

**THE FLOW SEPARATIONS IN THE TAIWAN STRAIT AND OCEAN
RESPONSES TO THE “HIATUS” OF THE GLOBAL MEAN SURFACE
TEMPERATURE**

by

Enhui Liao

A dissertation submitted to the Faculty of the University of Delaware in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Oceanography

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ABSTRACT

Coastlines are fundamental to humans for habitation, commerce, and natural resources. Many coastal ecosystem disasters, caused by extreme sea surface temperature (SST), were reported when the global climate shifted from global warming to global surface warming hiatus between 1998 and 2013. The 2008 cold event in the Taiwan Strait (northwestern Pacific) is one of these disasters. In studying the behind dynamical mechanism of the cold event, four topics were raised in sequence and consisted of my thesis. Two off-shore flows are identified and investigated through observed and model (ROMS) results. In order to better understand this cold event within global context, the global coastal responses and heat redistribution in the Indian Ocean are then studied using satellite data and a global climate model (CESM).

In winter, an off-shore flow of the coastal current can be inferred from satellite and in-situ data over the western Taiwan Bank. The dynamics related to this off-shore flow are examined here using observations as well as analytical and numerical models. The currents can be classified into three regimes. The downwind (southward) coastal current remains attached to the coast when the wind stress is stronger than a critical value depending on the upwind (northward) large-scale pressure gradient force, and an upwind current appears over the Taiwan Bank under a weak wind stress. The downwind coastal current and upwind current converge and the coastal current deflects onto the bank during a moderate wind. Analysis of the vorticity balance shows that the off-shore transport is a result of negative bottom stress curl that is

triggered by the positive vorticity of the off-shore flow. The negative bottom stress curl is reinforced by the gentle slope over the bank, which enhances the off-shore current. Composite analyses using satellite observations show cool waters with high chlorophyll in the off-shore current under the moderate wind, which support the model findings and may explain the high productivity over the western bank in winter.

The other off-shore flow in the northern strait often makes a U-turn to join the northward-flowing Taiwan warm current. In early 2008 of anomalously strong winter monsoon, cold water in the Taiwan Strait was moved sequentially by a cross-strait flow and a southward flow to the Penghu Island, causing a cold-related fish kill disaster. Except for the local wind forcing, the Coastal-Kelvin Waves (CKWs), intermittently propagating toward the TWS from north in winter, are an additional factor that could impact the flow patterns by changing cross-strait sea level gradient during the disaster. In the first stage (January 28-February 7), the reach of a large CKW trough induced an additional northward flow, which formed a cyclone after turning around the Zhangyun Ridge. Then, the cyclone led to an additional cross flow, which enhanced an eastward (offshore) movement of cold water. In the second stage (February 7-14), the arrival of a large CKW crest triggered an additional southward flow, which intensified a southward movement of the cold water. Due to the additional eastward and southward movements caused by the CKWs, the cold water could reach Penghu Island inducing a cold disaster.

The study on the global coastal SST from 1982 to 2013 revealed a significant cooling trend in the low and mid latitudes (31.4% of the global coastlines) after 1998, while 17.9% of the global coastlines changed from a cooling trend to a warming trend concurrently. The trend reversals in the Northern Pacific and Atlantic coincided with

the phase shift of Pacific Decadal Oscillation and North Atlantic Oscillation, respectively. These coastal SST changes are larger than the changes of the global mean and open ocean, resulting in a fast increase of extremely hot/cold days, and thus extremely hot/cold events. Meanwhile, a continuous increase of SST was detected for a considerable portion of coastlines (46.7%) with a strengthened warming along the coastlines in the high northern latitudes. This suggests the warming still continued and strengthened in some regions after 1998, but with a weaker pattern in the low and mid latitudes.

The study of heat transport in the Indian Ocean discloses a different pathway that the anomalous heat moves southward instead of westward caused by a strengthened southward transport and a weakened south equatorial current. This induces a striking heat build-up in the middle latitude of South Indian Ocean, contributes to the Southern Ocean warming, and intensifies heat hemisphere asymmetry. The heat increase has important climate impacts such as changes to rainfall over the western coast of Australia and increased coral bleaching. The new path discovered here may be an essential route linking the tropical Indo-Pacific Ocean and the Southern Ocean during the surface warming hiatus period.

Chapter 1

INTRODUCTION

There are four topics in my dissertation that two of them are related with two off-shore flows in the Taiwan Strait (TWS) and the other two of them are associated with global surface warming hiatus. In this chapter, I am going to first introduce the topography, the general flow pattern and related driving forcing in the TWS. The phenomenon and related explanations of the global surface warming hiatus are then given in this chapter.

1.1 The Topography in the Taiwan Strait

The TWS is an important channel that links the South China Sea and the East China Sea. As shown in 1.1, the coastline and seafloor topography in the TWS are complex. Off of Pingtan, a cross-strait ridge extends southward and then eastward to the eastern strait and separates the strait into two “sub-basins”. The eastern part of the ridge is called the Zhangyun Ridge, and the rest of the ridge is called the Pengbei Ridge [Wang and Chen, 1989]. The northern basin is the Guanyin Depression, and the southern basin is the Wuqiu Depression, which connects to the Penghu Channel in the south [Wu *et al.*, 2007]. In the southern strait, the Taiwan Bank (TWB) is a significant topographic feature with an area of approximately 160km×160km (characterized by the 40m isobath). The submarine bank consists of a large area of sandy shoals including submarine sand waves with crests as shallow as 20m [Shao *et al.*, 2011]. The shallowest portion of the bank is in the center with a deep submarine valley

(Penghu Channel) to the east and a shallow waterway to the west. The Penghu Channel slopes steeply to a depth of 2000-3000m into the South China Sea while the West Waterway widens to merge with the China coastline and has an averaged depth of 35 m. The TWS is located in the subtropical monsoon region. The prevailing southwesterly monsoon during the summer (between June and August) has an average wind speed of 5.1 m/s, whereas the prevailing northeasterly monsoon during the other seasons has an average wind speed of 10.6 m/s [Hu *et al.*, 2010].

1.2 The Winter Circulation Pattern in the Taiwan Strait

In winter, the TWS current is driven by two forces in opposing directions, i.e., southward wind stress (northeasterly monsoon) and northward pressure gradient force induced by Kuroshio. Two major currents are present under the influence of these two forces (1.1). The first is the China Coastal Current (CCC), a cold southward (downwind) current that is driven by the northeasterly monsoon and is characterized by low temperatures ($<20\text{ }^{\circ}\text{C}$) and low salinity ($<32\text{ psu}$); it flows southwestward along the western strait [Jan *et al.*, 2002]. The second current is the warm northward (upwind) current in the central and eastern strait [Chuang, 1985, 1986; Fang *et al.*, 1991]. The sea level in the TWS is generally higher in the south and lower in the north. Yang [2007] proposed that this pressure gradient, which is the reason the warm upwind current exists in the TWS throughout most of the entire year, is generated by the Kuroshio. The warm current, which has high temperatures ($>24\text{ }^{\circ}\text{C}$) and high salinity ($>34\text{ psu}$), has two origins: the extension of the South China Sea Warm Current and the extension of the Kuroshio into the eastern TWS. These two currents flow into the strait from the east side of the TWB, while the rest of the extension of the

South China Sea Warm Current flows into the strait from the west side of the TWB
[*Hu et al.*, 2010].

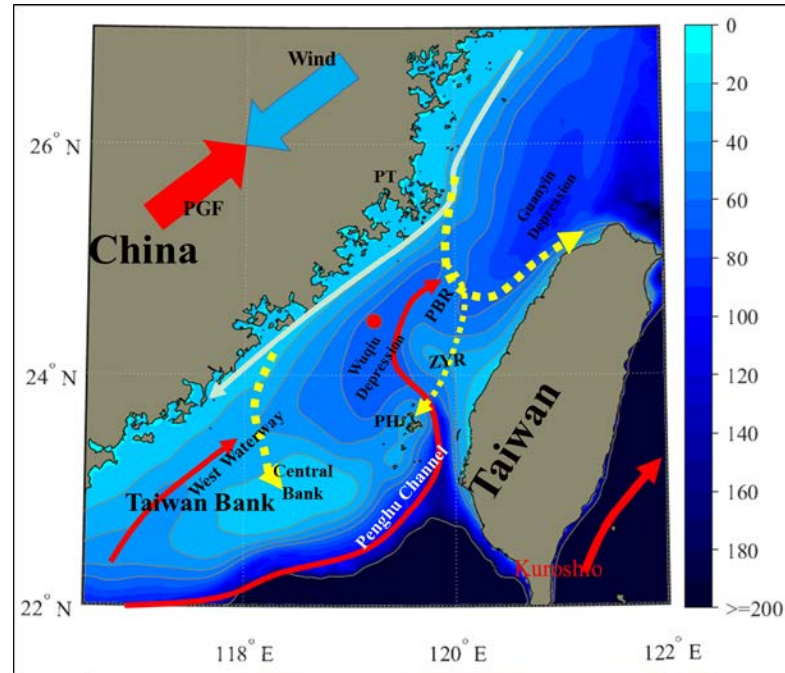


Figure 1.1 Seafloor topography and major currents in winter.

The grey line in the western strait denotes the China Coastal Current. The dashed yellow line cross the West Waterway is the off-shore branch in the southern strait, the dashed lines in the northern strait are the other off-shore branch that the northward arrow is the normal branch and the southward arrow is unusual branch in 2008. The red line in the West Waterway and Penghu Channel are the South China Sea Warm Current related to the Kuroshio. PH, PT, PBR, and ZYR are Penghu, Pingtan, Pengbei Ridge, and Zhangyun Ridge, respectively. PGF is the Pressure Gradient Force. The red dot represents the buoy's location.

Because of the varying northeasterly monsoon and complex seafloor topography, these two currents create a complicated flow pattern during the winter. In general, the CCC flows southwestward along the western coast and has two offshore

branches in the southern and northern strait respectively. The north-offshore branch reaches the Zhangyun Ridge and then returns northeastward, forming a U-shaped flow pattern that blocks the warm northeastward current [Jan *et al.*, 2002; Wu *et al.*, 2007]. The north-offshore branch that extends from the western strait to the Zhangyun Ridge was reported in 1989 [Huang, 1989; Wang and Chen, 1989] and has been reconfirmed by measurements [Hu *et al.*, 1999], model results [Jan *et al.*, 2002], and satellite data [Li *et al.*, 2006]. A zonal oceanic front over the Zhangyun Ridge is typically observed in winter [Chang *et al.*, 2006] and may be caused by the confluence of the CCC branch and the warm current from the south. The north offshore branch has been hypothesized to be caused by the bottom Ekman effect [Jan *et al.*, 2002; Lin *et al.*, 2005] and vorticity conservation [Wang and Chen, 1989]. The south off-shore flow is first time identified in this study and the related dynamics is investigated by theoretical and model analysis.

In contrast with the two aforementioned local factors (i.e., wind stress and sea level gradient), Ko *et al.* [2003] reported a huge influence of Coastal-Kelvin Waves (CKWs), generated in the Yellow and East China Sea, on the volume transport of the TWS. A positive (or negative) Sea Surface Height (SSH) anomaly, created by a wind stress event in the Yellow and East China Sea, can propagate southward along the China coast as the CKWs [Jacobs, 1998; Jacobs *et al.*, 1998b]. When CKWs pass through the East China Sea into the TWS, the CKWs crest induces a high sea level and strengthens the southward geostrophic current by changing the cross-strait sea level gradient [Ko *et al.*, 2003]. The CKWs are not only characterized by high sea level (wave crest), but also low sea level (wave trough). Is it possible that the low sea level can strengthen the northward geostrophic current like the high sea level strengthens

the southward geostrophic current? The northward warm current, as a counter-wind flow (the monsoon in winter is from northeast), has received strong interests [Ma *et al.*, 2010; Yang, 2007]. At present, there have been few factors reported that can drive the counter-wind flow except for the Kuroshio. Therefore, it is worth studying the impact of CKWs on the concerned counter-wind flow. If CKWs can impact both the northward and southward current, what is the role of the CKWs played in the cross-strait flow leading to the eastward (offshore) movement of cold water? These questions are going to be studied in this thesis.

Yin *et al.* [2014] indicated that the coastal-trapped waves (CTWs), propagating in the East China Sea, has three different modes: the free Kelvin wave mode, the forced first and second shelf wave modes. Their amplitudes are $O(10\text{ cm})$ and periods vary from 2 to 11 days. The sea level variation of the CKWs mainly results from the Kelvin wave mode from the coastal-trapped waves, whose phase speeds are about 15-18 m/s [Chen and Su, 1987; Li, 1989; Yin *et al.*, 2014]. Although there are several studies, involving the generations, mechanisms, and propagation properties, on the CKWs along the China coast, none of them elaborate the impact of CKWs on the 2008 cold disaster and north off-shore flow in the TWS.

