric SSH maps is given in section 1.2.1. In this study we analyze eddies tracked in SSH maps starting in October 1992 through December 2011. A total of 28,928 anticyclones and 30,857 cyclones with lifetimes longer than 12 weeks were identified in the 19-year data record considered here.

Measurements of eddy-induced Ekman pumping were made for every eddy identified during the 10-year QuikSCAT record beginning in July 1999 and ending in November
2009. The processing of QuikSCAT data and filtering applied to isolated the mesoscale are described in section 1.2.2.

Near-surface chlorophyll (\(CHL\)) and phytoplankton carbon (\(C_{phyto}\)) were estimated from water leaving radiance measured by SeaWiFS as described in section 1.2.3. The SeaWiFS mission began on 19 September 1997 and ended on 10 September 2010.

Phytoplankton growth rates \(\mu\) were computed from measurements of \(CHL\), \(C_{phyto}\) and mixed layer light levels as described in section 1.2.5. Estimates of mixed layer light levels from which \(\mu\) is estimated rely on measurements of the mixed layer depth (MLD) within eddy interiors. MLD was estimated from Argo floats as described in section 1.2.4 and for this study a total, 8089 and 8340 estimates were made for cyclones and anticyclones, respectively during the \(\sim\) 13 year SeaWiFS data record.

### 4.4.1 Collocation of Observations within Eddies Interiors

To investigate the signatures of eddies in the SSH, Ekman pumping and various bio-optical observations, we take advantage of the 18+ year AVISO SSH data record that spans a longer period than any of the other observations. For observation from QuikSCAT and SeaWiFS, we construct eddy-centric composites for the entire data record of each of these sensors. The advantage of using the entire data record for each satellite sensor individually is that it maximizes the numbers of weekly observation included in each eddy composite. Note, however, that this can result in composites that show one variable as a pseudo-color image (for example \(CHL'\)) overlaid with contours of another variable (e.g., SSH) that are constructed from observations with two different data record lengths. Since the response of both \(CHL\) and \(C_{phyto}\) is presumably invariant in time, the use of different data records doesn’t affect the interpretations presented herein.

The bio-optical and Ekman pumping fields, filtered as described in sections 1.2.2 and 1.2.3 were collocated with the interiors of the eddies as identified from their SSH signatures. Composites of \(CHL'\) and \(C'_{phyto}\), along with Ekman pumping velocity were
then constructed from thousands of irregularly-shaped eddies as discussed in section 4.5. The compositing was done on a common grid by scaling the distance from the eddy centroid to each $1/4^\circ$ grid point in the eddy interior by the instantaneous eddy radius $L_S$. Each normalized grid location was then interpolated onto a high resolution grid for radial distances ranging from 0 to $2L_S$.

It will become apparent in section 4.5.1 that the eddy influence on $CHL'$ composite constructed for all midlatitude eddies consists predominantly of horizontal advection of the background bio-optical gradient by the azimuthal advection of the background gradients from the rotational velocity within the eddy interior. The composites of $CHL'$ for all midlatitude eddies were therefore computed in a rotated frame of reference determined by the orientation of the large-scale background gradient computed from 200-day smoothing of the $CHL$ fields (see Chelton et al., 2011a, and chapter 2). When this ambient gradient vector had a northward component, the anomalies were rotated to orient the large-scale gradient vector to a polar angle of $90^\circ$. When the ambient gradient vector had a southward component, the anomalies were rotated to orient the large-scale gradient vector to a polar angle of $-90^\circ$.

For the five study regions investigated in section 4.5, we select eddies that originated in each of these regions and the composites consist only of those weekly observations when the eddy centroid is located within the bounds of the region. Observations associated with that eddy are no longer included in the composites after the eddy propagates outside of the study region.

Throughout this study, the composites constructed from hundreds to thousands of eddy realizations consist of the medians rather than the averages because they are less sensitive to occasional outliers in the anomalies.

To investigate the time evolution of bio-optical properties within the cores of eddies, we took the median of all the gridded values for each individual eddy realization within
a radial distance of $L_s/2$ from the eddy centroid. The evolution of the $CHL$ and $C_{phyto}$ within eddies is calculated from weekly medians as a function of eddy age, spanning the range from 1 to 12 weeks. The confidence interval about the median at each weekly average follows the standard formulation: \[ \pm q_t(0.025, N^* - 1)/\sqrt{N^*}, \] where $\sigma$ is the standard deviation of the various bio-optical properties in each weekly average, $N^*$ is the effective number of independent realizations of individual long-lived eddies (lifetimes $\geq 12$ weeks) included in each bin average. The factor $q_t(0.025, N^* - 1)$ is the 2.5% percentage point of the Student’s t random variable with $N^* - 1$ degrees of freedom. The weekly averages and confidence intervals were then interpolated onto a daily grid using a half-power filter cutoff of 2 days.

The time rate of change of $CHL$ and $C_{phyto}$ within eddies is calculated by centered finite differences of the daily means. Forward and backward first differences were used at the start and end of the time series, respectively. The confidence interval for each time derivative was estimated following the standard formula above, but with consideration of error propagation. The time rate of change of each of the bio-optical properties calculated by differencing is a function of two weekly bin averages. The variance of a linear combination of two weekly bin averages: \[ \sigma^2(z) = \sigma^2(x) + \sigma^2(y) + 2cov(x,y) \] where $cov$ is the covariance between $x$ and $y$. Assuming that the variance of any two weekly bin averages are uncorrelated ($cov(x,y) = 0$) and $\sigma^2(x) = \sigma^2(y)$, the variance of $z$ reduces to $\sigma^2(z) = 2\sigma^2(x)$. The standard deviation of $z$ is therefore $\sigma(z) = \sqrt{2} \sigma(x)$. The confidence interval about each estimate of the time rate of change of any particular bio-optical time series thus is \[ \pm \sqrt{2} \sigma q_t(0.025, N^* - 1)/\sqrt{N^*}. \]

### 4.4.2 A Simple NPZ Model for Ecosystems Trapped within Eddies

The phytoplankton populations within eddies in the 5 study regions are observed to oscillate quasi-regularly in time (see sections 4.5.2, 4.5.3 and 4.5.5). In order to investigate the nature of these oscillations, we adapt a simple nutrient (N), phytoplankton (P) and
zooplankton (Z) model to be representative of the ecosystems observed in the interiors of eddies. The details of the NPZ model, along with its initial conditions, are described by Franks et al. (1986). We step the model forward in time using the Euler method, as recommend by Miller (2004). The interactions of the three state variables are defined by:

\[
\frac{dP}{dt} = \frac{\mu NP}{K_s + N} - mP - ZR_m \left(1 - e^{-\lambda P}\right)
\]

(4.1)

\[
\frac{dZ}{dt} = \gamma ZR_m \left(1 - e^{-\lambda P}\right) - kZ
\]

(4.2)

\[
\frac{dN}{dt} = -\frac{\mu NP}{K_s + N} + mP + kZ + (1 - \gamma) ZR_m \left(1 - e^{-\lambda P}\right)
\]

(4.3)

where the state variables, parameters and their values are listed in table 4.2. More details and the justification for each of these interactions are given in both Franks et al. (1986) and Miller (2004).

This NPZ model is strictly time dependent and represents a gross over-simplification of ecosystem dynamics within eddies. The simplicity of this model is nevertheless advantageous as its sole purpose in this study is to test whether we can reproduce oscillatory behavior of phytoplankton communities observed within eddies using parameters that are based on our observations.

4.5. Results

4.5.1 Azimuthal Advection of Phytoplankton by Eddy Currents

The regional variability of the influence of mesoscale eddies on CHL and \(C_{phyto}\) is presented in the sections that follow. We begin by constructing CHL composites of all midlatitude eddies to investigate the dominant mechanism by which eddies influence CHL. This first section is presented as a follow-on study to the work presented by Chelton
et al. (2011a).

Midlatitude eddies are defined in this study as having originated between 15° and 45° latitude in both hemispheres. Histograms of the horizontal scales $L_S$ of these eddies are shown in figure 4.4. The middle 75% of the distribution of eddy scale $L_S$ ranges from 62 to 128 km. On average, midlatitude cyclones are somewhat larger in amplitude than cyclones (averages of 8.0 cm and 7.1 cm, respectively) with a preference for larger-amplitude eddies (amplitude $\geq$ 10 cm) to be cyclonic. This is consistent with the global analysis presented in Chelton et al. (2011b).

The investigation of the influence of midlatitude eddies on phytoplankton by Chelton et al. (2011a) considered anomalies of $\log_{10}$ transformed CHL, primarily because of a historical precedent. This appears to be traceable to Campbell (1995), who found the distribution of CHL to be very skewed towards smaller values. It is shown in appendix D that CHL in the open ocean is more symmetrically distributed than in coastal regions and the log transformation of CHL is therefore not necessary. Moreover, there are some advantages to working with CHL rather than $\log_{10}(CHL)$ (see appendix D). In the present study, we extend the results of Chelton et al. (2011a) to investigate the influence of eddies on non-log transformed CHL.

In their study of midlatitude eddies, Chelton et al. (2011a) showed that anomalies of $\log_{10}$ transformed near-surface chlorophyll $[\log_{10}(CHL)]'$ are dominated by the azimuthal advection of $\log_{10}(CHL)$ around the peripheries of eddies by the rotational eddy currents. A dipole spatial structure of the $[\log_{10}(CHL)]'$ anomalies results from the advection of the ambient $\log_{10}(CHL)$ field around eddies (figure 4.5a-d). This was shown schematically for the advection of the background SST gradient in chapter 2., figure 2.2, which is mechanistically the same for $\log_{10}(CHL)$. The dipole structures of $[\log_{10}(CHL)]'$ consists of anomalies of opposite sign with orientation and magnitude that depend on the direction and magnitude of the background $\log_{10}(CHL)$ gradients and the rotational speed and
direction of the eddy Chelton et al. (2011a).

The previously documented dominance of the azimuthal advection of near-surface $\log_{10}(CHL)$ around midlatitude eddies is also observed in anomalies of non-$\log$ transformed $CHL'$ (figures 4.5e-h). For eddies rotating clockwise in a $CHL$ gradient with a nonzero northward component, the leading (westward) pole has a negative $CHL$ anomaly (figure 4.5e). The northward velocity on the western side of the clockwise-rotating eddy advects low $CHL$ water from the southwestern quadrant to the northwestern quadrant, resulting in a negative $CHL$ anomaly in the northwestern quadrant. The clockwise-rotating surface currents on the trailing edge act to advect relatively high $CHL$ water from the northeastern quadrant to the southeastern quadrant, resulting in a positive $CHL$ anomaly in the southeastern quadrant. The opposite is true for clockwise-rotating eddies propagating in regions with a background $CHL$ gradient having a southward component; in this case, a positive $CHL$ anomaly in the northwestern quadrant is associated with the leading edge of the eddy and a negative $CHL$ anomaly in the southwest quadrant is associated with the trailing edge (figure 4.5g).

For counterclockwise-rotating eddies in a northward $CHL$ gradient, the currents associated with the leading edge of the westward propagating eddy advect high $CHL$ water from the north, thereby creating a positive $CHL$ anomaly in the southwest quadrant of the eddy (figure 4.5f). On the trailing side, the counterclockwise-rotating currents bring low $CHL$ water from the southeastern quadrant into the northeastern quadrant, resulting in a negative $CHL$ anomaly in the northeastern quadrant (figure 4.5f). The opposite is observed for counterclockwise-rotating eddies in a southward $CHL$ gradient (figure 4.5h).