1.3 Global Climate Change during the Hiatus Period (1999-2013)

Despite the continued increase of atmospheric greenhouse gases, the global surface mean temperature remained flat from 1998-2013 [Easterling and Wehner, 2009; Kosaka and Xie, 2013]. The global climate exhibited a shift from a rapid global surface warming to an unexpected deceleration with a notable cooling in the eastern tropical Pacific Ocean (a La-Niña-like pattern) and a strengthening of the trade winds [England *et al.*, 2014; Kosaka and Xie, 2013]. This change in the global surface

warming rate has received much notice with the title of “global surface warming hiatus” [Easterling and Wehner, 2009; England et al., 2014]. As indicated by Yan et al. [2016], this phenomenon is a surface characteristic and does not represent a slowdown in warming of the Earth climate system, but rather is an energy redistribution. The accurate description should be the “hiatus” of the global mean surface temperature increase, not a pause, but a slowdown of the warming rate. To be consistent with previous studies, we refer to this time period (1998-2013) as the “hiatus” period. One widely-accepted explanation is the change was induced by the heat sequestration from the atmosphere into the upper layer of equatorial Pacific via strengthened trade winds, and then the sequestered heat flowed westward into upper ocean in the Indian Ocean through strengthened Indonesian Throughflow (ITF) [Liu et al., 2016; Nieves et al., 2015] [Lee et al., 2015]. As a possible response to the surface warming hiatus, the recent global ocean heat anomaly pattern features an intensifying hemispheric asymmetry with 67-98% heat gain occurred in the southern Hemisphere extratropical ocean and a large build-up of heat in the mid-latitude of South Indian Ocean [Levitus et al., 2012; Roemmich et al., 2015; Wijffels et al., 2016].

1.4 Objectives and Motivation

Starting from a cold event, I first study the dynamics of two off-shore flow in the TWS. In February 2008, many coral reef fish froze to death on the beaches of Penghu Island [Hsieh et al., 2008]. Because of the massive loss to local aquaculture and the serious impacts on the local coral reef ecosystem, this event was called the cold disaster by many newspapers. The cold water that killed the fish at Penghu Island came from the offshore branch of the CCC in the northern strait, which was revealed by a remote sensing sea surface temperature (SST) map [Chang et al., 2009]. In

addition, another off-shore branch was also identified by satellite and in-situ data in the southern strait. The intrinsic dynamics of the off-shore flows have not been thoroughly investigated.

The cold disaster in the TWS was not an isolated event but may be related to global climate change. The extreme sea surface temperatures (SST) occurred recently in many coastal areas and caused serious physiological stress and large casualties to various coastal ecosystems[*Chen et al.*, 2014a; *Colella et al.*, 2012; *Eakin et al.*, 2010; *Feng et al.*, 2013; *Garrahou et al.*, 2009; *Hsieh et al.*, 2008; *Mills et al.*, 2013b]. It is imperative to study the corresponding coastal effects of the global surface warming hiatus. The understanding and prediction of the effects of the global surface warming hiatus on the coast are beneficial to the scientific field, policy makers, and general population.

These cold/hot events, as the responses to the global climate change from warming to slowdown, intrigue me to further study the behind mechanism within global context. As indicated by *Yan et al.* [2016], one of the major reasons may be related to the heat redistribution within oceans. The Indian Ocean, the warmest ocean during the hiatus period, is selected to study the heat transport. I hope this may answer part of question that why so many events occurred during the “hiatus” period. In addition, as the strengthened ITF moves the heat from the Pacific Ocean to the Indian Ocean, it is essential to continue tracking the anomalous heat to piece together a complete picture of energy redistribution within the global oceans during the period. An expected pathway is a continuous westward heat transport into the Atlantic Ocean via the global ocean conveyor belt [*Lee et al.*, 2015], i.e., South Equatorial Current and Agulhas Current. Surprisingly, here we disclose a different pathway that the

anomalous heat moves southward instead of westward caused by a strengthened southward transport and a weakened SEC. The path of anomalous heat transport discovered here may be an essential path linking the tropical Indo-Pacific and the Southern Ocean during the “hiatus” period.

Chapter 2

THE OFF-SHORE FLOW IN THE SOUTHERN STRAIT

2.1 Abstract

In winter, an off-shore flow of the coastal current can be inferred from satellite and in-situ data over the western Taiwan Bank. The dynamics related to this off-shore flow are examined here using observations as well as analytical and numerical models. The currents can be classified into three regimes. The downwind (southward) coastal current remains attached to the coast when the wind stress is stronger than a critical value depending on the upwind (northward) large-scale pressure gradient force, and an upwind current appears over the Taiwan Bank under a weak wind stress. The downwind coastal current and upwind current converge and the coastal current deflects onto the bank during a moderate wind. Analysis of the vorticity balance shows that the off-shore transport is a result of negative bottom stress curl that is triggered by the positive vorticity of the off-shore flow. The negative bottom stress curl is reinforced by the gentle slope over the bank, which enhances the off-shore current. Composite analyses using satellite observations show cool waters with high chlorophyll in the off-shore current under the moderate wind, which support the model findings and may explain the high productivity over the western bank in winter.

2.2 Introduction

A coastal current is observed along the coast of China in winter. Along the China Coastal Current (CCC), wave-like meanders and off-shore flows along the

length of the current exist as the current flows southward from East China Sea, through the Taiwan Strait and into the South China Sea. These variabilities can be due to an along-shore change in pycnocline undulation, triggered by changes in coastline and bathymetry, and flow convergence or divergence [Gan *et al.*, 2012; Oey *et al.*, 2014; Wang and Oey, 2016; Wu, 2015]. Two significant off-shore flows often occur in the northern around the Pingtan Island and southern Taiwan Strait near a prominent submarine bump, the Taiwan Bank (TWB) respectively (2.1). In some years of anomalously strong winter monsoon, the CCC carries unusually cold water ($<16^{\circ}\text{C}$) southward and the off-shore flow in the northern strait transports the coastal water off-shore to the Penghu Island [Hsieh *et al.*, 2008; Liao *et al.*, 2013a; Tang, 1978]. The unusually cold water can cause a devastating disruption of the local coral reef and the ecosystem. Lee *et al.* (2014) reported a 50-80% decrease in fish catch when the cold-water meander occurred in 2008. The off-shore flow also has a significant influence on the ecosystem at the TWB and is the focus of this study (Figures. 2.2 and 2.3).

The CCC water has a low salinity (28~34 psu), low temperature ($5\text{-}20^{\circ}\text{C}$), and high nutrient [Hong *et al.*, 2011; Huang, 1989; Oey *et al.*, 2013]. The current is along the China coastline with a speed of 0.1-0.3 m/s and its width extends to as far as the 60m isobath depending on the strength of the wind [Hu *et al.*, 1999; Pan *et al.*, 2013]. On the shelf of the northern South China Sea, the circulation is complicated by the presence of complex coastlines and variable bathymetry, as well as the existence of counter-wind (northward) South China Sea Warm Current [SCSWC, 2.1, [Hu *et al.*, 2010]]. In the East China Sea, offshore-penetrating tongues of CCC water are often observed [Wu, 2015]. In the northern Taiwan Strait, a branch of the CCC (2.1) often

makes a U-turn to join the northward-flowing Taiwan warm current [Liao *et al.*, 2013a; Oey *et al.*, 2014].

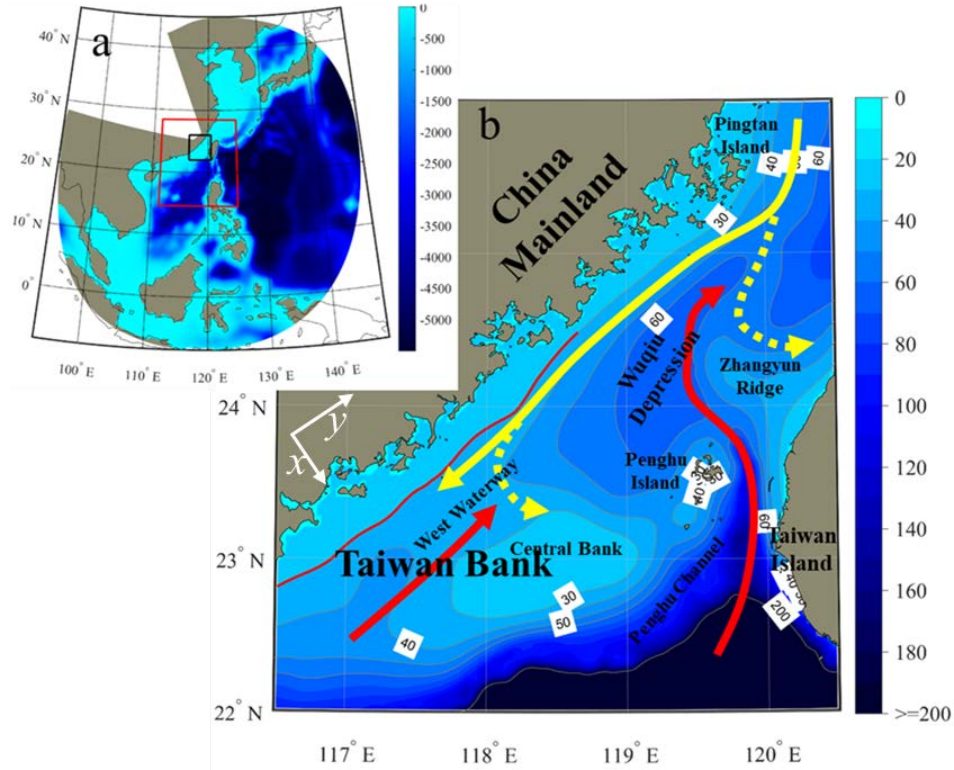


Figure 2.1 The model domain (a) and the topography around the Taiwan Bank (b). a, the color shading region is the domain of the course grid, the red box is the fine grid, the black box is the focus region which zooms in panel b. b, the Taiwan Bank as a focus region is divided into three parts: West Waterway, Central Bank (30 m isobaths in the center), and Penghu Channel. The grey lines are the isobaths. The red line is the 30m isobaths used for transect analysis. The yellow arrow represents the southward (downwind) coastal current and the dashed yellow arrows are the off-shore flows in the northern and southern strait. The red arrow in the West Waterway is the SCSWC's extension and the other red arrow in the Penghu Channel is the combination of SCSWC's extension and Kuoshio's extension. The schematic coordinate (along-shore and cross-shore) is plotted as the white line.

The TWB is a significant topographic feature with an area of approximately 160km×160km (characterized by the 40m isobath). The submarine bank consists of a large area of sandy shoals including submarine sand waves with crests as shallow as 20m [Shao *et al.*, 2011]. The shallowest portion of the bank is in the center with a deep submarine valley (Penghu Channel) to the east and a shallow waterway to the west. The Penghu Channel slopes steeply to a depth of 2000-3000m into the South China Sea while the West Waterway widens to merge with the China coastline and has an averaged depth of 35 m. In winter, circulation in the Taiwan Strait is influenced by two forces in opposite directions: northeasterly monsoon wind and open-ocean, northward pressure gradient force associated with the Kuoshio and warmer water of the South China Sea [Oey *et al.*, 2010; Yang, 2007]. Considering the different forcing influences, we divide the study region into three parts: West Waterway, Central Bank (30m isobath in the center), and Penghu Channel (2.1). A persistent northward flow dominated by the open ocean forcing is observed in the deep Penghu Channel [Jan and Chao, 2003; Lin *et al.*, 2005]. The West Waterway is alternatively occupied by the southward coastal current and the northward SCSWC depending on the balance between wind and open-ocean forcing [Guan and Fang, 2006; Zhu *et al.*, 2013]. Some studies stated that the CCC ends in the TWB, blocked by the TWB and the northward SCSWC [Wu, 1982; Xiao *et al.*, 2002], while others indicated that the coastal current can reach as far south as Hainan Island in the South China Sea in some years of strong winter monsoon [Guan and Fang, 2006].

Flow variations in the vicinity of submarine banks have been reported in the literature. Whitney and Allen [2009] indicated a downwelling jet is weakened over the Heceta Bank. Gong *et al.* [2015] reported an increased cross-shelf transport associated

with a coastal sea level setup during the strong winter storm period. *Gula et al.* [2015] suggested that local along-isobath pressure anomalies and topographic form stresses exerted by the Charleston Bump retard the Gulf Stream and steer the current seaward. *Gan et al.* [2012] pointed out that the bottom stress curl can induce an along-shore pressure gradient which then intensifies a geostrophic downslope cross-shore current over a widened shelf in the northern part of the South China Sea. The cross-shore transport in the northern Taiwan Strait is associated with the nonlinear term in the vorticity equation which is contributed by a stationary topographic Rossby wave over the lee of the Zhangyu Ridge [*Oey et al.*, 2014].

To understand the circulation and underlying dynamics of the off-shore transport in the TWB, a numerical model and theoretical analysis are used in this study. This paper is organized as follows. The spatial feature of the off-shore flow is identified first in observational data. After a description and validation of the numerical model, the criterion for the occurrence of off-shore flow is derived using model results. The mechanism is then investigated through vorticity equation and numerical experiments. The conclusions are summarized in the last section.