A noteworthy feature of the eddy-induced anomalies in figure 4.5 is that the anomaly on the leading (western) side of westward-propagating eddies is always larger in magnitude than that of the secondary anomaly of the opposite sign on the trailing (eastern) side. This is true for both $[\log_{10}(CHL)]'$ and $CHL'$ (with the exception of clockwise-rotating eddies
in a southward CHL gradient where the magnitudes of the two poles are effectively equal, figure 4.5g). From a model simulation of mesoscale eddies propagating through a passive tracer fluid with a meridional gradient, Chelton et al. (2011a) showed that this structure is to be expected. They suggest that this is a consequence of the trailing side of the eddy encountering a background field that has recently been perturbed by the leading side.

It is thus seen that the signatures of azimuthal advection of the background gradient found by Chelton et al. (2011a) for $\log_{10}(CHL)$ is also clearly evident in the CHL anomalies without the log transformation. The spatial patterns of anomalous CHL within the interiors of midlatitude eddies have essentially the same structures as $\log_{10}(CHL)$, indicative of horizontal advection of ambient CHL by the rotational velocities of the eddies. As is the case of the anomalies of log transformed CHL, the dipole patterns of CHL anomalies result from a combination of the rotational sense of the eddies and the direction of the background CHL gradient.

Also as in the case of anomalies of $\log_{10}(CHL)$, the dipole structure of the CHL anomalies results in a maximum cross correlation between $CHL'$ and SSH occurring at a non-zero time lag. This is because the eddy centroids lag the extremum of $CHL'$ in the westward propagating eddies. A secondary extremum of lagged cross correlation of opposite sign occurred at a lag of opposite sign in association with the secondary extremum of the dipole of $[\log_{10}(CHL)]'$. Chelton et al. (2011a) found that the cross correlation between SSH and $[\log_{10}(CHL)]'$ was maximum at a lag of 4 weeks in the region of the subtropical South Pacific considered in detail in their study. We found that the maximum lagged correlation in this region also occurs at a lag of 4 weeks for the $CHL'$ anomalies. A global map of the cross correlation of SSH and $CHL'$ lagged in time so that $CHL'$ leads SSH by 4 weeks is shown in figure 4.6a. Large zonal swaths of time-lagged correlation magnitudes in excess of 0.2 occur in the midlatitudes where the propagation speeds of the eddies in conjunction with the longitudinal offsets between $CHL'$ and SSH are consistent
with a 4-week lag. These regions of lagged cross correlation are collocated with the regions of the largest meridional $CHL$ gradient (figures 4.6a and b).

In summary, the conclusion of Chelton et al. (2011a) that the statistically dominant mechanism of the azimuthal advection of $\log_{10}(CHL)$ around eddies is confirmed to also be the dominant mechanism by which $CHL'$ responds to midlatitude eddies. Therefore the dominant influence of eddies on phytoplankton is to stir the ambient field, transporting phytoplankton communities around the periphery of eddies. This azimuthal advection generates dipoles of $CHL'$ within the eddy interiors. Because the centroids of westward propagating eddies lag the $CHL'$ extrema, a maximum cross correlation is found at a lag of approximately one month.

### 4.5.2 Mesotrophic boundary current systems eddies

Eddies formed in boundary current regions, including the major eastern boundary currents and most western boundary currents, are generally observed to have concentrations of $CHL$ and $C_{phyto}$ that are elevated within the interiors of cyclones and locally depressed within anticyclones, as illustrated by the negative correlation between SSH and $CHL'$ (figure 4.6c). These anomalies are consistent with the preferential entrainment of eutrophic water into cyclones and oligotrophic water into anticyclones in equatorward-flowing eastern boundary currents and poleward-flowing western boundary currents. The $CHL'$ anomalies in eddies of most boundary currents are also consistent with an ecosystem response to eddy pumping during eddy formation, i.e., the shoaling and deepening of subsurface isopycnals in cyclones and anticyclones, respectively. Satellite observations of eddies formed in a series of three boundary current regions are summarized in this section along with an investigation of how the different aspects of the $CHL'$ and $C_{phyto}'$ are related to the SSH structures of the eddies. We also look at the time evolution of these bio-optical signatures to assess whether the observations support the notion that the photoautotrophic ecosystems in the interior of eddies that form in
these regions are influenced by eddy pumping.

**Gulf Stream Eddies**

The first region investigated is the Gulf Stream in the western North Atlantic. The Gulf Stream (GS) region is defined in this study as: $35^\circ N - 45^\circ N$ and $290^\circ E - 325^\circ E$ (figure 4.8a). The histograms shown in figure 4.8b, along with the composites presented in figures 4.9 and 4.10, are based on only the times when the centroids of these eddies are located within this region.

Eddies formed from the pinching off of meanders of the GS are large in amplitude compared to the average midlatitude eddies (GS eddies have an average amplitude of $24.1 \text{ cm}$, whereas the global average midlatitude eddy amplitude is $7.6 \text{ cm}$). The largest amplitude GS eddies are predominantly cyclonic (figure 4.8b). These cyclones shed from the southern side of the GS and thus entrain waters from the inshore side of the GS that are high in $CHL$ and $C_{phyto}$, compared to the adjacent Sargasso Sea to the south (figure 4.7a). This preferential entrainment of eutrophic shelf water generates positive $CHL'$ and $C'_{phyto}$ in GS cyclones (figure 4.9). In contrast, GS anticyclones entrain nutrient-depleted subtropical waters during formation, resulting in negative $CHL'$ and $C'_{phyto}$ within their interiors (figure 4.9).

Cyclones are observed to have elevated $CHL'$ within their cores throughout most of the year (figure 4.9). In the fall and winter, however, the $CHL'$ are very weak with a slight tendency for dipolar structure with elevated $CHL'$ in the southwestern quadrant. This structure is clearest from September through November (figure 4.10). The $CHL'$ content of anticyclones is approximately a mirror image of cyclones. The dipole structures in fall and winter are consistent with the azimuthal advection of a northward background $CHL$ gradient around GS eddies.

The monthly composites of $C'_{phyto}$ within GS eddies are very similar in structure to
the CHL′ composites with the exception of August and September when \( C_{\text{phyto}}' \) reverses sign, i.e., \( C_{\text{phyto}}' \) becomes elevated in the interiors of GS anticyclones and depressed in GS cyclones. It is not clear why this happens in the late summer.

The evolution of \( CHL \) and \( C_{\text{phyto}} \) within GS eddies during the first 12 weeks of their life histories are shown in figure 4.11. As can be expected from the composite averages in figures 4.9 and 4.10, average values of these two bio-optical properties are higher in the interiors of cyclones than in anticyclones (figures 4.11a-b and d-e). If eddy pumping influences the response of photoautotrophic communities within GS eddies, the time rate of change of \( C_{\text{phyto}} \) should be positive during the growth of cyclones and negative during the growth of anticyclones. GS cyclones have just this relationship, with \( \partial C_{\text{phyto}}/\partial t \) remaining positive until week 7 (figure 4.11f) when GS eddies have completed their initial amplitude growth and eddy pumping is expected to cease (figure 4.3a). The time rate of change of \( C_{\text{phyto}} \) in anticyclones does not support the role of eddy pumping in reducing \( C_{\text{phyto}} \) during formation. It is not clear why negative \( \partial C_{\text{phyto}}/\partial t \) is not observed within GS anticyclones. One potential explanation is that GS anticyclones are smaller than GS cyclones. Downward eddy pumping velocities therefore are not sufficiently large to elicit an observable response of near-surface \( C_{\text{phyto}} \).

The time rate of change of \( CHL \) is somewhat different than that of \( C_{\text{phyto}} \). Initially, \( CHL \) increases in the cores of anticyclones and decreases in the cores of cyclones during the first 3 weeks of eddy formation (figure 4.11c). The time evolution of \( CHL \) after this initial period is in agreement with what would be expected as a result of eddy pumping (figure 4.11c) with positive \( \partial CHL/\partial t \) occurring in cyclones from weeks 3 to 6 and negative \( \partial CHL/\partial t \) in anticyclones roughly during the same time period (figure 4.11c).

The oscillations of both \( \partial C_{\text{phyto}}/\partial t \) and \( \partial CHL/\partial t \) within GS eddies are quite striking and appear to have a fairly regular periodicity of roughly a month (figures 4.11c and f). These oscillations will be explored further in section 4.6.
Brazil-Malvinas Confluence Eddies

The second mesotrophic western boundary current region investigated in this study is the Brazil-Malvinas Confluence (BMC) in the southwestern Atlantic. For this study, we define the BMC region as: $34^\circ S - 50^\circ S$ and $305^\circ E - 330^\circ E$ (figure 4.8a). Animations of the SSH signatures of eddies reveal that they generally form in the poleward-flowing Brazil current after it encounters the Malvinas current and begins to deflect to the east. These eddies subsequently propagate to the south and are advected around the periphery of the Zapiola Gyre (Fu, 2006, and figure 4.8a).

Eddies formed in the BMC, as those formed in the GS, are large compared to the average midlatitude eddies (BMC eddies have average amplitude of 18.9 cm). Composite medians of $CHL'$ and $C_{phyto}'$ within the interiors of BMC eddies are consistent with the entrainment of locally elevated $CHL$ into the interiors of cyclones and low $CHL$ into anticyclones formed in the Brazil Current (figure 4.9). The composites of $CHL'$ and $C_{phyto}'$ in BMC eddies are also consistent with an ecosystem response to eddy pumping during formation (figure 4.9).

The geographical structure of $CHL'$ within both cyclones and anticyclones of the BMC consists of a monopole centered on the eddy centroids throughout the year with maximum anomalies occurring during June through January (figure 4.12). Similarly, the structure of $C_{phyto}'$ in BMC eddies consists of a monopole for both polarities throughout the year with the strongest $C_{phyto}'$ occurring from August through February (figure 4.12).

The time series of $CHL'$ and $C_{phyto}'$ are shown in figure 4.13. Anomalies are persistently positive in the interiors of cyclones and negative in anticyclones (figures 4.13a-b and d-e). While this suggests that the combination of preferential entrainment and eddy pumping forms the observed anomalies, the time series of $\partial C_{phyto}/\partial t$ reveals that the initial evolution of $C_{phyto}$ during the first 3 weeks is not representative of what would be expected from eddy pumping. The time rate of change of $C_{phyto}$ is negative in cyclones
until week 3 when $\partial C_{\text{phyto}}/\partial t$ becomes positive. The $C_{\text{phyto}}$ content of cyclones peaks at week 4. The time rate of change of $C'_{\text{phyto}}$ within anticyclones is slightly positive during the first two weeks and becomes negative at week 2 and remaining negative until week 7 (figure 4.13e). Strong oscillations of $\partial C_{\text{phyto}}/\partial t$ are observed for both polarities.

In contrast to the GS eddies considered above, the evolution of $CHL$ within BMC eddies does not appear to support the role of eddy pumping. Initially, $\partial CHL/\partial t$ is positive in both cyclones and anticyclones. During the first 12 weeks of the eddy life cycle, $\partial CHL/\partial t$ appears to oscillate in phase within BMC eddies of both polarities.