2.3 Observations

We examine the general feature of the cold water deflection using satellite data and ship data. The satellite data consists of SST, chlorophyll, and wind. The SST and chlorophyll are from NASA Moderate Resolution Imaging Spectroradiometer (MODIS) Terra. The MODIS/Terra acquires data in 36 spectral bands covering the global in two days. In this study, we use mapped 4-km and 8-day resolution SST from 2002 to 2015 (<https://oceandata.sci.gsfc.nasa.gov>). The wind is from Cross-Calibrated Multi-Platform (CCMP) Version 2 gridded surface vector winds products. The CCMP

data is produced using Version-7 RSS radiometer wind speeds, QuikSCAT and ASCAT scatterometer wind vectors, moored buoy wind data, and ERA-interim model wind fields. The Variational Analysis Method is applied to produce four maps daily of 0.25 degree gridded vector winds (<http://www.remss.com/measurements/ccmp>). We also use data from a vessel-survey by the China Third Institute of Oceanography in December 2, 2007-January 8, 2008 [Pan *et al.* 2013; Qiu *et al.* 2012].

The separated cold water brings nutrients into the TWB, leading to phytoplankton growth. Therefore, we choose both chlorophyll and SST as indicators of cold water deflection. The coastal current tends to deflect under an “intermediate” wind ($-1.14 \times 10^{-4} < \tau < -0.67 \times 10^{-4} \text{ m}^2/\text{s}^2$) that will be theoretically determined below. 2.2 shows the SST, chlorophyll, and wind stress composites of those times when the wind is intermediate.

An off-shore movement of cold water can be identified in 2.2a. The cold (10-16°C) water is along the coast which is considered as the CCC and the warm (>20°C) water dominates the open-ocean which is regarded as the SCSWC and Kuroshio extension. The cold water (16°C) leaves the coast at 23.5°N, crosses the West Waterway, and extends to the Central Bank. Correspondingly, the chlorophyll anomaly shows a high concentration in the West Waterway (2.2b) when the cold water spreads to the open ocean. Wang *et al.* [2016] showed the off-shore flow in the northern TWS can trigger a winter bloom during the wind relaxation, when the CCC water with high nutrient and low density transport off-shore, enhancing the stratification. Further study is needed for the detailed mechanism in the south TWS. The separated pattern is much more distinct in the chlorophyll than SST, which may be due to the strong winter wind and surface heat fluxes that can rapidly modify the

SST. This relationship between cold water and high chlorophyll suggests that the physical process of deflection is important to the regional marine ecosystem and fishery, further justifying a study of the phenomenon.

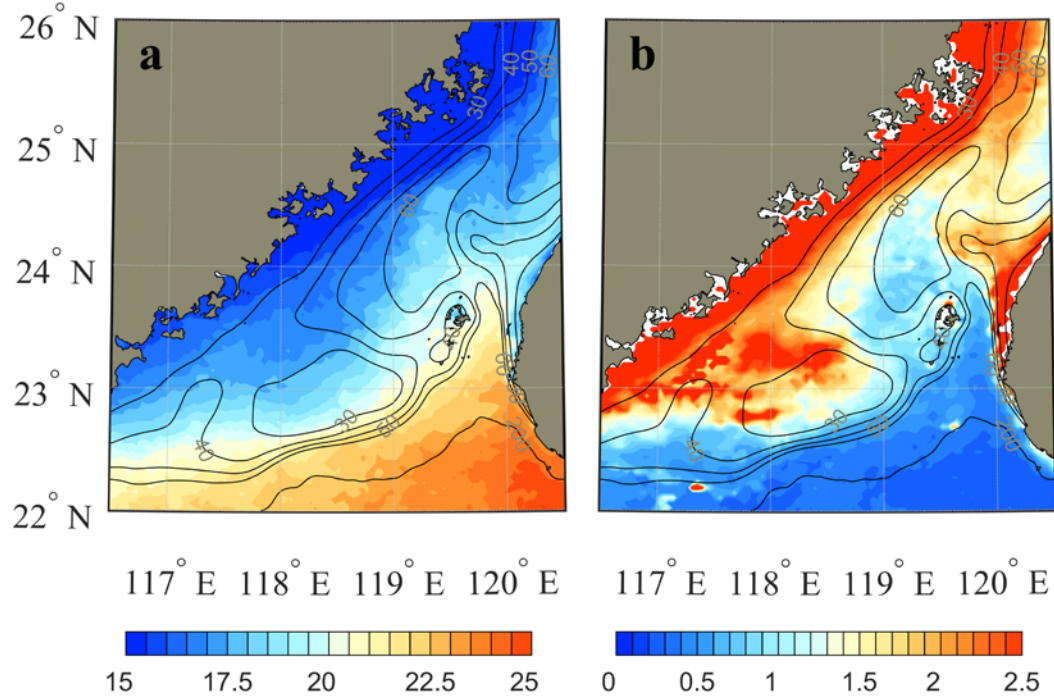


Figure 2.2 The SST (a, units: °C) and chlorophyll (b, units: mg/m³) composites of those times when the wind is intermediate (0.0922~0.144 N/m², or 8-10 m/s) in winter. The SST and chlorophyll are MODIS data. The wind criterial is computed in the Section 4. Winter time is defined as December, January, and February. The grey lines are the isobaths from 30m to 60m with a spacing of 10m.

A clear example of coastal current deflection over the TWB can be seen in observed salinity and temperature (20m depth) in January 2008 (2.3). Off-shore spreading of cold and fresh water is observed over the TWB. The cold and fresh water considered as the coastal current deflects from the 30m isobath near the coast at

23.5°N and crosses the West Waterway (2.3). Here too the feature is more distinct in salinity than SST, as the former is less affected by the strong surface heat flux. A clear front can be seen over the TWB with cold (17-19°C) and fresh (31-33 psu) water near the coast and warm (>20°C) and salty (>34 psu) water in the open ocean. Both satellite and *in situ* observations thus show a clear coastal current deflection pattern at 23.5°N.

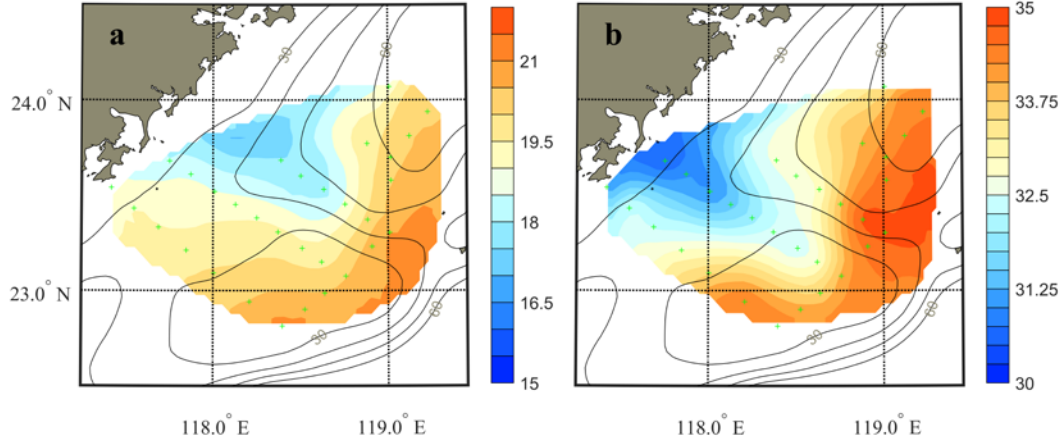


Figure 2.3. The observed temperature (a, units: °C) and salinity (b, units: psu) distributions (20m depth) around the Taiwan Bank. The observation is conducted by the China Third Institute of Oceanography in December 2, 2007-January 8, 2008. The green plus symbols are survey sites. The grey lines are the isobaths from 30m to 60m with a spacing of 10m.

2.4 Model Configuration

The numerical model used in this study is a one-way nested model based on the Regional Ocean Modelling System [ROMS, [Shchepetkin and McWilliams, 2003, 2005]]. The coarse grid is from the Taiwan Strait Nowcast/Forecast System (TFOR) which has served as an operational forecasting system for the Fujian Province Marine Forecasting Institute since 2003 [Lin *et al.*, 2016]. The TFOR has been extensively tested and validated, and the results have been used to study various oceanic

phenomenon in the Taiwan Strait [*Chen et al.*, 2014b; *Liao et al.*, 2013a; *Liao et al.*, 2013b; *Lu et al.*, 2015; *Wang et al.*, 2013]. The model covers the northwestern Pacific from 93.0°E to 148°E and 8.5°S to 45.0°N (2.1) with a curvilinear grid at a horizontal resolution of 1.5km to 45km and with 30 vertical terrain-following levels. The nested fine grid focuses on the Taiwan Strait from 111.4°E to 125.2°E and 14.5°N to 28.4°N (2.1) at a $1/32^\circ \times 1/32^\circ$ horizontal resolution and 25 vertical terrain-following levels. The model bathymetry consists of a survey data and ETOPO2V2 from the National Geophysical Data Center. In order to reduce the sigma-coordinate pressure gradient truncation errors, a depth filter by *Mellor et al.* [1998] is applied to smooth the bathymetry. The surface forcing data, consisting of wind stress, net heat flux, and freshwater flux are from MERCATOR PSY3V2R2 [<http://www.mercator-ocean.fr> (2008)]. The surface net heat flux is obtained by the prescribed climatological surface net heat flux and a correction term proportional to the difference between the climatological SST and the model SST [*Barnier et al.*, 1995]. The initial and lateral boundary conditions of the fine-grid model are interpolated from the coarse-grid model. In addition, the lateral boundary condition includes major rivers' discharge along the China coast and sea-level forcing by 10 main tidal components (i.e., M2, S2, N2, K2, K1, O1, P1, Q1, Mf, and Mm) from TPXO7.0 [*Egbert and Erofeeva*, 2002].

The default ROMS schemes for horizontal and vertical advection terms of momentum and tracer equation are used: the 3rd-order upstream bias and 4th-order centered for horizontal and vertical advections of 3D momentum respectively, 4th-order centered for 2D momentum and tracers' equation. The Smagorinsky diffusion is used for the harmonic horizontal mixing of tracers, which occurs along the geopotential surface. The vertical viscosity and diffusivity are calculated by the Mellor

and Yamada (1982), MY2.5 turbulence model. The radiation boundary condition employs the Flather [Flather, 1987] scheme for the 2-D velocity. The Chapman boundary condition is chosen for the model to include background elevation and tidal processes [Chapman, 1985] and the Clamped boundary condition is employed for the 3-D velocity and tracers (Temperature and salinity).

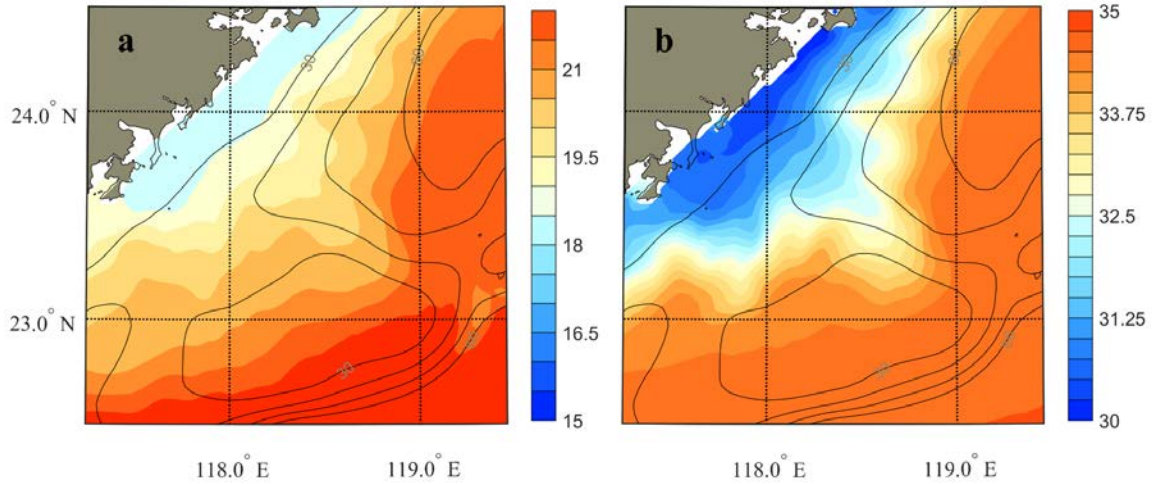


Figure 2.4. The simulated temperature (a, units: °C) and salinity (b, units: psu) distributions (20m depth) around the Taiwan Bank. The model results are on December 22, 2007. The grey lines are the isobaths from 30m to 60m with a spacing of 10m.

The model results from October 2007 to February 2008 were selected to compare with the in-situ data in 2008 and theoretical analysis. The simulated temperature and salinity distribution are plotted in 2.4. The simulations demonstrate an off-shore extension of the coastal current in the West Waterway which is similar with in-situ data during the winter of 2008 (2.3). The cold and fresh coastal current appears to be blocked by the submarine bump and warm and salty water from the south, and could not flow pass the TWB as a branch of it extends from the 30m isobath to the

Central Bank at 23.5°N. A north-southward temperature gradient over the TWB can be seen. The model simulates well the off-shore flow and will be used to analyze the underlying dynamics. The model was extensively validated against time series of velocity and temperature (January and February in 2008) in *Liao et al.* [2013a], in which we studied an unusually cold event in February 2008 in the northern Taiwan Strait. The velocity validation in *Liao et al.* [2013a] shows the correlation coefficient of the cross-shore and along-shore velocities between observations and model outputs are 0.61 and 0.74 respectively.