In summary, eddies in the BMC region are characterized by positive anomalies of $CHL'$ and $C'_{\text{phyto}}$ within cyclones and negative anomalies within anticyclones. The evolution of these eddies suggest that the entrainment of mesotrophic Brazil current water into cyclones and oligotrophic water from offshore into anticyclones primes the ecosystems trapped within BMC eddies to have the observed bio-optical characteristics. The influence of eddy pumping is not observed in the time evolution of BMC eddies. This lack of a clearly observable response to eddy pumping could be a result of a maximum response occurring toward the bottom of the euphotic zone, as was observed in a number of in situ studies (Backus et al., 1981; Falkowski et al., 1991; McGillicuddy et al., 2007; Thompson et al., 2007).

**California Current Eddies**

The third and final mesotrophic boundary current evaluated in this study is the California Current system (CCS), an equatorward-flowing eastern boundary current that generates mesoscale eddies that are much smaller in amplitude (an average amplitude of 4.8 cm) than the eddies observed in the GS and BMC. For this study, the CCS system is defined as $45^\circ N - 20^\circ N$ and $230^\circ E - 250^\circ E$ (figure 4.8a).

The CCS is a seasonally productive upwelling system (e.g., Huyer, 1983) that spawns
shallow cyclonic eddies that entrain the high productive waters produced during the summertime upwelling when the shallow equatorward surface current is strongest (and hence most unstable) and coastal waters are rich in nutrients, $CHL$ and $C_{phyto}$. The anticyclones that from in the CCS extend deeper in the water column and are believed to form from the pinching off of meanders of the poleward undercurrent (Simpson and Lynn, 1990; Huyer et al., 1998; Chereskin et al., 2000; CORNUELLE et al., 2000; Jerónimo and Gómez-Valdés, 2007; Kurian et al., 2011). There is a strong eastward gradient of both $CHL$ and $C_{phyto}$ in the CCS region (figures 4.7a and b). Eddies originating in the CCS propagate offshore into the oligotrophic North Pacific gyre. Ecosystems entrained during the formation of cyclones could therefore transport elevated levels of biomass when they propagate to the oligotrophic region offshore.

The composites medians of $CHL'$ within the interiors of CCS cyclones are in agreement with what would be expected from the preferential entrainment of eutrophic coastal waters and eddy pumping during formation (figure 4.9). Elevated $CHL'$ and $C'_{phyto}$ is observed in the interiors of these cyclones. Anticyclones formed in the CCS contain low values of $CHL$ and $C_{phyto}$. The spatial structure of these anomalies appear to be dominated by the advection of the background $CHL$ and $C_{phyto}$ fields around their peripheries. In particular, a dipole geographical structure is observed in the composites of both $CHL$ and $C_{phyto}$ in CCS anticyclones (figure 4.9).

The dipole signature of advection around the eddy periphery is observed in the $CHL$ content of anticyclones throughout the year with a seasonal intensification of the magnitude of the $CHL'$ anomalies during the summer (figure 4.14). Positive $CHL'$ anomalies are observed in the cores of cyclones throughout the year, with an offset of the $CHL'$ to the west of their centroids that is largest in the summer and early fall (June through October). The advection of recently upwelled and highly eutrophic water around cyclones during the summer can account for the observed westward displacement of the maximum $CHL'$ in
cyclones, as well as the accentuated dipole of $CHL'$ in anticyclones. This structure of the composites of $CHL'$ thus consists of a superposition of the advection of water around eddies and the entrainment of coastal waters replete with nutrients and phytoplankton into cyclones, or the entrainment of oligotrophic water from offshore in anticyclones.

During most of the year, phytoplankton biomass ($C_{phyto}$) is elevated above background levels in the interiors of cyclones. In July and August, the advection of the background $C_{phyto}$ gradient around cyclones and the resulting dipole pattern dominates the geographical structure of $C'_{phyto}$ within CCS cyclones. CCS anticyclones advect the background $C'_{phyto}$ gradient around their peripheries during the majority of the year, suggesting that vertical motions occurring in the interiors of CCS anticyclones from eddy pumping (during the growth phase) or eddy-induced Ekman pumping have little effect on $C'_{phyto}$.

Eddies formed in the CCS propagate westward against the background $CHL$ and $C_{phyto}$ gradients. Time series of $CHL$ and $C_{phyto}$ in CCS cyclones reveal that both biological properties decrease rapidly after formation (figures 4.15a and d). No mechanism exists that could sustain the high $CHL$ and $C_{phyto}$ concentrations in CCS cyclones following their formation. Therefore the loss of both of these biological properties in cyclones should come as no surprise. Why this is not observed in cyclones of both the GS and BMC is puzzling. This might be because many of the “eddies” identified in the GS and BMC might in fact be meanders that eventually pinch off into eddies.

Unlike CCS cyclones, CCS anticyclones are not observed to lose $CHL$ and $C_{phyto}$ following formation (figures 4.15a and d). The relatively constant $CHL$ and $C_{phyto}$ values within anticyclones can be explained by CCS anticyclones entraining water from offshore during formation that is subsequently only “stirred” around their peripheries.

The time rate of change of $CHL$ and $C_{phyto}$ is dominated by the loss of initial $CHL$ and $C_{phyto}$ within cyclones (figures 4.15c and f), which is opposite of what is expected
from eddy pumping. It should be noted, however, that any signal attributable to eddy pumping may be weak because of the relatively small amplitudes of CCS eddies, and hence small eddy pumping velocities during formation. CCS anticyclones also have negative $\partial C_{\text{phyto}} / \partial t$ during the first few weeks of their life cycles, becoming positive until week 6 (figure 4.15f). The time rate of change of $CHL$ is somewhat similar to $\partial C_{\text{phyto}} / \partial t$ (figure 4.15c) and also appears to be dominated by the loss of initial $CHL$ in cyclones. The notion that eddy pumping plays an important role in the creation of $CHL$ and $C_{\text{phyto}}$ within CCS eddies during formation is thus not supported for either eddy polarities.

The evolution of both $CHL$ and $C_{\text{phyto}}$ within CCS anticyclones has a peculiar “bump” starting at 6 weeks (figures 4.15a-b and d-e). This might be attributable to an ecosystem response to eddy-induced Ekman pumping. However, composites of CCS anticyclones constructed excluding the first 5 weeks of the eddy life cycles do not support this notion. If this “bump” is in fact a result of eddy-induced Ekman pumping, the resulting increase in both $CHL$ and $C_{\text{phyto}}$ is not great enough to generate positive anomalies of $CHL'$ and $C_{\text{phyto}}'$ within CCS anticyclones (figures 4.15b and e).

4.5.3 Oligotrophic Central Ocean Gyres

In contrast to eddies formed in western boundary currents and equatorward-flowing eastern boundary currents for which high $CHL'$ is found in cyclones, eddies formed in the oligotrophic central oceanic gyres are observed to have elevated $CHL'$ in the interiors of anticyclonic eddies. As was illustrated in section 4.3.1, the direction of the background $CHL$ gradient predetermines whether positive $CHL'$ anomalies are associated with cyclonic or anticyclonic eddies. In the two regions investigated in detail in this section, which are both in the southern hemisphere, the ambient $CHL$ gradient is northward and the surface velocity is westward, resulting in the entrainment of water with locally elevated $CHL$ into anticyclones. This preferential entrainment of water high in $CHL$ into anticyclones results in the SSH and $CHL'$ signatures of eddies being positively
correlated in these two regions (figure 4.6c). An objective of this study is to determine the mechanism that sustains elevated $\text{CHL}'$ in anticyclones following their formation. From the analysis in chapter 3, eddy-induced Ekman upwelling is a possible candidate. Eddy-induced Ekman downwelling likewise suppresses $\text{CHL}$ in the interiors of cyclones (negative SSH anomalies), further reinforcing the observed positive correlation between SSH and $\text{CHL}'$ in oligotrophic central ocean gyres.

South Indian Ocean Eddies

Eddies of the South Indian Ocean (SIO) are large in amplitude (a mean of 7.5 cm) compared with eddies in the equatorward-flowing eastern boundary currents such as the CCS considered in section 4.5.2. The entrainment of high $\text{CHL}$ water preconditions anticyclones to respond to the eddy-induced Ekman upwelling that can then sustain elevated $\text{CHL}'$ within anticyclones throughout their lifetime (chapter 3). Cyclones shed from the Leeuwin Current entrain oligotrophic water from offshore and thus precondition the already low productivity water within their interiors to remain low in response to Ekman downwelling. In this study, we define the SIO as the region: $15^\circ S − 30^\circ S$ and $60^\circ E − 120^\circ E$ (figure 4.8a).

The year-round composites of $\text{CHL}'$ are consistent with the entrainment of high $\text{CHL}$ water into anticyclones and low $\text{CHL}$ water into cyclones as well as with the expected ecosystem response to eddy-induced Ekman pumping (figure 4.9), which on average exceeds $8 \text{ cm day}^{-1}$ of upwelling in anticyclones and $5 \text{ cm day}^{-1}$ of downwelling in cyclones (figure 4.16). The asymmetry of the upwelling and downwelling velocities is a result of the largest SIO eddies being predominantly anticyclonic (see figure 4.8b as well as chapter 3, figure 3.2). The $C'_{\text{phyto}}$ signatures appear to be dominated by the advection of the background $C'_{\text{phyto}}$ gradient around the peripheries of SIO eddies (figure 4.9).

The $\text{CHL}'$ signature of SIO eddies varies seasonally (figure 4.17). SIO anticyclones
have a maximum \(CHL'\) centered on the eddy centroid during May through October (figure 4.17), coincident with the region of maximum eddy-induced Ekman upwelling (figure 4.16). During November through April, the \(CHL'\) composites for SIO anticyclones have the dipole structure expected from azimuthal advection of the background \(CHL\) gradient (figure 4.17). This seasonal variability of the \(CHL'\) signatures of SIO anticyclones appears to be a result of the uncoupling and subsequent recoupling of the mixed layer to the nutricline (chapter 3). SIO cyclones have a similar seasonal variability with a monopole of minimum \(CHL'\) occurring at the eddy centroid during the winter and a dipole structure during the summer. As discussed in chapter 3, the seasonal variability of the magnitude of the \(CHL'\) within cyclones at least partly is an artifact of filtering in combination with the wintertime increases in \(CHL'\) within surrounding anticyclones.

The seasonal variability of \(C'_{phyto}\) within SIO eddies is quite interesting. In anticyclones, \(C'_{phyto}\) anomalies have the dipole structure associated with the advection of the background \(C_{phyto}\) around eddies for most of the year (figure 4.17). From December through February, however, \(C'_{phyto}\) appears to be enhanced in the cores of SIO anticyclones. This is out of phase with the observed \(CHL'\) within the same eddies. The analysis presented in chapter 3 revealed that eddy-induced Ekman upwelling was maximum during these months (see the contours of Ekman pumping in figure 3.4, chapter 3). There is thus an accumulation of \(C_{phyto}\) within the cores of anticyclones during the time period of the strongest upwelling. The \(C_{phyto}\) composites must be interpreted with some caution, however, since they are quite noisy compared with the \(CHL'\) composites (figure 4.17). The dipole structure of the \(C'_{phyto}\) anomalies in cyclones indicates a dominance of advection of the background \(C_{phyto}\) around the periphery of cyclones throughout the year.