2.5 Classification of Circulation Regimes

Before analyzing the mechanism of the off-shore flow, we present in this section a simple analysis to examine the wind criterion for the off-shore flow. When the off-shore flow emerges over the TWB, the concurrent flow pattern along the coastline is a flow convergence with an along-shore northward flow at the TWB and an along-shore southward flow to the north of the TWB. Here, we use the along-shore flow along the coastline to identify the off-shore flow. The along-shore flow is easily associated with the wind stress if we select two near-coast regions (i.e., R_{twb} and R_{ntwb} in 2.6b) where the cross-shore velocity is negligible (Coriolis term is not important (shown in supplementary FigureA 1 in Appendix A)). The wind criterion for the off-shore flow is then obtained through the link between along-shore flow and wind stress.

Considering barotropic dynamics for a weak stratification due to the strong winter monsoon and tides [*Jan et al.*, 2002; *Lefèvre et al.*, 2000; *Li et al.*, 2006; *Oey et al.*, 2014], a flow under surface and bottom stresses with small spatial and temporal Rossby number has shallow-water momentum equation,

$$-fv = -g \frac{\partial \eta}{\partial x} + \frac{\tau^{sx}}{H} - \frac{\tau^{bx}}{H} \quad (2.1)$$

$$fu = -g \frac{\partial \eta}{\partial y} + \frac{\tau^{sy}}{H} - \frac{\tau^{by}}{H} \quad (2.2)$$

Here, x is cross-shore, positive in the offshore direction, y is along-shore (along China coastline), positive poleward. H is the undisturbed water depth, η is sea level, g is the acceleration due to gravity, $\mathbf{u} = (u, v)$ is the depth-averaged velocity. $\boldsymbol{\tau}^s = (\tau^{sx}, \tau^{sy})$ and $\boldsymbol{\tau}^b = (\tau^{bx}, \tau^{by})$ are kinematic surface and bottom stress vectors respectively with a units of m^2/s^2 . In the near-coast regions (i.e., R_{twb} and R_{ntwb} in 2.6b) where $v \gg u \approx 0$, Figure A 1 (Appendix A) shows surface wind stress can be balanced by the pressure gradient and bottom stress. Therefore Eq. (2.2) becomes

$$rv = -g \frac{\partial \eta}{\partial y} + \frac{\tau^{sy}}{H} \quad (2.3)$$

Here τ^b can be parameterized as: $\tau^b = rHu$, where r is a constant friction coefficient [$\sim 1.66 \times 10^{-5} \text{ s}^{-1}$ [Oey *et al.*, 2014]]. This equation indicates that the alongshore flow is driven by a balance between along-shore wind stress and sea level gradient. While the latter should strictly be treated as part of the solution, it is useful in coastal seas [Csanady, 1982] to write η as a sum of a slowly-varying large-scale sea level η_0 which is externally imposed, and a fluctuating part η' which evolves as part of the flow: i.e $\eta = \eta_0 + \eta'$. Thus the pressure-gradient acceleration likewise consists of a large-scale, externally imposed part $g\partial\eta_0/\partial y$, and a fluctuating part $g\partial\eta'/\partial y$ that influences and depends on the flow $\mathbf{u}$. In the Taiwan Strait, Oey *et al.* [2014] show that the fluctuating pressure gradient $g\partial\eta'/\partial y$ is small compared to $g\partial\eta_0/\partial y$ due to e.g. the Kuroshio and/or warmer water of the South China Sea. Moreover, the wind varies at time scales from a few days to 1~2 weeks, while $\eta \approx \eta_0$ varies at time scales of

months and longer [Oey *et al.*, 2014]. The different time scales indicate that the along-shore sea level gradient can be considered as being steady compared to the wind. The large-scale forcing generally imposes a poleward pressure-gradient acceleration: $-g\partial\eta_0/\partial y > 0$, which in winter is opposite to the acceleration due to the wind stress: $\tau^{sy}/H < 0$ in the Taiwan Strait. When the northeasterly wind is stronger (weaker) than the poleward pressure gradient, the along-coast velocity is negative or equatorward (positive or poleward). On the other hand, one expects the emergence of a cross flow, $u \neq 0$ but $v \approx 0$, when the wind is moderate and the wind stress and pressure gradient terms nearly balance each other in Eq. (2.3) [c.f. Oey. *et al.* 2014].

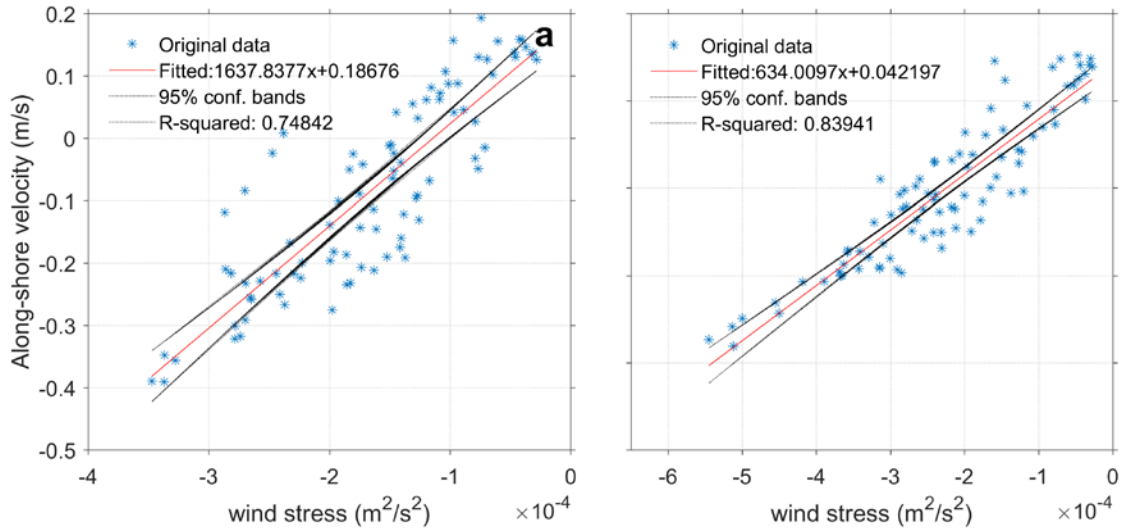


Figure 2.5. The regression of regional-mean along-shore velocities (units: m/s) and wind stress (units: m²/s²) in the region Rtwb (a, TWB) and Rntwb (b, north TWB).

The red line in the middle is the linear fitted line and the black lines outside are the confidence bands with a level of 95%. The equation of the regressed line can be seen in the legend. Region Rtwb and Rntwb are the green and red rectangles in 2.6b respectively. These two regions represent the TWB and north of the TWB.

We use the model results to regress the along-shore velocity (v) against the along-shore wind stress τ^{sy} according to Eq. (2.4) [Oey *et al.*, 2014].

$$v = -\frac{g}{r} \frac{\partial \eta}{\partial y} + \frac{\tau^{sy}}{rH} \quad (2.4)$$

Regressions (2.5) are calculated for v and τ^{sy} averaged in the near-coast regions Rtwb and Rntwb at the TWB and to the north of the Bank. The goodness of the fit (high r^2 value) and non-zero intercept indicate the existence and validity of $\eta \approx \eta_0$ as assumed *a priori*; the intercepts yield $(\partial \eta / \partial y)_{\text{Rtwb}} \approx -4.65 \times 10^{-7}$ and $(\partial \eta / \partial y)_{\text{Rntwb}} \approx -2.72 \times 10^{-7}$. These values are comparable to those obtained in other studies [Li *et al.*, 2017; Oey *et al.*, 2014; Wu and Hsin, 2005]. It is interesting that the along-shore pressure gradient in the TWB is almost 2 times larger than that further north in the Taiwan Strait [Li *et al.* 2017]. Following Oey *et al.* [2014], we can calculate a critical wind stress τ_c^{sy} such that the along-shore flow is equatorward (poleward) for northeasterly wind stress stronger (weaker) than τ_c^{sy} , i.e. since τ^{sy} is negative: $\tau^{sy} < \tau_c^{sy}$ ($\tau^{sy} > \tau_c^{sy}$). For Rtwb, the $\tau_c^{sy} = \tau_{c1}^{sy} \approx -0.67 \times 10^{-4} \text{ m}^2/\text{s}^2$ ($\sim 7 \text{ m/s}$), while $\tau_c^{sy} = \tau_{c2}^{sy} \approx -1.14 \times 10^{-4} \text{ m}^2/\text{s}^2$ ($\sim 9 \text{ m/s}$) for Rntwb. Physically, the currents at both sites Rntwb and Rtwb are equatorward for $\tau^{sy} < -1.14 \times 10^{-4} \text{ m}^2/\text{s}^2$ that is sufficiently strong to overcome the poleward pressure gradient force. The currents are poleward for weak wind stress $\tau^{sy} > -0.67 \times 10^{-4} \text{ m}^2/\text{s}^2$ that is overcome by the pressure gradient force, The currents are convergent for intermediate wind stresses ($-1.14 \times 10^{-4} < \tau^{sy} < -0.67 \times 10^{-4} \text{ m}^2/\text{s}^2$) when the windstress is sufficiently strong to overcome the pressure gradient force at Rntwb, but is being overpowered by the stronger pressure gradient at Rtwb. In the latter, intermediate case, one expects the CCC to deflect from the coast onto the TWB.

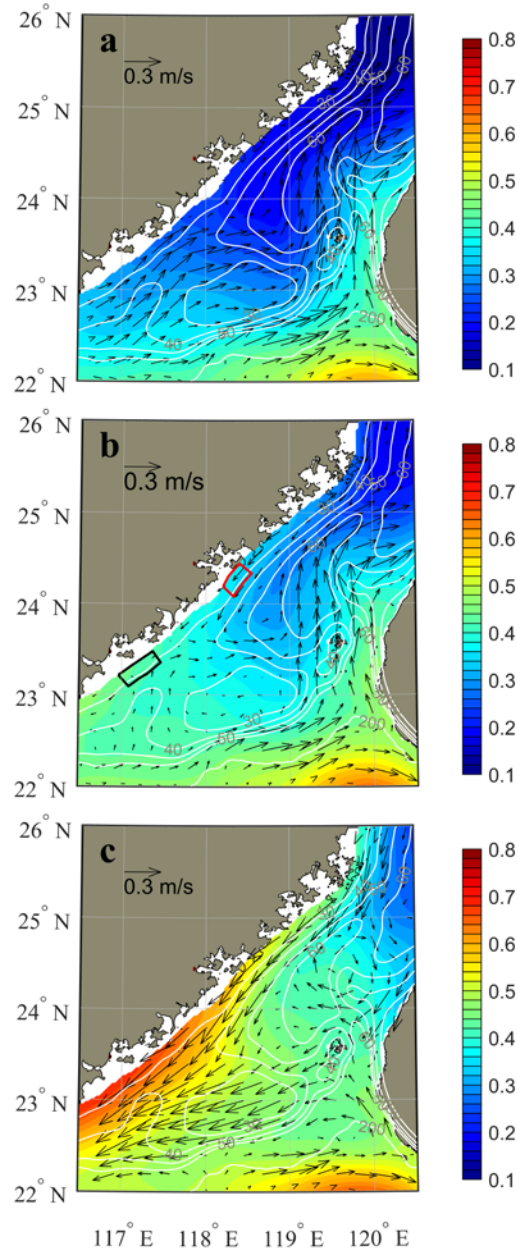


Figure 2.6. The mean composite patterns of sea level (color, units: m) and depth-averaged current (vector) for weak (a, $\tau^{sy} > -0.92 \times 10^{-4} \text{ m}^2/\text{s}^2$), moderate (b, $-1.36 \times 10^{-4} < \tau^{sy} < -0.92 \times 10^{-4} \text{ m}^2/\text{s}^2$), and strong (c, $\tau^{sy} < -1.36 \times 10^{-4} \text{ m}^2/\text{s}^2$) wind stress.

The mean pattern is the mean model results when the along-shore wind stress in different ranges. The green and red rectangles are the Region Rtwb and Rntwb respectively. These two regions represent the TWB and north of the TWB. The grey lines are the isobaths.

To test these ideas, we composite the model currents according to the above critical windstresses (2.6). Despite the simplicity of the theoretical model, currents near the coast are equatorward when the wind stress is strong $\tau^{sy} < -1.14 \times 10^{-4}$ (2.6c), but are poleward when the wind stress is weak $\tau^{sy} > -0.67 \times 10^{-4} \text{ m}^2/\text{s}^2$ (2.6a). For intermediate wind stress, two oppositely directed near-coast flows converge (2.6b), and a cross flow in the West Waterway appears. This is the phenomenon we often observe in winter (Figures. 2.2 and 2.3). The intermediate wind stress -1.14×10^{-4} to $-0.67 \times 10^{-4} \text{ m}^2/\text{s}^2$ is close to the along-shore winter-mean wind stress $-1.49 \times 10^{-4} \pm 0.84 \times 10^{-4} \text{ m}^2/\text{s}^2$ (mean $\pm$ 1std), which explains why the coastal deflection phenomenon occurs frequently near the TWB. The occurrence criteria computed above can be verified through the relationship between chlorophyll and wind stress. The chlorophyll over the TWB tends to be maximum during the intermediate wind stress. Chlorophyll observations from 2002-2015 show that high chlorophyll concentration occurs for windstress in the range, but low for windstress outside of the range (Figure A 2 in the Appendix A).

2.6 Vorticity Dynamics of the Off-shore Flow

The simple model above is useful, but it is linear and does not include realistic forcing and topography. To gain further insights into the relevant dynamics, we analyze the results from ROMS simulation, focusing on December 2008 when a deflection of the China Coastal Current onto the TWB occurred [Figures. 2.3 and 2.4, Qiu *et al.*, 2012]. In the experiment, the surface and lateral boundary conditions of December 2008 were applied for three months. The model results in the last month are considered as being quasi steady and averaged to analyze the vorticity dynamics.

2.6.1 Along-shore Variation

According to observation (2.3), the cold and fresh water leaves the coast at about 23.5°N and spreads from the 30m isobath to the TWB. The 30m isobath (red line in the 2.1) in the West Waterway is therefore selected to analyze the along-shore variability of the deflection process. The along-shore pressure gradient term, flow convergence, and depth-averaged velocity along this 30m isobath are shown in 2.7. The along-shore pressure gradient term (with minus sign) is positive with a peak at 23.5°N (2.7a) indicating a large-scale southward elevating sea level with a huge-local increase at 23.5°N . The large-scale sea level slope is related to the open ocean forcing (i.e., Kuroshio) while the steep local slope is associated with a local effect (e.g., topographic effect). Correspondingly, a positive (off-shore) cross-shore velocity occurs near the same location (2.7c). The cross-shore velocity at other locations is weak and negative (i.e., on-shore). The along-shore velocity is weak and positive (northward) in the TWB ($23\text{-}23.5^{\circ}\text{N}$) but is strong and negative (southward) to the north and south of the bank (2.7c). This leads to a flow convergence around the 23.5°N where off-shore flow occurs.

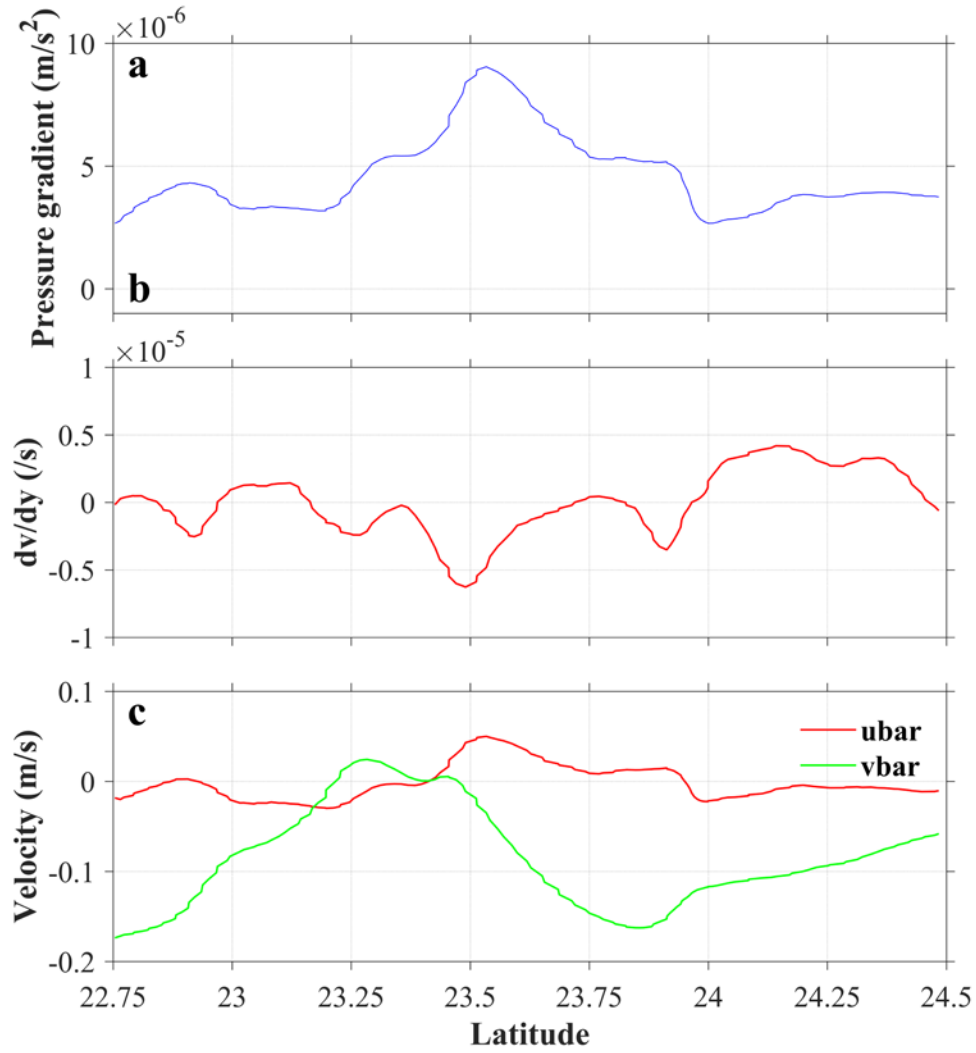


Figure 2.7. The along-shore pressure gradient term ($-\partial p / \partial y / \rho$, a), flow convergence indicated by $\partial v / \partial y$, (b), and depth-averaged along-shore ($\bar{v}$) and cross-shore ($\bar{u}$) velocities (c) along the 30m isobath.

The positive values of $\bar{u}$ and $\bar{v}$ are eastward (off-shore) and northward respectively. The 30m isobath is referred as the red line in the 2.1. The TWB location is between 23 and 23.75°N and the deflection is around 23.5°N.

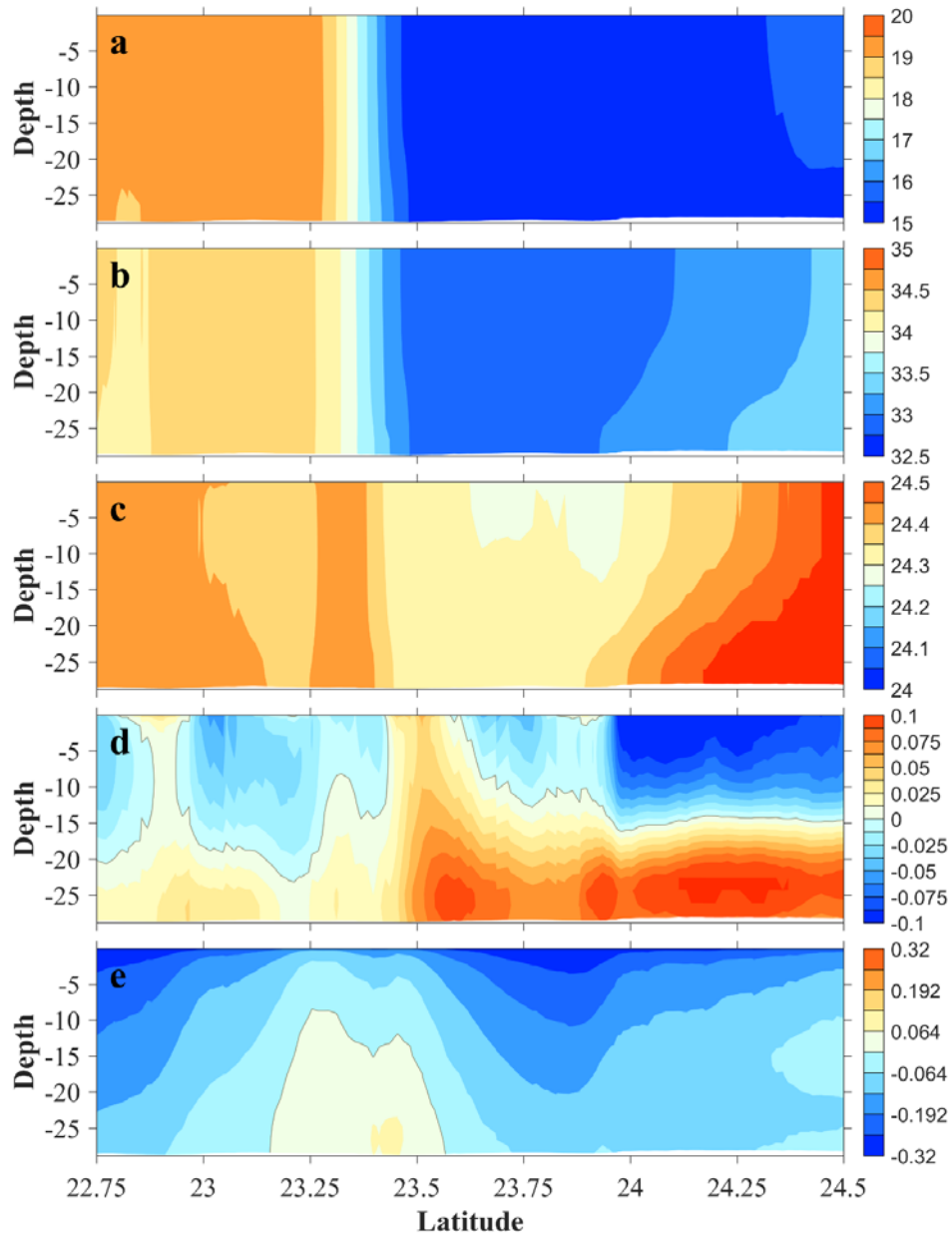


Figure 2.8. The transect profiles of temperature (a, units: °C), salinity (b, unites: psu), density (c, unites: +1000kg/m³), cross-shore (d) and along-shore (e) velocities (units: m/s) along the 30m isobath. The positive values of velocity are eastward (off-shore) and northward respectively. The 30m isobath is referred as the red line in the 2.1. The TWB location is between 23 and 23.75°N and the deflection is around 23.5°N.

As a result of the great flow convergence near the TWB (shown by dv/dy in 2.7b), a strong temperature and salinity front is found over the TWB, separating cold and fresh water to the north from warm and salty water in the south (Figures. 8a and 8b). The two different water masses indicate that the CCC does not flow pass the TWB when off-shore transport occurs. The temperature and salinity profiles also show a well-mixing water column, suggesting that the assumption of barotropic dynamics (weak stratification) is appropriate. However, there is a weak density gradient on the bank (2.8c). The cross-shore velocity (2.8d) is one layer, offshore, on the bank, but shows a two-layer flow south and north to the bank: on-shore in the surface layer and off-shore in the bottom layer, as in downwelling circulation driven by the northeasterly wind. The off-shore velocity extends over the entire water column with decreasing magnitude from bottom to surface over the TWB (near 23.5°N). The along-shore velocity (2.8e) is strongly sheared and is negative everywhere, except near the TWB where it is negative in the upper layer but positive in the lower layer (around 23.5°N). In summary, the transect profiles thus show significant along-shore variations that include a steep drop of along-shore sea level near the Bank with a strong off-shore velocity, a sheared and even reversing along-shore velocity, and a strong temperature and salinity gradient over the TWB. The southward coastal current is weakened around the bank, and a reversed (northward) current (SCSWC) is produced over the bank. The two opposite currents converge and produces an off-shore transport with large temperature and salinity gradients. The dynamics determining these along-shore variations are presented next.

2.6.2 Vorticity Balance

Following Oey *et al.* [2014], we derived a vorticity Eq. (2.7) from continuity [Eq. (2.5)] and depth-averaged momentum [Eq. (2.6)] equations to understand the effect of topography and wind and open-ocean forcing on the deflection process. The vorticity balance is analyzed in this section and the detailed vorticity and flow fields is shown in the following section.