Eddies formed in the eastern portion of the SIO region propagate westward into regions of lower \(CHL\) and higher \(C_{phyto}\) (figures 4.7a and b). The time series of both \(CHL\) and \(C_{phyto}\) within SIO eddies reveal that eddies of either polarity lose \(CHL\) and
gain $C_{phyto}$ during at least the first 8 weeks of their life cycles (figures 4.18a and d). This is likely the result of the eddies propagating down the ambient $CHL$ gradient and up the ambient $C_{phyto}$ gradient (figures 4.7a and b). The time series of $CHL'$ within SIO eddies reveals that $CHL'$ remains positive in anticyclones and negative in cyclones throughout the first 12 weeks of the eddy life cycles (figure 4.18b). The time rate of change of $C_{phyto}$ in figure 4.18f indicates that eddies of both polarities gain $C_{phyto}$ during the first few weeks and then begin to oscillate. The evolution of $CHL$ is dominated by the loss of $CHL$ following eddy formation (figure 4.18c). Eddies formed in the southern portion of the Leeuwin Current are highly nonlinear (see figure 17 of Chelton et al., 2011b). Elsewhere in the SIO, eddies are much less nonlinear. In weakly-nonlinear eddies, the region of trapped fluid is small. This could result in the time series of $CHL$ averaged within a distance of $L_s/2$ being influenced by the ambient $CHL$ and $C_{phyto}$ fields.

In summary, SIO eddies are observed to have elevated $CHL$ within the interiors of anticyclones and negative $CHL'$ within the interiors of cyclones during the winter, which is in agreement with an ecosystem response to eddy-induced Ekman pumping (chapter 3). During the summer, eddies of the SIO are observed to advect the background $CHL$ gradient around their peripheries.

*Oligotrophic South Pacific Eddies*

The central South Pacific is one of the most oligotrophic regions of the World Ocean (Behrenfeld et al., 2005). In this study we define the oligotrophic South Pacific gyre (OPAC) as the region: $15^\circ S - 25^\circ S$ and $200^\circ E - 250^\circ E$ (figure 4.8a). OPAC eddies are small (average amplitude of 4.1 cm), and presumably form by baroclinic instability in the ocean interior (Chelton et al., 2011b). The background $CHL$ gradient in the portion of the oligotrophic South Pacific analyzed here is northeastward and the surface flow transitions from slightly westward to eastward in the southern portion of the region.
From the discussion in section 4.3.1, anticyclonic eddies therefore entrain high CHL water and cyclonic eddies entrain low CHL water during formation.

OPAC anticyclones have locally elevated \( CHL' \) within their cores throughout their lifetimes, consistent with the initial entrainment of high \( CHL \) water during formation and subsequent eddy-induced Ekman upwelling bringing nutrients into the euphotic zone from below the nutricline, thereby sustaining the phytoplankton populations (figure 4.9). Conversely, OPAC cyclones have depressed \( CHL' \) through their lifetimes, in agreement with the initial entrainment of low \( CHL \) during formation and subsequent eddy-induced Ekman downwelling transporting nutrients and phytoplankton downward and out of the euphotic zone (figure 4.9). This results in a relatively large region of positive cross-correlation of SSH and \( CHL' \) within the OPAC (figure 4.6a).

In the northeastern portion of the OPAC domain, the cross-correlation between SSH and \( CHL' \) at 4 weeks lag is observed to be in excess of 0.2 (figure 4.6a). This 4-week lag is symptomatic of a dipole structure in \( CHL' \) in the westward propagating eddies (Chelton et al., 2011a). This suggests that the composites of \( CHL' \) might consist of the superposition of the monopole from the combined effects of entrainment of \( CHL \) during formation and Ekman pumping with a dipole structure from the azimuthal advection of \( CHL' \) around the eddy periphery. From June through August, the \( CHL' \) of OPAC anticyclones is monopole structure with maximum \( CHL' \) collocated with the maximum eddy-induced Ekman upwelling velocities at the eddy centroid (figure 4.19). A shift to dipole structure of the \( CHL' \) anomalies of anticyclones occurs from September through May, indicating that azimuthal advection plays the dominate role in the observed \( CHL' \). Cyclones of the OPAC have \( CHL' \) anomalies that indicate the importance of eddy-induced Ekman downwelling throughout the year with a shift to a superposition of a monopole and dipole from September through March (figure 4.19).

The \( C'_{phyto} \) anomalies of OPAC eddies are very noisy (figure 4.19), but appear to
suggest that eddies of the OPAC are dominated by advection of $C_{phyto}$ around their peripheries most of the year, which is clearly seen in the year-round composites in figure 4.9.

The time series of $\text{CHL}$ and $\text{CHL}'$ within OPAC eddies are in agreement with the observed eddy composites; $\text{CHL}'$ anomalies are positive within anticyclones and negative within cyclones during the entire first 12 weeks of the eddy life cycles (figures 4.20b). The magnitudes of $\text{CHL}'$ anomalies in OPAC eddies are very similar to those of SIO eddies (figure 4.18a). Time series of $C_{phyto}$ and $C_{phyto}'$ are noisy and not very informative (figures 4.20d and e), which comes as no surprise considering the level of noise in the $C_{phyto}'$ composites (figure 4.19). The time rate of change of $\text{CHL}$ within OPAC eddies (figures 4.20c and f) displays the same types of oscillations that are observed in eddies in the other regions considered in this chapter, with much smaller amplitudes.

In summary, eddies formed in the OPAC entrain high $\text{CHL}$ water into anticyclones and low $\text{CHL}$ water into cyclones during formation. The $\text{CHL}'$ anomalies of OPAC eddies appear to respond to eddy-induced Ekman pumping in the winter and the azimuthal advection of phytoplankton communities around their peripheries in the summer.

4.5.4 The Relationship Between Chlorophyll and Phytoplankton Carbon Within Eddies

Eddies formed in the two oligotrophic regions investigated in section 4.5.3 can be characterized as having $\text{CHL}'$ composites that vary seasonally from a monopole to a dipole structure (figures 4.17 and 4.19). The monopole structure results from the entrainment of $\text{CHL}$ (and the associated nutrients assimilated into the phytoplankton that generate the observed $\text{CHL}'$) into the eddy interior during formation and a subsequent ecosystem response to eddy-induced Ekman pumping. The dipole structures result from the azimuthal advection of $\text{CHL}$ around the eddy periphery.

Eddies of the SIO and OPAC advect $C_{phyto}$ around their peripheries throughout the
year, suggesting that $C_{phyto}$ is not responding to eddy-induced Ekman pumping. The observed difference in the geographical structures of $CHL'$ and $C'_{phyto}$ anomalies suggest that eddies in these oligotrophic regions influence $CHL$ and $C_{phyto}$ differently.

To investigate whether the different geographical structures of $CHL$ and $C_{phyto}$ composites from eddies in the two oligotrophic regions considered in section 4.5.3 are a property of all eddies formed in oligotrophic regions, we constructed composites segregated by the observed relationship between SSH and $CHL'$.

The first subset of eddies investigated originate in regions where the zero-lag cross correlation between SSH and $CHL'$ is $r \geq 0.2$ (shown in all panels of figure 4.21 as enclosed by the black contours). A common characteristic of these regions is that they have a predominately equatorward ambient $CHL$ gradient (figure 4.21c). Another characteristic of these regions is that they generally have low $CHL$ (figure 4.21b). The average concentrations of all regions enclosed by the $r = 0.2$ contour is only 0.08 $mg \ m^{-3}$. We will therefore refer to the eddies formed in these regions oligotrophic eddies.

Regions of the World Ocean observed to have a negative cross correlation between SSH and $CHL'$ at zero-lag ($r \leq -0.2$) can be classified into two primary categories: (1) Poleward-flowing western and equatorward-flowing eastern boundary currents (figure 4.21b) and (2) regions of the mid-ocean gyres that have a predominate poleward ambient $CHL$ gradient (figure 4.21c). The average $CHL$ concentration in the regions enclosed by the $r = -0.2$ contour is 0.13 $mg \ m^{-3}$. Since this is higher than that of the oligotrophic eddies, but far from being considered eutrophic, we will refer to the eddies formed in these regions as mesotrophic eddies.

The mesotrophic cyclones have monopole $CHL'$ anomalies that are positive and maximum at the eddy centroids (figures 4.22a). Conversely, mesotrophic anticyclones have negative $CHL'$ anomalies. This is consistent with the entrainment of water replete

\footnote{Although positive $C'_{phyto}$ anomalies are observed in the interiors of SIO anticyclones from December through February.}
with phytoplankton into cyclones and low in phytoplankton into anticyclones formed in the poleward-flowing western and equatorward-flowing eastern boundary currents considered in section 4.5.2. The composites of mesotrophic eddies are also consistent with the entrainment of high $CHL$ water into cyclones and low $CHL$ water into anticyclones formed from meanders of zonal currents with poleward $CHL$ gradients, as discussed in section 4.3.1.

The $C'_{phyto}$ anomalies associated with mesotrophic eddies consist of a monopole of positive $C'_{phyto}$ in cyclone and negative $C'_{phyto}$ in anticyclones. Mesotrophic eddies thus affect both $CHL$ and $C_{phyto}$ similarly.

To supplement the $CHL'$ and $C'_{phyto}$ signatures of eddies originating in mesotrophic regions described above, direct estimates of the phytoplankton growth rate $\mu$ within eddies were made as described in section 1.2.4. Individual Argo float profiles were collocated to the interiors of mesoscale eddies and the MLD estimates from each of these profiles were used to compute $\mu$ using equation (1.7). Histograms of the distribution of $\mu$ estimates within the interiors of eddies in mesotrophic regions (figure 4.23a) reveal that there is a slight preference for growth rates in excess of 1 $day^{-1}$ to occur in cyclonic eddies, suggesting that mesotrophic cyclones act to enhance phytoplankton growth rates within their interiors.

Composites of $CHL'$ anomalies in oligotrophic eddies are approximately mirror images of $CHL'$ composites of mesotrophic eddies (figures 4.22). Oligotrophic anticyclones have monopole $CHL'$ anomalies that are positive and maximum near eddy centroids (figure 4.22c). This is consistent with the entrainment of high $CHL$ water into anticyclones and low $CHL$ water into cyclones formed from meanders of zonal currents with equatorward $CHL$ gradients, as discussed in section 4.3.1. These $CHL'$ anomalies are also indicative of an phytoplankton response to eddy-induced Ekman pumping, as shown by the coincidence of the largest magnitude $CHL'$ with the largest Ekman upwelling velocities.
Oligotrophic cyclones have negative $CHL'$ anomaly monopoles collocated with the eddy centroids as a result of the entrainment of low $CHL$ water during formation and the transport of $CHL$ and nutrients down and out of the euphotic zone by Ekman downwelling (figure 4.22c).