$$\frac{\partial \eta}{\partial t} + \nabla \cdot \mathbf{u}D = 0 \quad (2.5)$$

$$\frac{\partial \mathbf{u}}{\partial t} + (\xi + f)\mathbf{k} \times \mathbf{u} + \nabla \frac{(|\mathbf{u}|^2)}{2} = -g\nabla \eta - \frac{g}{\rho_0} \frac{1}{D} \int_{-H}^{\eta} \int_z^0 \nabla \rho dz' dz + \frac{\boldsymbol{\tau}^s}{D} - \frac{\boldsymbol{\tau}^b}{D} \quad (2.6)$$

where η is sea level, $\mathbf{u}D$ is the transport vector per unit length, where $\mathbf{u}(u,v)$ is the depth-averaged velocity and $D=(H+\eta)$ is the total water depth. H is the undisturbed water depth. ξ is the vertical component of vorticity ($\nabla \times \mathbf{u}$), $\mathbf{k}$ is the unit vector in the vertical direction. f is the Coriolis coefficient. ∇ is the horizontal gradient operator. g is gravity coefficient. ρ is the density, ρ_0 is the reference density, z and z' are the depths in the vertical direction. $\boldsymbol{\tau}^s$ and $\boldsymbol{\tau}^b$ are kinematic surface and bottom stress vector (unit: m^2/s^2) respectively. After taking the curl of the depth-averaged momentum equation [Eq. (2.6)] and approximating $D \approx H$ and the flow is steady ($\partial \eta / \partial t \approx 0$), we get the depth-averaged vorticity equation:

$$\underbrace{\frac{\partial \xi}{\partial t}}_{TR} + \underbrace{\mathbf{u}H \cdot \nabla \frac{f}{H}}_{BPV} \approx \underbrace{-\mathbf{u}H \cdot \nabla \frac{\xi}{H}}_{RPV} + \underbrace{J\left[\chi, \frac{1}{H}\right]}_{JEBAR} + \underbrace{\nabla \times \frac{\boldsymbol{\tau}^s}{H}}_{WSC} + \underbrace{\left[-\nabla \times \frac{\boldsymbol{\tau}^b}{H}\right]}_{BSC} \quad (2.7)$$

$$\text{where } \chi = \frac{g}{\rho_0} \int_{-H}^{\eta} z \rho dz'$$

and $J(A,B)=A_x B_y - A_y B_x$. In order to differentiate the second and third terms in the Eq. 2.7, we define f/H as background potential vorticity, ξ/H as relative potential vorticity. In the Eq. (2.7), the TR term in the left side is the

time rate term, the BPV term is the advection of background potential vorticity (PV), the terms in the right hand side are advection of relative PV (RPV, nonlinear term), joint effect of baroclinicity and relief [JEBAR; [Huthnance, 1984; Mertz and Wright, 1992]], wind stress curl (WSC), and bottom stress curl (BSC) respectively. The time rate term is negligible for a steady flow ($\partial \xi / \partial t = 0$, 2.9a). When the right hand side equals zero and TR is negligible, a steady flow is along f/H contours. The BPV is considered as an index that positive (negative) for upslope (downslope) transport. The upslope (downslope) transport is on-shore (off-shore) in coastal areas where water depth deepens from coastline to open ocean. These four terms in the right hand side represent different processes that can modify a cross flow. The RPV ($-\mathbf{u}H \cdot \nabla \frac{\xi}{H}$, with minus sign) is related to the flow field and gradient of relative PV. A positive (negative) RPV is regarded as a down-gradient (up-gradient) advection of relative PV by the flow field. For the JEBAR ($J(\chi, \frac{1}{H})$), the χ is a vertical integration of density with a weight of depth and physically represents the baroclinicity. The JEBAR is considered as the combination of baroclinicity and sloping bottom topography and arises for an along isobath density gradient. The WSC is externally imposed input of vorticity by a curl in the wind stress, while the BSC is the dissipation of vorticity by bottom friction.

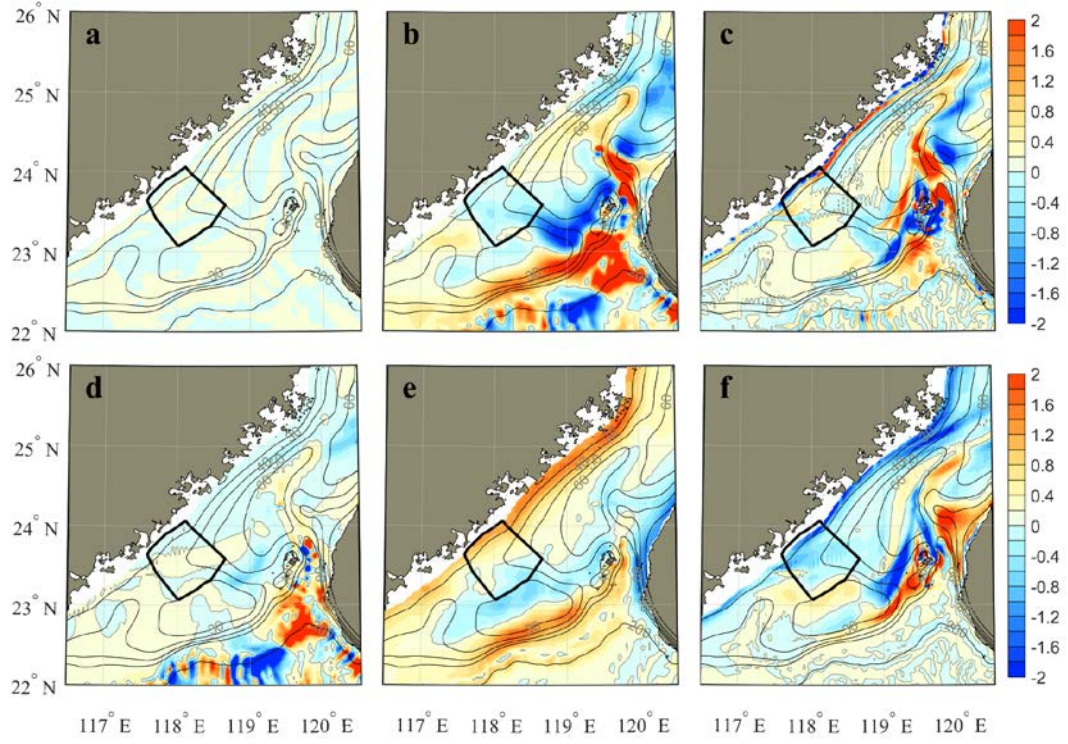


Figure 2.9. The composite patterns of different terms in the vorticity equation [Eq. (4.3), units: $\times 10^{-10} \text{s}^{-2}$]. a is the TR, b is BPV, c is the RPV, d is JEBAR, e is the wind stress curl (WSC), f is the bottom stress curl (BSC) in the Eq. (2.7). Since f is considered as constant in this small region, we plot isobaths (H , grey lines) to represent f/H . Black box is the focus region where off-shore flow occurs.

These terms are plotted in 2.9 to determine which process(es) control the cross flow on the TWB. Since f is regarded as constant in this small region, contours of H almost coincide with the f/H in 2.9. The flow along f/H contours indicates approximate geostrophic flow when BPV is very weak. Comparing with the five terms in Eq. (2.7), time rate term (TR, $\partial \xi / \partial t$) in the 2.9a is negligible. The BPV has large negative values in the northern part of the West Waterway (about 23.5°N , 118°E , black box, 2.9b),

suggesting a strong downslope transport of the coastal current in this region where the off-shore flow is observed (Figures. 2 and 3). The mean BPV in the black box (2.9) is $-0.25 \times 10^{-10} \text{ s}^{-2}$ while the mean BSC is $-0.42 \times 10^{-10} \text{ s}^{-2}$ which indicates the negative BPV here is mainly contributed by the bottom stress curl of these four processes. The RPV (nonlinear term, 2.9c) has reversed effect (upslope transport) and the JEBAR term (2.9d) is weak (mean is $0.018 \times 10^{-10} \text{ s}^{-2}$) and positive (upslope) in this region. The reversed effect of RPV is induced by the down-gradient advection of relative PV which is related to the flow field and gradient of relative PV which could be referred in the 2.11. The weak JEBAR term is related to the weak along-shore density gradient shown in 2.8c. The wind stress curl (2.9e, mean is $0.18 \times 10^{-10} \text{ s}^{-2}$) can balance most of the negative values induced by the bottom stress curl (2.9f).

Along the China coast, another significant negative BPV region is around the Pingtan Island in the northern Taiwan Strait (2.9b). The downslope flow in this region is related to the nonlinear term in Eq. (2.7) [Oey *et al.*, 2014]. The BPV is positive (on-shore flow) at other coastal areas to the north and south of the TWB, being mainly produced by the wind stress curl term in the so called “topographic- β ” Sverdrup balance [Oey *et al.*, 2014], while the curl of bottom stress and nonlinear term nearly cancel in a thin near-coast boundary layer (Figures 2.9c and 2.9f). A strong negative BPV distributes between Wuqiu Depression and the TWB where is predominantly controlled by wind stress curl and bottom stress curl. Since water depth shallows from west to east in this region, wind stress curl induces a negative value (westward and downslope flow) in this region. In the eastern TWB, an upslope flow (positive BPV) is strengthened by the nonlinear term, wind stress curl, and bottom stress curl. The outer shelf and shelf break ($>100\text{m}$) is dominant by the JEBAR term (2.9d). The reason for

such a great JEBAR effect here may be tied to a strong density gradient produced by the cold water in the strait and warm water from Kuroshio intrusion.

2.6.3 The Along-isobath Balance

To further examine the cross flow, we examine the vorticity equation [Eq. (2.7)] along the 30m isobath. Since the 30m isobath is almost parallel to the shoreline, the coordinate system is the same as before. The BPV ($\mathbf{u}H\nabla \frac{f}{H}$) is written as $uH \frac{\partial}{\partial x} \frac{f}{H} + vH \frac{\partial}{\partial y} \frac{f}{H}$ and the second part ($vH \frac{\partial}{\partial y} \frac{f}{H}$) is ignored along the isobath (y direction) on a f -plane. The first part ($uH \frac{\partial}{\partial x} \frac{f}{H}$) remains in the left side and the cross-isobath velocity (u) is then linked to the vorticity dynamics after dividing by $H \frac{\partial}{\partial x} \frac{f}{H}$.

$$u \approx \underbrace{\frac{H}{fH_x} \left(\overline{uH} \cdot \nabla \frac{\xi}{H} \right)}_{u_{RPV}} - \underbrace{\frac{H}{fH_x} J \left(\chi, \frac{1}{H} \right)}_{u_{JEBAR}} - \underbrace{\frac{H}{fH_x} \left(\nabla \times \frac{\overline{\tau^s}}{H} \right)}_{u_{WSC}} + \underbrace{\frac{H}{fH_x} \left(\nabla \times \frac{\overline{\tau^b}}{H} \right)}_{u_{BSC}} \quad (2.8)$$

where the H_x is the partial differentiation of depth normal to the isobaths (bathymetry slope). The cross-isobath velocity in Eq. (2.8) is produced by four processes which are RPV, JEBAR, WSC, and BSC. These terms along the 30m isobath are shown in 2.10a. The intensified off-shore flow in the bank (positive u_{all} around 23.5°N) is evidently induced by the u_{BSC} as noted previously in 2.9. The u_{BSC} increases gradually from 24°N upstream (north) of the bank and reaches a maximum at 23.5°N and the u_{RPV} has a reversed pattern, while the u_{WSC} and u_{JEBAR} keep flat along the 30m isobaths. The increase of u_{BSC} overcomes the other three terms leading to a positive velocity in the bank. Both u_{BSC} and u_{RPV} weaken rapidly to the south of 23.5°N where the u_{WSC} is again dominant. Except for the TWB, the cross-isobath velocity in other areas is primarily driven by the wind stress curl.

To examine the bottom stress curl in detail, we split this term into three components [Eq. (2.9)] according to $\tau_b = C_d |u_b| u_b$, where u_b is the bottom velocity and $|u_b|$ is the magnitude of the bottom velocity.

$$u_{BSC} = \frac{H}{fH_x} \nabla \times \frac{\vec{\tau}^b}{H} = \underbrace{\frac{C_d |\vec{u}_b|}{fH_x} \nabla \times \vec{u}_b}_I + \underbrace{\frac{H}{fH_x} C_d |\vec{u}_b| \nabla H^{-1} \times \vec{u}_b}_{II} + \underbrace{\frac{C_d}{fH_x} \nabla |\vec{u}_b| \times \vec{u}_b}_{III} \quad (2.9)$$

where u_{BSC} is the cross-isobath velocity induced by the bottom stress curl, $C_d \approx 2.5 \times 10^{-3}$ is the bottom drag coefficient. The Eq. (2.9) indicates that the u_{BSC} is attributed to three processes. The term I is the cross flow caused by frictional dissipation of the vorticity. If the flow has positive relative vorticity, the effect of friction is to cause a downslope (off-shore) velocity as parcel moves to region with lower background PV, and vice versa for negative vorticity which then induces an upslope (on-shore) velocity. The cross-isobath velocity induced by this term (I) is in geostrophic balance with an along-isobath pressure gradient (see supplemental material). The term II is described as a “drag torque” by Oey *et al.* [2014], and can be further expressed as $C_d |u_b| v_b / H / f$, which is the bottom Ekman transport. The velocity caused by this term is confined in the bottom Ekman depth. The term III is considered as the “speed torque” by Oey *et al.* [2014] and represents a similar physical process of term I that this term is also geostrophic and extends through the water column. However, this term only involves shear vorticity and the term I includes shear and curvature vorticity [Liu and Gan, 2014; Oey *et al.*, 2014]. The shear vorticity depends on a spatial variation of flow field and the curvature vorticity is only associated with the curving streamline. In addition to the relative vorticity, magnitudes of terms I and III are amplified by a gentler slope (H_x) in the Eq. (2.9) over the TWB. If Eq. (2.9)

times f , *Gan et al.* [2012] stated the terms I and III, related to local topography, induce a locally along-isobath pressure gradient which is derived in the Appendix A. The peaks of terms I and III in the bank are consistent with the sharp increase of sea level in 2.7a. This confirms the cross-isobath velocity induced by term I and III are in geostrophic balance with the along-shore sea-level gradient.