In contrast to the mirror imaging of the composites of $CHL'$, the composites of $C'_{phyto}$ for oligotrophic eddies are dramatically different from their counterparts for mesotrophic eddies. The $C'_{phyto}$ signatures of oligotrophic eddies is clearly dominated by the azimuthal advection of $C_{phyto}$ around the peripheries of eddies (figure 4.22d). Oligotrophic eddies evidently affect $CHL$ differently than $C_{phyto}$. One possible mechanism that could result in the lack of biomass accumulation within the cores of oligotrophic anticyclones is the grazing of phytoplankton by zooplankton at rates that prevent the accumulation of biomass within the eddy interiors. It is known that phytoplankton in oligotrophic ecosystems are generally tightly coupled to their grazers (Burkill et al., 1993; Froneman and Perissinotto, 1996). Paterson et al. (2007) investigated the grazing of phytoplankton by zooplankton in a pair of Leeuwin Current eddies of opposite polarities. The anticyclonic eddy contained significantly more zooplankton biomass than the cyclone. The zooplankton community was dominated by dinoflagellates that primarily graze upon diatoms. Estimates of the phytoplankton mortality from grazing by zooplankton in both the cyclone and anticyclone suggested that grazing rates were sufficient to account for all of the primary production measured within the eddies.

A tight coupling between phytoplankton biomass and the grazer community might thus explain why an accumulation of biomass is not observed within oligotrophic anticyclones, even when high growth rates give rise to increased $CHL'$ (figures 4.22c and d). While this presents an interesting hypotheses, it cannot be tested from the satellite data analyzed in this study. Additional in situ measurements will be necessary to test this hypothesis rigorously.
In oligotrophic regions, the distribution of $\mu$ as a function of eddy polarity does not reveal a clear preference for large phytoplankton growth rates in eddies of either polarity (figure 4.23b).

4.5.5 The Observed Oscillations of Phytoplankton Concentration within Eddy Interiors.

Surprisingly consistent oscillations in phytoplankton carbon ($C_{phyto}$) and, to a lesser degree, $CHL$ were observed in eddies originating in both mesotrophic boundary current regions (section 4.5.2) and oligotrophic regions (section 4.5.3). A representative estimate of the period of oscillation of both $\partial C_{phyto}/\partial t$ and $\partial CHL/\partial t$ can be obtained by time-lagged autocorrelations in figure 4.24 by determining the non-zero lag at which the autocorrelation. The oscillatory periods for each of the 5 study regions are reported in table 4.1. The $\sim$ monthly periodicity can be compared with observations of phytoplankton biomass from a moored fluorometer at ocean station Papa in the subarctic Pacific Ocean, which found an oscillation at a period of $\sim$ 30 days (Miller, 2004). Additional observations of $CHL$ and ammonium ($NH_4$) exhibited a similar oscillation with a period of $\sim$ 20 days (Miller et al., 1991). One of the hypotheses proposed to explain these oscillations was predator-prey interactions between grazers and phytoplankton (Miller et al., 1991).

The observed monthly oscillation of $\partial C_{phyto}/\partial t$ and $\partial CHL/\partial t$ within the interiors of mesoscale eddies may arise from zooplankton predation. Since these oscillations are most apparent in GS eddies, we investigate whether they can be reproduced using a simple NPZ model (see section 4.4.2 for a description of the model). The model is initialized with a phytoplankton growth rate of $\mu = 0.7 \text{ day}^{-1}$, which is the average value within all GS eddies. All other initial conditions (see table 4.2) are taken directly from Franks et al. (1986), who developed this model to investigate ecosystem dynamics in idealized GS anticyclones. The resulting N, P and Z concentrations as a function of time are shown in figure 4.25a and the time lagged auto-correlation of P is shown in figure
4.25b. This simple NPZ model has an oscillatory period (∼ 30 days) that is very similar to the observed oscillatory periods of 23-37 days for $CHL$ and $C_{phyto}$ within GS eddies (table 4.1).

It is important to note that this NPZ model is a gross over simplification of ecosystem interaction within eddies. For example, there is no physical forcing in this model and the physical environment is assumed to be in steady state with no fluxes into or out of the model. The initial condition and values of the state variables used in this model might not correctly represent the conditions within eddies, especially since we chose $\mu$ to match our observations. Alterations of these initial conditions can significantly change the results of the model. In light of these shortcoming, and others not mentioned, this simple modeling study does provide qualitative support for the hypothesis that the observed monthly oscillations in $CHL$ and $C_{phyto}$ within eddies could be the result of predator-prey (zooplankton-phytoplankton) interactions. Further investigations would be require test this hypothesis more rigorously.

4.6. Summary and Conclusions

The regional investigation presented in this chapter revealed that cyclones that form in mesotrophic boundary current systems (GS, BMC and CCS) have elevated $CHL$ and $C_{phyto}$ within their interiors (figure 4.9). In western and most eastern boundary current systems, the preferential entrainment of water high in $CHL$ and $C_{phyto}$ primes the interiors of cyclones to have positive $CHL'$ and $C'_{phyto}$. On the other hand, the preferential entrainment of water from offshore primes the interiors of anticyclones to have negative $CHL'$ and $C'_{phyto}$.

In the GS, our observations indicate $C_{phyto}$ within cyclones might also be influenced by eddy pumping during formation (figure 4.11f). This was not observed in eddies of the
BMC or CCS, possibly because of the largest phytoplankton response to eddy pumping occurs at the base of the euphotic zone (Backus et al., 1981; Falkowski et al., 1991; McGillicuddy et al., 2007; Thompson et al., 2007).

The global analysis of mesotrophic eddies (identified as originating in regions of negative cross correlation between SSH and $CHL'$) revealed that $CHL'$, $C'_{phyto}$ and $\mu$ are elevated in cyclones and suppressed in anticyclones. This suggests that mesotrophic eddies have a direct effect on phytoplankton biomass and growth rates within their interiors that is similar to that formed in the boundary currents.

In contrast, observations of eddies formed in the SIO and OPAC are characterized as having elevated $CHL$ within the interiors of anticyclones and suppressed $CHL$ within cyclones. This results in a positive cross correlation coefficient between SSH and $CHL'$ (figure 4.6c). Eddies formed in the SIO and OPAC only advect ambient $C'_{phyto}$ around their peripheries, as is evident from the consistent dipole structure of $C'_{phyto}$ composites.

The observed $CHL'$ anomalies of oligotrophic eddies (identified as originating in regions of positive cross correlation between SSH and $CHL'$) suggest that the preferential entrainment of high $CHL$ water into anticyclones in conjunction with eddy-induced Ekman upwelling sustains high $CHL'$ within anticyclones. An accumulation of biomass ($C_{phyto}$) is not observed. It was suggested that this lack of biomass accumulation within the cores of oligotrophic anticyclones may be an indication of the grazing of phytoplankton by zooplankton at rates that prevent the accumulation of biomass within the eddy interiors.

$CHL$ and $C_{phyto}$ was observed to oscillate with fairly regular periods within the interiors of many eddies. We hypothesize that the oscillatory nature of phytoplankton biomass may be a result of predator-prey interactions between phytoplankton and their grazers. This was tested for GS eddies by forcing a simple NPZ model with observed phytoplankton growth rates, which showed similar oscillations. Despite the gross oversim-
plification of the representation of ecosystem dynamics within eddies in this model, it at least offers qualitative support of our hypotheses.

Satellite observations of near-surface $CHL$ and $C_{phyto}$ only provide a glimpse into the true nature of how mesoscale eddies influence marine ecosystems. These observations nonetheless provide a baseline from which coupled ocean circulation and biogeochemical models could be compared.
4.7. Tables
TABLE 4.1: Estimated oscillatory period (in days) of $\partial C_{phyto}/\partial t$ and $\partial CHL/\partial t$ within eddies of the 5 study regions. Averages shown in bottom row exclude those periods $\geq$ 45 days.

<table>
<thead>
<tr>
<th>region</th>
<th>$\partial C_{phyto}/\partial t$ anticyclones</th>
<th>$\partial C_{phyto}/\partial t$ cyclones</th>
<th>$\partial CHL/\partial t$ anticyclones</th>
<th>$\partial CHL/\partial t$ cyclones</th>
</tr>
</thead>
<tbody>
<tr>
<td>GS</td>
<td>24</td>
<td>23</td>
<td>37</td>
<td>33</td>
</tr>
<tr>
<td>BMC</td>
<td>$&gt; 45$</td>
<td>37</td>
<td>$&gt; 45$</td>
<td>39</td>
</tr>
<tr>
<td>CCS</td>
<td>45</td>
<td>30</td>
<td>22</td>
<td>$&gt; 45$</td>
</tr>
<tr>
<td>SIO</td>
<td>40</td>
<td>41</td>
<td>26</td>
<td>25</td>
</tr>
<tr>
<td>OPAC</td>
<td>24</td>
<td>$&gt; 45$</td>
<td>34</td>
<td>37</td>
</tr>
</tbody>
</table>

average 33 33 27 33
4.8. Figures
TABLE 4.2: Initial condition and values for the Franks et al. (1986) NPZ model.

<table>
<thead>
<tr>
<th>Variable or Parameters</th>
<th>Definition</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\mu$</td>
<td>maximum phytoplankton growth rate</td>
<td>0.7 day$^{-1}$</td>
</tr>
<tr>
<td>$N$</td>
<td>nutrient concentration</td>
<td>starts at 1.6 $\mu$ mol N liter$^{-1}$</td>
</tr>
<tr>
<td>$K_s$</td>
<td>half saturation constant for nutrients</td>
<td>1 $\mu$ mol N liter$^{-1}$</td>
</tr>
<tr>
<td>$P$</td>
<td>phytoplankton concentration</td>
<td>starts at 0.3 $\mu$ mol N liter$^{-1}$</td>
</tr>
<tr>
<td>$m$</td>
<td>phytoplankton mortality rate (apart from grazing)</td>
<td>0.1 day$^{-1}$</td>
</tr>
<tr>
<td>$Z$</td>
<td>zooplankton concentration</td>
<td>starts at 0.1 $\mu$ mol N liter$^{-1}$</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>zooplankton growth efficiency</td>
<td>0.3</td>
</tr>
<tr>
<td>$R_m$</td>
<td>maximum zooplankton ratio</td>
<td>1.5 day$^{-1}$</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>Levlev constant</td>
<td>$1 (\mu$ mol N liter$^{-1})^{-1}$</td>
</tr>
<tr>
<td>$d$</td>
<td>zooplankton mortality rate</td>
<td>0.2 day$^{-1}$</td>
</tr>
</tbody>
</table>
FIGURE 4.1: SeaWiFS CHL image illustrating the preferential entrainment of the ambient CHL gradient into eddies formed in the open ocean. (a) $\log_{10}(CHL)$ overlaid with contours of SSH at an interval of 2 cm starting at $\pm 2$ cm. Positive SSH anomalies are shown as solid contours, negative as dashed contours. (b) CHL anomaly overlaid with SSH contours.
Mechanism by Which Eddies Influence Phytoplankton