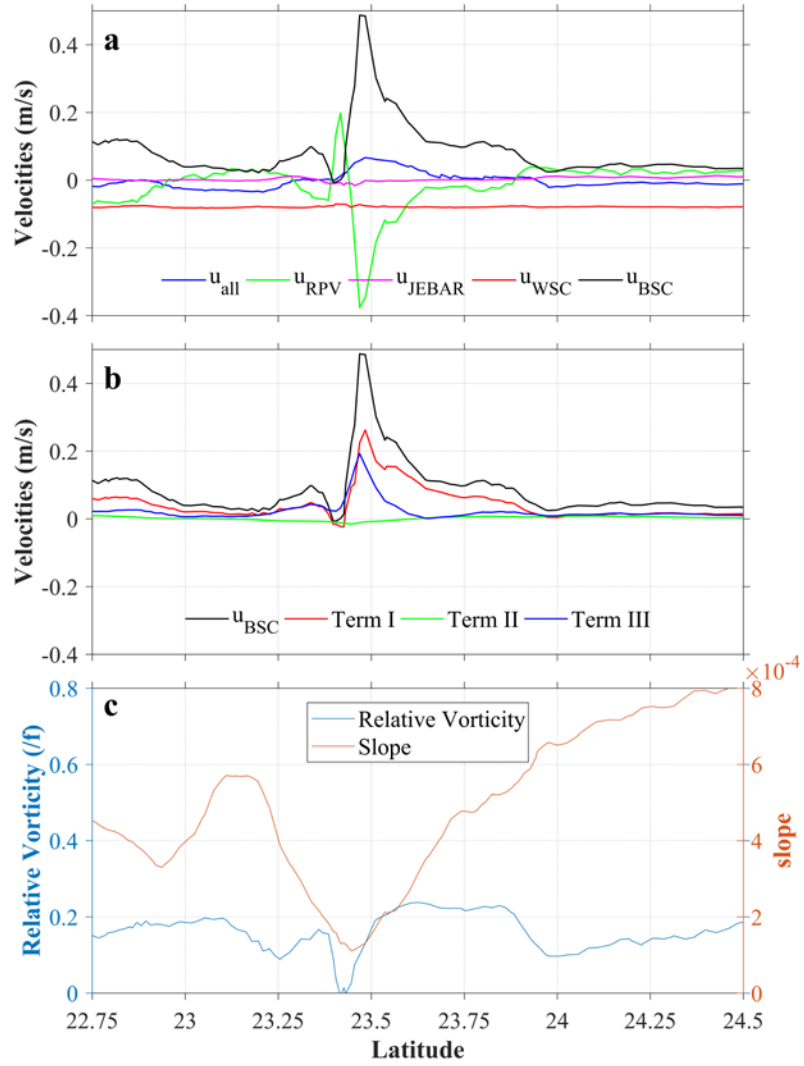


Figure 2.10. a. The overall cross-isobath velocity (u_{all}) and the cross-isobath velocities caused by different terms according to Eq. (2.8).

b. The cross-isobaths velocities induced by bottom stress curl and its three components according to Eq. (2.9). c. The relative vorticity and slope along the 30m isobath. The relative vorticity is normalized by the Coriolis coefficient (ζ/f). All three figures are plotted along the 30m isobath referred as the red line in the 2.1. The TWB location is between 23 and 23.75°N and the deflection is around 23.5°N.

The relative strength of each term in Eq. (2.9) is plotted in 2.10b. The u_{BSC} is mainly contributed by the terms I and III. The bottom Ekman transport term (II) contributes little. The term I slowly increases at 24°N while term III grows rapidly in the northern TWB (23.5-24°N) and decays fast in the southern side. The dominance of terms I and III clearly demonstrates a tight relation between this off-shore flow and the relative vorticity and slope. 2.10c reveals a decline of slope and an increase of relative vorticity at 23.5°N. This suggests both slope and vorticity contribute to the reinforcement of the bottom stress curl here. Note that the contribution of the slope decline seems more than the relative vorticity based on their ratio changes. The slope is about 8×10^{-4} to the north of the bank, and decreases as low as 1×10^{-4} at 23.5°N in the bank. To the south of 23.5°N, the slopes increases to 5.8×10^{-4} around 23.1°N and then falls to 4×10^{-4} . The relative vorticity decreases from $0.2f$ to $0.1f$ to the north of the bank, and increases as large as $0.21f$ in the bank. The vorticity is zero at 23.4°N where velocity is zero shown in the 2.7c. To the south of the bank, the vorticity varies between $0.1f$ and $0.2f$. Through Eqs. (2.7 and 2.9), the mechanism can be explained that a flow with a positive relative vorticity results in a negative bottom stress curl for friction removal. This negative bottom stress curl is strengthened at a gentle slope in the TWB which induces an enhanced off-shore current over the bank. In another point

of view [Gan *et al.*, 2012], the off-shore flow is regulated by the dynamics related to the locally along-shore pressure gradient generated by the local topography.

2.6.4 Numerical Experiments

As shown above, the strong positive vorticity and weak slope result in the off-shore flow in the TWB. The vorticity depends on flow field which is modified by topography, the northeasterly monsoon wind, and open ocean forcing. We wish to understand how they are related. Numerical experiments are conducted to investigate their relationship (Table 1). The slope along the 30m isobath ranges from 4×10^{-4} to 6×10^{-4} in the non-TWB case while varies between 1×10^{-4} and 8×10^{-4} in the control case.

Table 1 The experiments descriptions

Case names	Descriptions
Control case	Dynamic analysis
Wind only	Northward pressure gradient (NPG) at the lateral boundary is removed in the fine-grid model.
NPG only	Wind is removed in the fine-grid model.
non-TWB case	The Taiwan Bank (gentle slope) is removed in the fine-grid model.

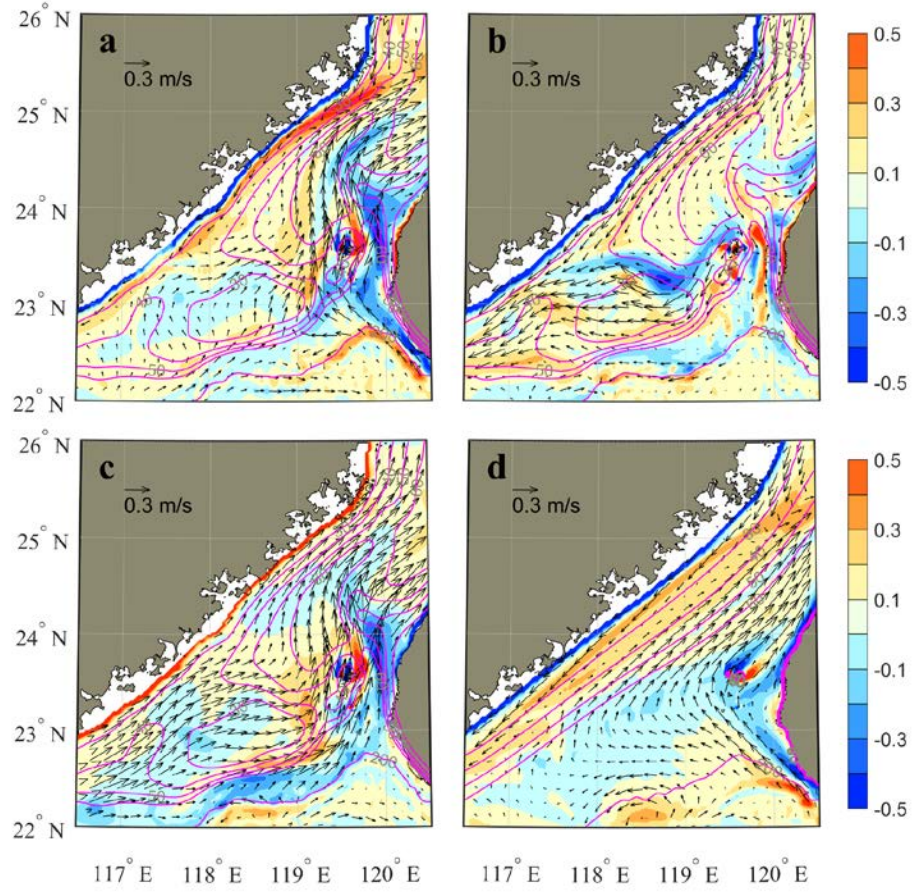


Figure 2.11. The composite patterns of depth-averaged relative vorticity (shading) and currents (arrows) in the experiments. a is the control case, b is the wind only, c is the NPG only, d is the non-TWB case. The grey lines are the isobaths from 30m to 60m. The relative vorticity is normalized by the Coriolis coefficient (ζ/f).

The depth-averaged current and relative vorticity of all experiments are plotted in the 2.11. An apparent cross flow only occurs in the control case (2.11a) which suggests the topography and both forces all play an important role. In the control case, a stripe of positive vorticity is along the coast and enlarges in the West Waterway. The stripe also emerges in the non-TWB case (2.11d), but disappears in the cases with only

one force (i.e., Wind only and NPG only, Figures 2.11b and 2.11c). This indicates the stripe of positive vorticity is related to the concurrence of two opposite currents, i.e. the coastal current attached along the coastline and the SCSWC in the shelf. However, this stripe of positive vorticity alone leads to a weak cross flow without a gentle slope in the non-TWB case (2.11d). The apparent cross flow in the control case suggests this also needs the contribution of a gentle slope around 23.5°N. In the Wind only and NPG only cases, a gentle slope exists but a large positive vorticity vanishes, there is evidently no cross flow in the West Waterway (23.5°N). These experiments suggest that the cross flow is associated with the TWB, wind and open-ocean sea-level gradient. The latter two opposite forces induce a southward coastal current along the coastline and a northward SCSWC in the shelf, which converge over the TWB forming an increased positive vorticity. This increased positive vorticity plus a gentle slope cause an amplified a cross flow in the bank.

2.7 Conclusions

This study provides evidence of and explains the cold-water deflection in winter over the TWB using observations and models. The cold and fresh coastal current leaves the 30m isobath at 23.5°N, crosses the West Waterway, and spreads from the coast to the Central Bank. The chlorophyll shows high concentration in the same region, in agreements with the temperature and salinity distributions. The separated, nutrient-rich coastal water is important to the marine ecosystem and fishery. Model simulation and analyses are conducted to understand the deflection process and associated mechanism.

Three circulation regimes are derived depending on the balance of wind stress and pressure gradient associated with the poleward tilt of sea level along the Taiwan

Strait. The flow is controlled by a southward (northward) current when the wind stress is stronger (weaker) than an upper (lower) bound. The two opposite flows occur simultaneously and converge, and the coastal current deflects when the wind stress is moderate.

There is a striking along-shore variation around 23.5°N along the 30m isobath. The southward coastal current is weakened in the TWB and a reversed (northward) current is generated over this bank. Associated with this two-flow concurrence, an off-shore transport is strengthened and large temperature and salinity gradients are formed over the bank.

Vorticity analysis and numerical experiments indicate that the off-shore transport is associated with the topography of the TWB and two opposite external forces: wind and sea-level gradient. Over the TWB, the northeasterly monsoon drives a southward coastal current along the coastline and the open ocean forcing induces a northward warm current (SCSWC) on the shelf. These two currents meet at the western bank with the former limited in the inner shelf and latter located outer, generating a strong relative vorticity. The high relative vorticity trigger a negative bottom stress curl which is reinforced in the bank by the gentle slope over the Bank. This reinforced negative bottom stress curl dominates the other terms in Eq. (2.7) and leads to an amplified off-shore flow over the bank.

Chapter 3

THE ROLE OF COASTAL-KELVIN WAVES ON THE 2008 COLD DISASTER IN THE TAIWAN STRAIT

The influence of local forcing on the 2008 cold event in the TWS was studied in *Liao et al.* [2013]. In this chapter, I am going to discuss the influence of a remote factor, coastal Kelvin waves, on the 2008 cold event.

3.1 Abstract

In early 2008, cold water in the Taiwan Strait (TWS) was moved sequentially by a cross-strait flow and a southward flow to the Penghu Island, causing a cold-related fish kill disaster. Except for the local wind forcing, the CKWs, intermittently propagating toward the TWS from north in winter, are an additional factor that could impact the flow patterns by changing cross-strait sea level gradient during the disaster. In the first stage (January 28-February 7), the reach of a large CKW trough induced an additional northward flow, which formed a cyclone after turning around the Zhangyun Ridge. Then, the cyclone led to an additional cross flow, which enhanced an eastward (offshore) movement of cold water. In the second stage (February 7-14), the arrival of a large CKW crest triggered an additional southward flow, which intensified a southward movement of the cold water. Due to the additional eastward and southward movements caused by the CKWs, the cold water could reach Penghu Island inducing a cold disaster.

3.2 Introduction

In the La Niña winter (January and February) of 2008, extremely strong northeasterly winds (> 6.7 m/s) lasted an exceptional long duration (> 30 days) in the Taiwan Strait (TWS, [Chang *et al.*, 2009]). The strong winds strengthened the China Coastal Current, carrying a large amount of extremely cold water from north into the northwestern TWS [Zhu *et al.*, 2013]. An unusual offshore branch of the cold China Coastal Current then moved the cold water to the Penghu Island (PHI), southeastern strait [Liao *et al.*, 2013a]. The PHI, well known for its coral reef ecosystem, is generally surrounded by warming water (> 20 °C) in winter [Jan *et al.*, 2006]. The unexpected cold water (< 16 °C) intrusion induced the temperature-sensitive coral reefs and fish to die around the PHI, resulting in serious mortality to the entire coral reef ecosystem [Hsieh *et al.*, 2008]. Similar cold disaster at the PHI occurred four times before 2008 (three times between 1930 and 1934 and once in 1976; [Tang, 1978]).