FIGURE 4.2: Schematic representation of how eddies influence phytoplankton communities. Red colors represent an increase in phytoplankton, blue a decrease. Eddy stirring resulting from the azimuthal advection of phytoplankton communities around the periphery of eddies for counterclockwise (a) and clockwise (b) rotating eddies in a northward ambient gradient of phytoplankton. The sign, orientation and magnitude of the leading (westward) pole of the phytoplankton anomalies is a function of the rotational sense of the eddy and the direction and magnitude of the background gradient. Eddy-induced Ekman pumping drives downwelling in the cores of cyclones (c) and upwelling in the cores of anticyclones (d) during the entirety of an eddies lifetime. Ekman pumping velocities within eddies are $O(10$ cm day$^{-1}$). Eddy pumping resulting from the baroclinic adjust of isopycnals during eddy formation generates upwelling in the cores of cyclones (e) and downwelling occurs in the cores of anticyclones (f). Eddy pumping only occurs during eddy formation and demise and drives velocities on the order of $1$ m day$^{-1}$.
FIGURE 4.3: Time series of the evolution of eddy amplitude shown as the time rate of change of eddy amplitude. The evolution of the amplitude of anticyclones is shown in red and the negative of the time rate of change of the amplitudes of cyclones is shown in blue. Each panel refers to one of the five study regions shown in figure 4.8a. (a) Gulf Stream eddies. (b) Eddies of the Brazil-Malvinas confluence. (c) Eddies of the California Current. (d) South Indian Ocean eddies. (e) Eddies of the oligotrophic South Pacific.
FIGURE 4.4: (a) Trajectories of midlatitude eddies between 15° and 45° latitude in both hemispheres. Red and blue tracks correspond to cyclones and anticyclones, respectively. (b) Cumulative probability function (top) and histogram (middle) of eddy amplitudes, cyclones in blue and anticyclones in red, and the ratio of cyclonic to anticyclonic histograms (bottom). (c) Same as (b) but for the eddy radial scale $L_S$. (c) Same as (a) except for the rotational speed $U$ of the eddy.
FIGURE 4.5: (a) Composite medians of $\log_{10}(CHL)'$ for clockwise-rotating eddies propagating in a northward background $CHL$ gradient in the upper panel and counterclockwise-rotating eddies in the lower panel. Composites have been horizontally normalized by the eddy scale $L_s$ and rotated to align the background $CHL$ gradient to a polar angle of 90° (see section 4.4.1). (b) Rotated composites of $\log_{10}(CHL)'$ for clockwise-rotating eddies propagating in a southward background $CHL$ gradient in the upper panel and counterclockwise-rotating eddies in the lower panel. (c) Same as (a) but for the non-log transformed $CHL'$. (d) Same as (b) but for $CHL'$. The magnitude ($r$) of the ratio of the primary pole in the leading (left) half of each composite to the secondary pole in the trailing (right) half are labeled on each panel.
FIGURE 4.6: (a) Map of the cross correlation of $CHL'$ and SSH at a time lag of 4 weeks (SSH lags $CHL'$) overlaid with a contour of ±0.2 cross correlation. (b) The median meridional $CHL$ gradient (positive values are northward and negative values are southward). The same contours shown in panel (a) are overlaid on panel (b). (c) Map of the cross correlation of $CHL'$ and SSH at 0 time lag.
FIGURE 4.7: Global maps of $1/4^\circ \times 1/4^\circ$ averages over the 8 + year data record: (a) $\log_{10}(CHL)$ and (b) $\log_{10}(C_{phyto})$ The bounding boxes of each of the 5 study regions are shown in both panels.
a) Track of mesoscale eddies in the study regions

b) Histograms of the bio/physical properties of eddies

**FIGURE 4.8:** (a) Map of the eddies originating in the 5 study regions. Only eddies that are within the bounds of the black boxes are used for eddy statistics and composite medians. (b) Histograms separated by the ecosystem type with mesotrophic boundary currents on the left, enclosed by green border, eddies of the oligotrophic open ocean on the right enclosed by blue border. Histograms of eddy amplitude (cm) in the first row, CHL$'$ (mg m$^{-3}$) in second row and C$'$ phyto (mg m$^{-3}$) in third row.
FIGURE 4.9: Composite medians as a function of polarity for the 5 study regions in each row. The study regions are separated by ecosystem type with eddies of mesotrophic boundary currents enclosed by a green border and overlaid with contours of SSH. Eddies of the oligotrophic open ocean enclosed by a blue border and overlaid with contours of Ekman pumping. The first columns are the composite medians of $CHL'$ for anticyclones (AC) and cyclones (CC) and the second are the composite medians of $C_{phyto}'$ for anticyclones (AC) and cyclones (CC). Note that the colorbar range is different for the composites of each variable and between regions.
FIGURE 4.10: Monthly composites medians of $\text{CHL}'$ and $C'_{\text{phyto}}$ as a function of polarity for eddies originating and propagating in the Gulf Stream region as shown on figure 4.8. Contours are of the same composites shown in the pseudo-color images.
FIGURE 4.11: The evolution of bio-optical properties within the interiors of Gulf Stream eddies. (a) CHL within eddy core (distance ≤ L_s/2 from eddy centroid) bin averaged by eddy age (in weeks). (b) CHL' bin averaged by eddy age. (c) Time rate of change of CHL (∂CHL/∂t) within eddy cores in units µg m^{-3} per week (d) Same as (a) but with bin averages of C_{phyto}. (e) Same as (b) but with bin averages of C_{phyto}'. (f) Same as (c), except ∂C_{phyto}/∂t in units of mg m^{-3} per week.
FIGURE 4.12: Monthly composites medians of $CHL'$ in the left panel and $C'_{phyto}$ in the right panel for anticyclones (AC) and cyclones (CC) originating and propagating in the Brazil-Malvinas Confluence as shown on figure 4.8a. Contours are of the same composites shown in the pseudo-color images.
FIGURE 4.13: The evolution of bio-optical properties within the interiors of Brazil-Malvinas Confluence eddies. (a) CHL within eddy core (distance ≤ \(L_s/2\) from eddy centroid) bin averaged by eddy age (in weeks). (b) \(CHL'\) bin averaged by eddy age. (c) Time rate of change of \(CHL\) (\(\partial CHL/\partial t\)) within eddy cores in units \(\mu g\ m^{-3}\) per week (d) Same as (a) but with bin averages of \(C_{phyto}'\). (e) Same as (b) but with bin averages of \(C_{phyto}'\). (f) Same as (c), except \(\partial C_{phyto}/\partial t\) in units of \(mg\ m^{-3}\) per week.
FIGURE 4.14: Monthly composites medians of $CHL'$ in the left panel and $C_{phyto}'$ in the right panel for anticyclones (AC) and cyclones (CC) originating and propagating in the Californian Current system as shown on figure 4.8a. Contours are of the same composites shown in the pseudo-color images.
FIGURE 4.15: The evolution of bio-optical properties within the interiors of California Current system eddies. (a) CHL within eddy core (distance ≤ $L_s/2$ from eddy centroid) bin averaged by eddy age (in weeks). (b) CHL' bin averaged by eddy age. (c) Time rate of change of CHL ($\partial$CHL/$\partial t$) within eddy cores in units $\mu g\ m^{-3}\ per\ week$ (d) Same as (a) but with bin averages of $C_{phyto}'$. (e) Same as (b) but with bin averages of $C_{phyto}'$. (f) Same as (c), except $\partial C_{phyto}/\partial t$ in units of $mg\ m^{-3}\ per\ week$. 
FIGURE 4.16: Composite medians as a function of polarity for the 5 study regions in each row. The study regions are separated by ecosystem type with eddies of mesotrophic boundary currents enclosed by a green border and eddies of the oligotrophic open ocean enclosed by a blue border. The first set of columns are the composite medians of SSH for anticyclones (AC) and cyclones (CC). The second set of columns are the composite medians of eddy-induced Ekman pumping for anticyclones (AC) and cyclones (CC).
FIGURE 4.17: Monthly composites medians of $CHL'$ in the left panel and $C_{phyto}'$ in the right panel for anticyclones (AC) and cyclones (CC) originating and propagating in the South Indian Ocean as shown on figure 4.8a. Contours are of the same composites shown in the pseudo-color images.
FIGURE 4.18: The evolution of bio-optical properties within the interiors of South Indian Ocean eddies. (a) CHL within eddy core (distance ≤ L_s/2 from eddy centroid) bin averaged by eddy age (in weeks). (b) CHL' bin averaged by eddy age. (c) Time rate of change of CHL (∂CHL/∂t) within eddy cores in units µg m⁻³ per week (d) Same as (a) but with bin averages of C'_{phyto}. (e) Same as (b) but with bin averages of C'_{phyto}. (f) Same as (c), except ∂C_{phyto}/∂t in units of mg m⁻³ per week.
FIGURE 4.19: Monthly composites medians of $CHL'$ in the left panel and $C_{phyto}'$ in the right panel for anticyclones (AC) and cyclones (CC) originating and propagating in the oligotrophic South Pacific, as shown on figure 4.8a. Contours are of the same composites shown in the pseudo-color images.
FIGURE 4.20: The evolution of bio-optical properties within the interiors of oligotrophic South Pacific eddies. (a) CHL within eddy core (distance ≤ $L_s/2$ from eddy centroid) bin averaged by eddy age (in weeks). (b) CHL’ bin averaged by eddy age. (c) Time rate of change of CHL ($\partial$CHL/$\partial t$) within eddy cores in units $\mu g \: m^{-3}$ per week (d) Same as (a) but with bin averages of $C_{phyto}$. (e) Same as (b) but with bin averages of $C_{phyto}'$. (f) Same as (c), except $\partial C_{phyto}/\partial t$ in units of $mg \: m^{-3}$ per week.
FIGURE 4.21: (a) Map of the cross correlation of \( CHL \) and SSH at 0 time lag overlaid with contours of 0.2 cross correlation in black and -0.2 in white. (b) Mean \( CHL \) concentration from SeaWiFS overlaid with the same contours as in panel (a). (c) The median meridional \( CHL \) gradient (positive values are northward and negative values are a southward) overlaid with the same contours as in panel (a).
FIGURE 4.22: Composite medians of eddies selected by the cross correlation of SSH and CHL at 0 time lag. The composites shown in the left panels, enclosed by a green rectangle, are of eddies that are in regions where the cross correlation coefficient is $\leq -0.2$, generally representative of mesotrophic regions. (a) $CHL'$ and (b) $C'_{\text{phyto}}$ overlaid with contours of SSH at an interval of 3 cm. The composite shown in the right panels, enclosed by a blue rectangle, are of eddies that are in regions where the cross correlation coefficient is $\geq 0.2$, which are generally oligotrophic regions. (c) $CHL'$ and (d) $C'_{\text{phyto}}$ overlaid with contours of Ekman pumping velocities at an interval of 1 cm day$^{-1}$. 

Mesotrophic ecosystems ($r\leq -0.2$)

Oligotrophic ecosystems ($r\geq 0.2$)
FIGURE 4.23: Statistical distribution of phytoplankton growth rate $\mu$ estimates made within the interiors of cyclones (blue) and anticyclones (red). (a) Cumulative probability function (top) and histogram (middle) of $\mu$, cyclones in blue and anticyclones in red, and the ratio of cyclones to anticyclones (bottom) for eddies in mesotrophic regions. (b) Same as (a) but for eddies in oligotrophic regions.
Auto Correlation of Time Rate of Change of Bio-optical Properties Within Eddies

FIGURE 4.24: Autocorrelation of (a) $\partial C_{phyto}/\partial t$ within the interiors of anticyclones and (b) cyclones, (c) $\partial CHL/\partial t$ in anticyclones and (d) cyclones originating in the 5 study regions.
a) N, P and Z from modified Franks et al., 1987 NPZ model

b) auto-correlation of phytoplankton biomass (P)

FIGURE 4.25: (a) Nutrient (N), phytoplankton (P) and zooplankton (Z) concentrations (in $\mu$ mol N liter$^{-1}$) from the modified Franks et al. (1986) NPZ model. See table 4.2 for values used to initiate model. (b) Autocorrelation of P. Vertical line demarcates 30 days, the averaged observed oscillatory period of $CHL$ and $C_{phyto}$ in the 7 study regions (table 4.1).
5. GENERAL CONCLUSIONS

The work presented in this dissertation investigated the observed influence of mesoscale ocean eddies on SST, surface stress and near-surface phytoplankton. Eddy effects were isolated by collocating temporally and spatially high-pass filtered observations to the interiors of eddies identified and tracked in SSH fields.

The primary conclusions of the analysis presented in this dissertation can be summarized in four parts:

1. The vorticity of surface currents in mesoscale eddies generate upwelling in the cores of anticyclones and downwelling in the cores of cyclones. Eddy-induced SST anomalies also generate Ekman pumping through air-sea interaction in which SST modifies the overlying wind field. The geographical structure of eddy-induced SST anomalies generate a curl of the wind stress that in turn generates Ekman pumping. This SST-induced Ekman pumping is a function of both the geographical structure of the SST signatures of the eddies and the wind direction. SST-induced Ekman pumping can augment the current-induced Ekman pumping either constructively or destructively, depending on the magnitudes of the SST anomalies and the magnitude and direction of the ambient wind. Our observations conclude that the influence of eddy surface currents and eddy-induced SST perturbations cannot be neglected from ocean circulation models that attempt to reproduce the observed properties of mesoscale eddies.

2. Our analysis of eddies in the South Indian Ocean indicates that eddy-induced Ekman upwelling in anticyclones is capable of sustaining positive CHL' anomalies throughout their life cycles. The response of CHL to eddy-induced Ekman upwelling is seasonal. Analysis of climatological potential density and nitrate concentrations in
the South Indian Ocean suggest that seasonal coupling and decoupling of the mixed layer to the nutricline is the likely mechanism for in the observed summertime disappearance of monopole-like CHL anomalies coincident with maximum eddy-induced Ekman upwelling in anticyclones. In situ profiles of density and nutrients are not sufficient to test this hypothesis rigorously. The seasonal coupling and decoupling of the mixed layer to the nutricline could be further explored in coupled eddy-resolving biogeochemical models.

3. Analysis of global composite of CHL anomalies showed that azimuthal advection around eddy peripheries is the statistically dominant mechanism by which midlatitude eddies influence the CHL in their interiors. This supports the interpretation of Chelton et al. (2011a) which investigated anomalies of \( \log_{10} \) transformed CHL.

4. The direction of the eddy-generating currents in conjunction with the ambient gradient of phytoplankton concentration predetermine whether high concentrations of phytoplankton are observed in cyclones or anticyclones. Cyclonic eddies formed from western and most eastern boundary currents entrain water from the continental shelf with high concentrations of phytoplankton. Conversely, anticyclones formed in these classic boundary current systems entrain water from offshore that has low concentrations of phytoplankton. Both CHL and \( C_{phyto} \) were observed to be elevated within cyclones and depressed in anticyclones within these regions.

In contrast to eddies formed in the classic boundary current systems, those formed in the Leeuwin Current (a poleward-flowing eastern boundary current) and in the oligotrophic mid-ocean gyres with equatorward ambient CHL gradients entrain high CHL water into anticyclones and low CHL in cyclones. Unlike eddies originating in western and most eastern boundary currents, however, an accumulation of biomass \( (C_{phyto}) \) is not observed in eddies formed in oligotrophic regions. It was suggested that this lack of biomass accumulation within the cores of oligotrophic anticyclones
may be an indication of the grazing of phytoplankton by zooplankton at rates that prevent the accumulation of biomass within the eddy interiors. Additional in situ measurements would be necessary to test this hypothesis rigorously.

The observational results presented in this dissertation motivate future studies from eddy-resolving, coupled bio/physical models and in situ observations. Such studies will yield more definitive conclusions about exactly how mesoscale eddies influence marine ecosystems globally.
APPENDICES
A APPENDIX  Estimation of Surface Current-Induced Ekman Pumping

The surface current-induced Ekman pumping associated with the mesoscale eddies was estimated from the geostrophic surface currents computed from the AVISO SSH fields. Because of the nonlinear relation (1.2) between wind stress and the relative wind (1.1), surface currents induce Ekman pumping depends on the large-scale background wind field. This was obtained by smoothing the QuikSCAT wind fields to remove variability with wavelength scales shorter than $6^\circ$. The resultant relative wind was then obtained by subtracting the geostrophic surface currents from the large-scale wind fields. This surface wind stress was estimated from the relative wind by (1.2) and the Ekman pumping (1.3) was estimated using centered finite differencing. The resulting current-induced Ekman pumping was then spatially high-pass filtered in the same manner as the QuickSCAT wind fields (see section 2.4.).
B APPENDIX  Estimation of SST-Induced Ekman Pumping

To estimate the SST-induced Ekman pumping fields, we utilized the empirical linear relationship between the perturbation wind stress curl $\nabla \times \tau'$ and the crosswind and downwind components of the SST gradients (O’Neill et al., 2012):

$$\nabla \times \tau'_{SST} = -\alpha_{strcrl}^{c} \left( \frac{\partial T}{\partial n} \right)' - \alpha_{strcrl}^{d} \left( \frac{\partial T}{\partial s} \right)'$$  \hspace{1cm} (B.1)

where $\alpha_{strcrl}^{c}$ and $\alpha_{strcrl}^{d}$ are the coupling coefficients and $\frac{\partial T}{\partial n}$ and $\frac{\partial T}{\partial s}$ are the crosswind and downwind SST gradients, respectively. The primes denote spatial high-pass filtering (see section ??). The perturbation wind stress curl and divergence, as well as the perturbation wind vorticity and divergence fields, are linearly related to the crosswind and downwind SST gradients (Chelton et al., 2004; O’Neill et al., 2012, and references therein). The crosswind and downwind SST gradients are calculated from the meridional ($\frac{\partial T}{\partial y}$) and zonal ($\frac{\partial T}{\partial x}$) SST gradients by:

$$\frac{\partial T}{\partial n} = -\sin \psi \frac{\partial T}{\partial x} + \cos \psi \frac{\partial T}{\partial y}$$  \hspace{1cm} (B.2)

$$\frac{\partial T}{\partial s} = \cos \psi \frac{\partial T}{\partial x} + \sin \psi \frac{\partial T}{\partial y}$$  \hspace{1cm} (B.3)

$n$ and $s$ are the local crosswind and downwind spatial coordinates, respectively and $\psi$ is the counterclockwise wind direction estimated from individual orbital passes of QuikSCAT. The resultant crosswind and downwind SST gradients were filtered in space and time in the same manner as the SST and wind fields (see section 1.2.2). O’Neill et al. (2012) showed that the coupling between the perturbation wind stress curl and SST is varies both spatially and temporally, depending primarily on the magnitude of the background wind.
stress field. To construct a SST-induced wind stress curl field, we calculated a monthly climatology of the coupling coefficients $\alpha^\text{strcrl}_c$ and $\alpha^\text{strcrl}_d$ at each $\frac{1}{4}^\circ$ grid point for the 7.5 year study period by the regression relations:

$$
\alpha^\text{strcrl}_c = \rho_c \frac{\sigma^\text{strcrl}}{\sigma_c}
$$

$$
\alpha^\text{strcrl}_d = \rho_d \frac{\sigma^\text{strcrl}}{\sigma_d}
$$

where $\rho_c$ and $\rho_d$ are the cross-correlations between the perturbation wind stress curl and the cross- and downwind SST gradients, respectively, $\sigma^\text{strcrl}$, $\sigma_c$ and $\sigma_d$ are the standard deviations of the perturbation wind stress curl, cross- and downwind SST gradients, respectively. To reduce the noise in the estimation of the coupling coefficients, the cross-correlation and standard deviations were computed spatially from weekly averages for each calendar month in overlapping $4^\circ$ box centered on each $\frac{1}{4}^\circ$ grid point.
C APPENDIX The Asymmetry of Eddy SST and Wind Speed Anomalies

While the composite medians of SST within midlatitude eddies (figures 2.7a and b) show cold and warm interiors in cyclones and anticyclones, respectively, the common notion of “cold-core cyclones” and “warm-core anticyclones” is not entirely supported. In particular, the centers of the cold and warm SST anomalies are displaced westward and either northward or southward relative to the eddy centroids. Moreover, they are paired with an SST anomalies of opposite sign and much smaller amplitude on the opposite side of the eddy interiors. It is shown in this appendix that this asymmetric dipole structure is consistent with an interpretation as a superposition of a core of trapped cold or warm water and azimuthal advection of an ambient SST field by the rotational velocities within eddy interiors.

As described in section 2.4.2, the composites presented in figure 2.3 are segregated by the rotational sense of the eddy (clockwise or counterclockwise) and by direction of the background SST gradient. The large-scale SST gradients are predominately southward in the northern hemisphere and northward in the southern hemisphere. This hemispheric bias in the direction of the background SST gradient by hemisphere is reflected in the fact that approximately 99% of the midlatitude eddies in a southward SST gradient are in the northern hemisphere. Likewise, approximately 99% of the midlatitude eddies in a northward gradient are in the southern hemisphere.

The effect of horizontal advection of a tracer field by mesoscale eddies has been investigated by Chelton et al. (2011a) from a quasigeostrophic numerical model. When there is a large-scale gradient of the tracer field, the rotational velocities of the eddy perturbs the background tracer field resulting in an asymmetric dipole anomaly of the tracer within the eddy interior. The degree of asymmetry and the location of each pole depends on the intensity of the eddy. Typically, the pole on the leading edge of the
westward-propagating eddies has a magnitude that is 25 – 50% stronger than the pole of opposite sign on the trailing (eastward) edge and is centered somewhat closer to the eddy centroid. The center of the primary pole is northward of due west of the eddy centroid for clockwise-rotating eddies and southward of due west for counterclockwise-rotating eddies. The center of the secondary pole is opposite the eddy centroid from the center of the primary pole.

The characteristics of eddy-induced anomalies of oceanic chlorophyll are well represented by the dipole structure of the tracer field in the model (Chelton et al., 2011a). On time scale longer than 2-3 weeks, the chlorophyll field thus behaves essentially as a passive tracer that is advected horizontally by the rotational velocities of the eddies. The SST field presumably responds similarly to horizontal advection by the rotational velocities in the eddy interior. Indeed, a dipole structure is readily apparent from the composites of SST in figures 2.7a and b. The asymmetries of the SST dipoles are much larger, however, with ratios of 2.1 to 2.6 compared with ratios of 1.3 to 1.7 found by Chelton et al. (2011a) for chlorophyll and the tracer field in the numerical model.

The greater asymmetry of the SST dipoles are likely related to the fact that the SST field is also influence by the structures of isopycnals in the interiors of the eddies. In the case of cyclones, the negative SSH anomalies in the eddy interiors is associated with a doming of subsurface isopycnals and hence a cold core in the interiors of the eddy. For anticyclones, the positive SSH anomaly in the eddy interiors is associated with depress subsurface isopycnals and hence a warm core.

The hypothesis that the greater asymmetry of the SST dipoles is a result of the superposition of dipole from rotational advection and a monopole SST anomaly from cold or warm core can be tested from consideration of a simple simulated SST composite. A scaled version of the dipole with an asymmetry ratio of 1.5 (Chelton et al., 2011a) is shown in figure 0.1a, combined with the Gaussian shaped SST monopole in figure 0.1b). The
resultant SST composite (figure 0.1c) is very similar to the midlatitude SST composite median shown in the top panel of figure 2.3b.
FIGURE 0.1: Simulated composite median of SST for counterclockwise-rotating eddies in a southward background SST gradient. a) Eddy composite of asymmetric dipole with asymmetry ratio $r = 1.5$ scaled to have a maximum amplitude of $0.3^\circ C$. b) Gaussian SST anomaly scaled to have a maximum amplitude of $0.3^\circ C$ at the eddy centroid. c) The summation of panels (a) and (b). The x and y coordinates of the composite medians are normalized by the eddy scale $L_s$. Note the different colorbar scaling in panel (c) compared with (a) and (b).
When working with bio-optical observations such as CHL or \( C_{phyto} \) concentrations, the characteristics of the distribution of the observations must be considered before spatially and temporally filtering, constructing averages, calculating statistical properties, or testing hypotheses. In satellite based studies of bio-optical properties, a historic precedent exists to analyze log\(_{10}\) transformed \( CHL \) (e.g., Uz et al., 2001; Cipollini et al., 2001; Uz and Yoder, 2004; Siegel et al., 2007; Killworth et al., 2004; Gutknecht et al., 2010; Siegel et al., 2011; Chelton et al., 2011a). This precedent appears to be traceable to an early analysis of in situ observations and individual scenes from the Costal Zone Color Scanner (CZCS) by Campbell (1995). Our analysis focused on the mesoscale variability of \( CHL \) and utilizes regional satellite observations compiled over nearly 8 years. To investigate the question of whether or not to log transform the \( CHL \) fields in our study, we examine in this appendix the distribution of \( CHL \) observations with and without log\(_{10}\) transformation in two different locations within the South Indian Ocean. One of these is coastal and the other in the oligotrophic SIO (see map inset in figure 0.2a).

The \( CHL \) values in the Leeuwin Current region of formation have a long-tailed distribution (green curves in figure 0.2b) that has been described in the past as being “log-normal” (Campbell, 1995). Approximately 40% of the \( CHL \) values exceed 0.15 \( mg \ m^{-3} \). The focus of this study is on the influence of Ekman pumping on ecosystems trapped within the interiors of long-lived mesoscale eddies during their transit across the open ocean, long after their formation (figure 0.2a). We therefore look at the distribution of \( CHL \) at a point in the middle of the South Indian Ocean. At this offshore location, \( CHL \) is more symmetrically distributed with a probability density function resembles a Gaussian, but is clearly skewed slightly towards lower \( CHL \) values (blue curve in figure 0.2a). Although the \( CHL \) observations at this offshore location are not truly normally
distributed, the investigation of non-log\(_{10}\) transformed \(CHL\) anomalies has merit as it allows for comparisons of the results presented in this study to ship based observations discussed in section 3.5..

To ensure that the conclusions of the analysis in this study are not dependent on our choice of analyzing anomalies of the raw \(CHL\) data, we repeated our analysis using anomalies calculated from the \(\log_{10}\) transformed \(CHL\) data. The \(CHL\) observations were first \(\log_{10}\) transform and then smoothed in space as described in section 1.2.3 to calculate the large-scale \(\log_{10}\)-transformed \(CHL\ \log_{10}(CHL)\). The anomalies fields are defined to be

\[
[\log_{10}(CHL)]' = \log_{10}(CHL) - \overline{\log_{10}(CHL)}.
\]  

Our results were found to be insensitive to the choice of \(CHL'\) or \([\log_{10}(CHL)]'\). For example, one of the primary conclusions of this study is that the structure of \(CHL\) anomalies in the interiors of mesoscale SIO eddies varies seasonally. In January, anticyclonic eddies are observed to advect \(CHL\) around their periphery (figure 3.4). The same pattern is observed in the anomaly \(\log_{10}\)-transformed fields (figure 0.3). In June, when Ekman upwelling is hypothesized to stimulate \(CHL\) growth, the anomalies of \(CHL\) and \(\log_{10}(CHL)\) both consist of monopole structures centered within the regions of strong Ekman upwelling over the interiors of anticyclones (figures 3.4 and 0.3).
FIGURE 0.2: (a) Histogram of time series of SeaWiFS CHL values at 2 different locations: coastal (25°S, 114°E) shown as the green histogram and an offshore location (25°S, 81°E) shown as the blue histogram. The inset map shows the locations of the two time-series. (b) The same as a, except for log_{10}(CHL).
Monthly composite medians of $\log_{10}(\text{CHL})$ anomalies

FIGURE 0.3: Monthly composite medians of $[\log_{10}(\text{CHL})]'$ anomalies overlaid with contours of eddy-induced Ekman pumping velocity with a contour interval of 1 cm day$^{-1}$ and negative velocities shown as dashed lines. The title of each panel shows the number of eddy realizations used in the composite for each respective month.
In this study, we have suggested that most of the eddies in the SIO originate from the pinching off of meanders of the Leeuwin Current. To investigate the impact of this assumption on our results, we analyzed a subset of eddies that were observed to form in the Leeuwin Current.

This subset was defined to be eddies that originated in the region $35^\circ S - 21^\circ S$ and $108^\circ E - 117^\circ E$ (figure 0.4a). This resulted in nearly 205 cyclonic and 149 anticyclonic eddies with lifetimes of 12 weeks and longer consisting of 8,765 and 4,965 weekly realizations, receptively, over the nearly 8 year study period (January 2001 through December 2007). Composite medians of these Leeuwin Current eddies are shown in figure 0.5. The average amplitudes are $11.1 \, cm$ and $10.7 \, cm$ for cyclones and anticyclones, respectively, with average radial scales $L_s$ of $80.7 \, km$ and $96.4 \, km$ (figure 0.4b and c). Although cyclones on average are larger in amplitude and smaller in horizontal scale (which would imply faster average rotational speeds $U$, all other things being equal), the largest amplitude eddies ($amp \geq 30 \, cm$) are nearly all anticyclones (figure 0.4a). This results in the average $U$ of anticyclones being $25.8 \, cm \, s^{-1}$ which is $1.4 \, cm \, s^{-1}$ faster than for cyclones (figure 0.4d). The average eddy-induced Ekman pumping in anticyclones spawned from the Leeuwin Current is $7.8 \, cm \, day^{-1}$ and the average Ekman downwelling in cyclones is $-5.9 \, cm \, day^{-1}$.

The background average CHL gradient across is the Leeuwin Current is dominantly zonal, with CHL decreasing westward from the continent (figure 3.3b). This strong zonal $CHL$ gradient in this region, where eddies form from the pinching off of meanders of the Leeuwin Current is manifest as a dipole structure of the composite median of $CHL'$ (figure 0.5, see Chelton et al. (2011a) for an overview of how azimuthal advection of $CHL$ results in a dipole structure in the composites). The extrema of the dipoles of $CHL'$ are located in
the northwest and southeast quadrants (figure 0.5). This differs from the open-ocean SIO eddies for which the dipole extrema are located in the southwest and northeast quadrants (figure 3.4). The dipole structure of CHL' in Leeuwin Current eddies from advection of the zonal background CHL gradient dominates the composites of CHL' during the Austral summer (December through March, figure 0.5). As winter approaches, CHL' in the interior of Leeuwin Current anticyclones begins to respond to Ekman upwelling, with maximum CHL' occurring closer to the regions of maximum Ekman upwelling velocity, but always displace somewhat to the northwest nor north (figure 0.5).

The offset between the extrema of eddy-induced Ekman upwelling and CHL' is symptomatic of a superposition of a dipole structure from horizontal advection and a monopole structure collocated with contours of Ekman pumping velocity. This is shown schematically by combing a dipole CHL anomaly (figure 0.6a) with a monopole structure (figure 0.6b), resulting in a simulated composite CHL anomaly (figure 0.6c) that is qualitatively similar to what is observed in Leeuwin Current eddies (figure 0.5).

In cyclones, the lowest CHL' in winter is also located close to the region of maximum Ekman downwelling. The offset is again symptomatic of the effects of advection of the zonal background CHL gradient superimposed on the monopole structure associated with Ekman downwelling (figure 0.6).

The composites of CHL' of Leeuwin Current eddies in figure 0.5 are qualitatively similar to those of SIO eddies in figure 3.4. Leeuwin Current eddies have larger CHL' when compared to SIO eddies (note the different dynamic ranges of the color bars in figures 3.4 and 0.5) which is expected since the CHL content of anticyclones decreases rapidly as they propagate westward out of the Leeuwin Current region of formation (figure 0.7). Even with the observed difference in the magnitudes of the CHL anomalies and the different locations of the dipole contribution to the CHL anomalies, the physical mechanism for the monopole contribution to the CHL in the interiors of eddies in both the Leeuwin Current
and SIO is the same eddy surface-current Induced Ekman pumping. Ekman upwelling drives increased $CHL'$ during the winter in anticyclones and downwelling drives decreased $CHL'$ in cyclones throughout the year. As summarized above, this is augmented by the advection of the background $CHL$ around the eddy periphery in eddies of both polarities, most noticeably during the Austral summer.

As discussed in section 3.4., for this study we selected eddies based on the cross-correlation of $CHL'$ and eddy-induced Ekman pumping. A defining characteristic of eddies formed in the Leeuwin Current is the association of positive $CHL'$ with anticyclones. This results in a positive cross-correlation between $CHL'$ and eddy-induced Ekman pumping in eddy interiors. We conclude from the investigation of Leeuwin Current eddies presented in this appendix that our criteria of selecting eddies in the SIO based on a positive correlation $\geq 0.2$ between their $CHL'$ and eddy-induced Ekman pumping signatures aids in the identification of eddies that form from the Leeuwin Current. This results in nearly twice as many eddy realizations compared with the restricted set of eddies that from in the region shown in figure 0.4a.
FIGURE 0.4: (a) Trajectories of eddies originating in the Leeuwin Current, defined as the region 35°S – 21°S and 108°E – 117°E. Cyclones and anticyclones in blue and red, respectively. The start location of each eddy is shown as a black circle. (b) Histograms of eddy amplitude with cyclones shown in blue and anticyclones in red. (c) The ratio of the numbers of cyclones to anticyclones as a function of eddy amplitude (b). (d) and (e) are the same as (b) and (c) except for the eddy radius scale $L_s$. (f) and (g) are the same as (b) and (c) except for the rotational speed $U$. 
FIGURE 0.5: Monthly composite medians of CHL' anomalies of eddies originating in the Leeuwin Current (figure 0.4a) overlaid with contours of eddy-induced Ekman pumping velocity with a contour interval of 1 cm day$^{-1}$ and negative velocities shown as dashed lines. The title of each panel shows the number of eddy realizations used in the composite for each respective month.
FIGURE 0.6: Simulated composite median of CHL consisting of the superposition of a dipole CHL anomaly (a) with a monopole (b) resulting in a composite (c) that is qualitatively similar to observed CHL' composites of Leeuwin Current eddies (figure 0.5).
FIGURE 0.7: True color image of Western Australia and the Leeuwin Current. The region of the image over ocean is overlaid with pseudo color image of CHL. Data collected by SeaWiFS on 5 April, 2002. Image courtesy of NASA/GSFC, all rights reserved.
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