Normally, the cold offshore branch of the China Coastal Current moves the cold water offshore and turn northward forming a U-turn in the northern TWS [Jan *et al.*, 2002; Li *et al.*, 2006; Oey *et al.*, 2014]. However, Liao *et al.* [2013a] stated the cold offshore branch, instead of a normally northward turn, unusually flowed southward to the PHI, southern strait in 2008. In detail, during January 28 – February 7, a persistent cross-strait flow controlled the TWS and moved the cold water offshore to the central strait. Another strengthened along-strait flow next dominated the TWS and carried the separated cold water southward to the Penghu Island In February 7 – February 14 (PHI, [Liao *et al.*, 2013a]). The period between January 28 and February 7 is defined as the first stage and the subsequent period in February 7 – February 14 is called the second stage hereafter. As a result, the unexpected cold water intrusion is

related to two strengthened flow patterns that follow each other in a particular order. In a normal year, the combination of off-shore and southward flows may happen, but the flows are weak and impersistent that cannot move the cold water through the Zhangyun Ridge to the PHI, southern strait [Jan *et al.*, 2002; Li *et al.*, 2006]. Liao *et al.* [2013a] attributed the strengthened off-shore and southward flows to a particular strong local wind stress and a likely weakened northward warm current. However, it is still wondering that whether there are any other factors involved in the two unusual strengthened processes besides local forcing.

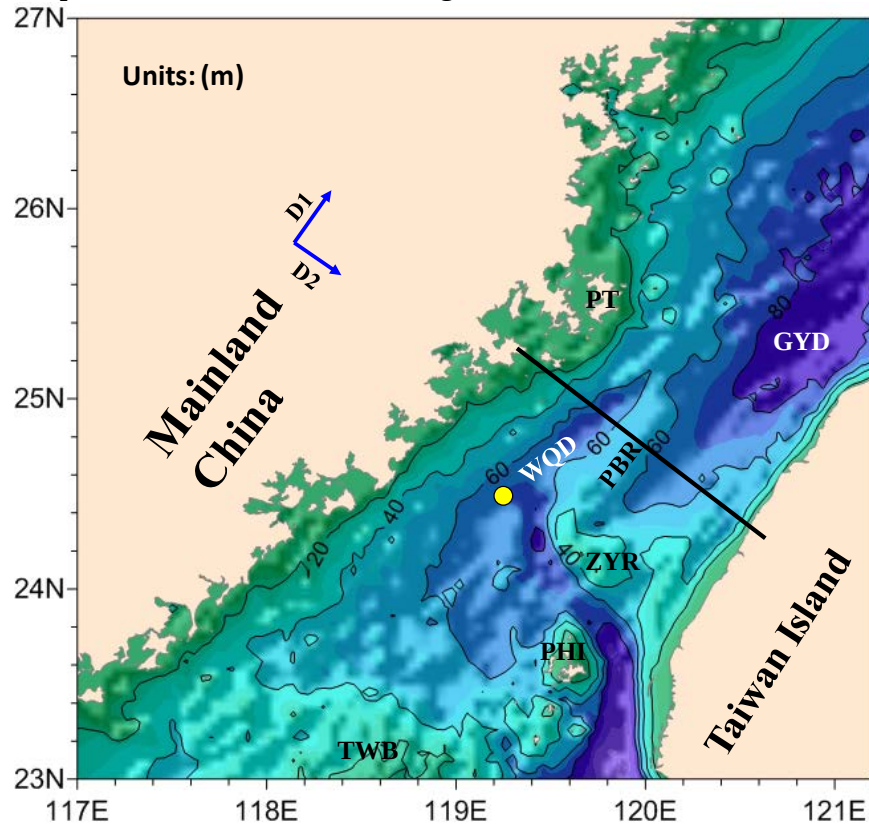


Figure 3.1 The seafloor in the Taiwan Strait.

D1 and D2 are the along-strait and cross-strait directions, respectively. PT, PHI, TWB, ZYR, PBR, WQD, and GYD denote the Pingtan, Penghu Island, Taiwan Bank, Zhangyun Ridge, Pengbei Ridge, Wuqiu Depression, and Guanyin Depression, respectively. The yellow dot

represents the buoy's location. The black straight line is the cross-strait section used for volume transport calculation.

In winter, the TWS current is driven by two forces in opposing directions, i.e., southward wind stress (northeasterly monsoon) and northward pressure gradient force induced by Kuroshio [Yang, 2007]. Both two forces interact with each other forming three complex current patterns: the southward (northward) current for northeasterly winds stronger (weaker) than an upper (lower) bound and the cross-strait flow for relaxing northeasterly winds between two bounds [Oey *et al.*, 2014]. In contrast with the two aforementioned local factors, Ko *et al.* [2003] reported a huge influence of CKWs, generated in the Yellow and East China Sea, on the volume transport of the TWS. A positive (or negative) Sea Surface Height (SSH) anomaly, created by a wind stress event in the Yellow and East China Sea, can propagate southward along the China coast as the CKWs [Jacobs *et al.*, 1998a; Jacobs *et al.*, 1998b]. When CKWs pass through the East China Sea into the TWS, the CKWs crest induces a high sea level and strengthens the southward geostrophic current by changing the cross-strait sea level gradient [Ko *et al.*, 2003].

The CKWs are not only characterized by high sea level (wave crest), but also low sea level (wave trough). Is it possible that the low sea level can strengthen the northward geostrophic current like the high sea level strengthens the southward geostrophic current? The northward warm current, as a counter-wind flow (the monsoon in winter is from northeast), has received strong interests [Ma *et al.*, 2010; Yang, 2007]. At present, there have been few factors reported that can drive the counter-wind flow except for the Kuroshio. Therefore, it is worth studying the impact of CKWs on the concerned counter-wind flow. If CKWs can impact both the

northward and southward current, what is the role of the CKWs played in the cross-strait flow leading to the eastward (offshore) movement of cold water? These questions are going to be studied in this paper.

Yin et al. [2014] indicated that the CTWs, propagating in the East China Sea, has three different modes: the free Kelvin wave mode, the forced first and second shelf wave modes. Their amplitudes are $O(10\text{ cm})$ and periods vary from 2 to 11 days. The sea level variation of the CTWs mainly results from the Kelvin wave mode, whose phase speeds are about 15-18 m/s [*Chen and Su*, 1987; *Li*, 1989; *Wang et al.*, 1988; *Yin et al.*, 2014]. Although there are several studies, involving the generations, mechanisms, and propagation properties, on the CKWs along the China coast, none of them elaborate the impact of CKWs on the 2008 cold disaster in the TWS.

The objective of this study is to investigate the impacts of CKWs on the circulation during the 2008 cold disaster in the TWS. The paper is organized as follows. The data, method, and model are described in Section 2. The impacts of CKWs on the northward (southward) flow and transport in the TWS are studied in Sections 3. The relation between the CKWs and the cold water intrusion is explored in Sections 4. The discussion and summary are given in Section 5.

3.3 Model, Data and Method

The model used in the study is a one-way nested model which is based on the Regional Ocean Modelling System (ROMS), a free-surface, hydrostatic, primitive equation ocean circulation model with nonlinear terrain-following coordinates [*Shchepetkin and McWilliams*, 2003, 2005]. The coarse grid is from the TWS Nowcast/Forecast System (TFOR), which has been an operational forecasting system for the Fujian Province Marine Forecasting Institute since 2003. The TFOR has been

extensively assessed and verified by *Lin et al.* [2016] and the results has been used to study various oceanic phenomenon in the TWS [*Chen et al.*, 2014b; *Liao et al.*, 2013b; *Liao et al.*, 2015; *Lu et al.*, 2015; *Wang et al.*, 2013]. The model spans the northwestern Pacific from 99.0°E to 148°E and 9.0°S to 44.0°N at 1/8° horizontal resolution with 25 vertical levels. The fine-grid model covers the TWS from 111.4°E to 125.2°E and 14.5°N to 28.4°N at 1/32° horizontal resolution with 25 vertical levels. The model bathymetry is a combination of survey data and ETOPO2v2 from National Geophysical Data Center (<http://www.ngdc.noaa.gov/mgg/fliers/01mgg04.html>). The air-sea flux data (heat and mass) and wind forcing for both nested grids are from MERCATOR PSY3V2R2 (<http://www.mercator-ocean.fr>, 2008). The initial and lateral boundary conditions of the fine-grid model are interpolated from the results of the coarse-grid model. In addition, the major rivers' discharge along the coast and sea-level forced by 10 main tidal components are included in the lateral boundary condition.

The default schemes in the ROMS are adopted to resolve horizontal and vertical advection terms of momentum and tracer equation in the fine-grid model (3rd-order upstream bias and 4th-order centered for horizontal and vertical advections of 3D momentum respectively, 4th-order centered for 2D momentum and tracers' equation). The Smagorinsky diffusion is used for the harmonic horizontal mixing of tracers, which occurs along the geo-potential surface. The vertical viscosity and diffusion rates are calculated by the MY2.5 turbulence module. The Flather scheme is employed for the radiation boundary condition [*Flather*, 1987]. The tidal processes are added to the model by the Chapman boundary condition [*Chapman*, 1985]. The other

model configurations and validations are detailed in *Liao et al.* [2013a] and *Lin et al.* [2016].

The CKWs originated in the Yellow and East China Sea and transported into the TWS through sea level and velocity variability. The coarse-grid domain covers the whole area and the fine-grid domain only focuses on the TWS. The CKWs is generated in the coarse-grid model and transports into the fine-grid model through boundary conditions, i.e., sea level and velocity. Therefore, the CKWs' transport into the TWS can be controlled by filtering the sea level and velocity variability in the north-lateral boundary of the fine-grid model. In order to investigate the role of CKWs in the circulation during the cold disaster, a sensitive numerical test (non-CKWs case, here after) was performed in the fine grid from 30 December 2007 to 20 March 2008. The only difference between control and non-CKW case is the north-lateral boundary condition. For the control case, the north-lateral boundary conditions were interpolated from coarse-grid model without any change. For the non-CKW case, the north-lateral boundary conditions, i.e., sea level and velocity, were interpolated from the coarse-grid model and then were filtered with a low-pass filter. The CKWs periods vary from 2 to 11 days in the East China Sea [*Yin et al.*, 2014], where the north boundary of the fine grid is located. Therefore, in the north boundary of the non-CKWs case, the momentums (sea level and velocities in 2D and 3D), temperature, and salinity were filtered by a 15-day low-pass filter to remove the CKWs signals entering the model from the north. The other three lateral boundary conditions and surface forcing, e.g., wind stress, surface heat flux, and surface mass flux are the same in the control and non-CKW cases. The control case includes two factors: local wind stress and remote CKW, while the non-CKW case only include local wind stress. Note that the wind

stress in 2008 is much stronger than the wind stress in a normal year. The non-CKW case is not a normal case, but a special case with an extremely-strong local wind.

The hourly velocities, measured by an acoustic Doppler current profile (ADCP) buoy (location is shown in 3.1), were used to compute the wave frequency during the cold disaster. The buoy was included in the Fujian Coastal Monitored System Project and is used in the monitoring system at the Fujian Marine Forecasts Institute (<http://www.fjmf.gov.cn/OceanObservation/BigBuoy.aspx>, 2014). The measured depth is from 5.5 m to 50.5 m with a vertical resolution of 5 m. The 36-h low-pass-filtered depth-mean velocities are used and rotated to the cross-strait and along-strait directions.

The Ensemble Empirical Mode Decomposition (EEMD) method is applied to analyze the frequency (Hilbert spectrum) of buoy velocities. This is an adaptive method which decomposes a time series into a set of intrinsic mode functions and a residual monotonic function through a sifting process. The EEMD can overcome the mode mixing problem in the Empirical Mode Decomposition [*Huang et al.*, 1998]. The EEMD scheme can extract physically meaningful modes from a non-stationary and nonlinear time series, and thus has been widely used in oceanography [*Ezer et al.*, 2013; *Li et al.*, 2012; *Yin et al.*, 2014]. The Hilbert spectrum can be calculated by applying Hilbert transform after obtaining the intrinsic mode functions.

3.4 The Impact of CKWs on the Northward and Southward Flow

The major characteristics of SSH anomaly (tide-filtered SSH minus monthly mean SSH) were notably low and high sea levels in the TWS during the first and second stage, respectively (3.2). Eight panels of SSH anomaly at different times demonstrate an excitation and transportation of the low and high sea levels from the

Yellow Sea to the TWS along the China coast. For the low SSH anomaly event, sea level fell with a weak wind stress in the Yellow Sea on February 3, and then propagated southward through the East China Sea to the TWS in one day. For the high SSH anomaly event, sea level rose with a strong wind stress in the Yellow Sea on February 11, and then transported southward. The high sea level also took almost one day to transport from the Yellow Sea to the TWS. Note that the coast is located to the right of the propagation direction.

The time series of the SSH anomaly along the 30 m isobath (the isobath is shown in 3.2b) is plotted in 3.3a. The figure illustrates propagations of low and high sea levels from north to south, which coincided well with 3.2. According to the propagation distance and period (3.3a), the propagation speed is estimated to be 16.8 m/s, which almost equals the phase speed of long surface gravity wave in 30 m depth. The speed approximates to the phase speed of the first mode of CTWs (Kelvin wave) reported in CTWs studies along the China coast as well [*Chen and Su, 1987; Yin et al., 2014*].

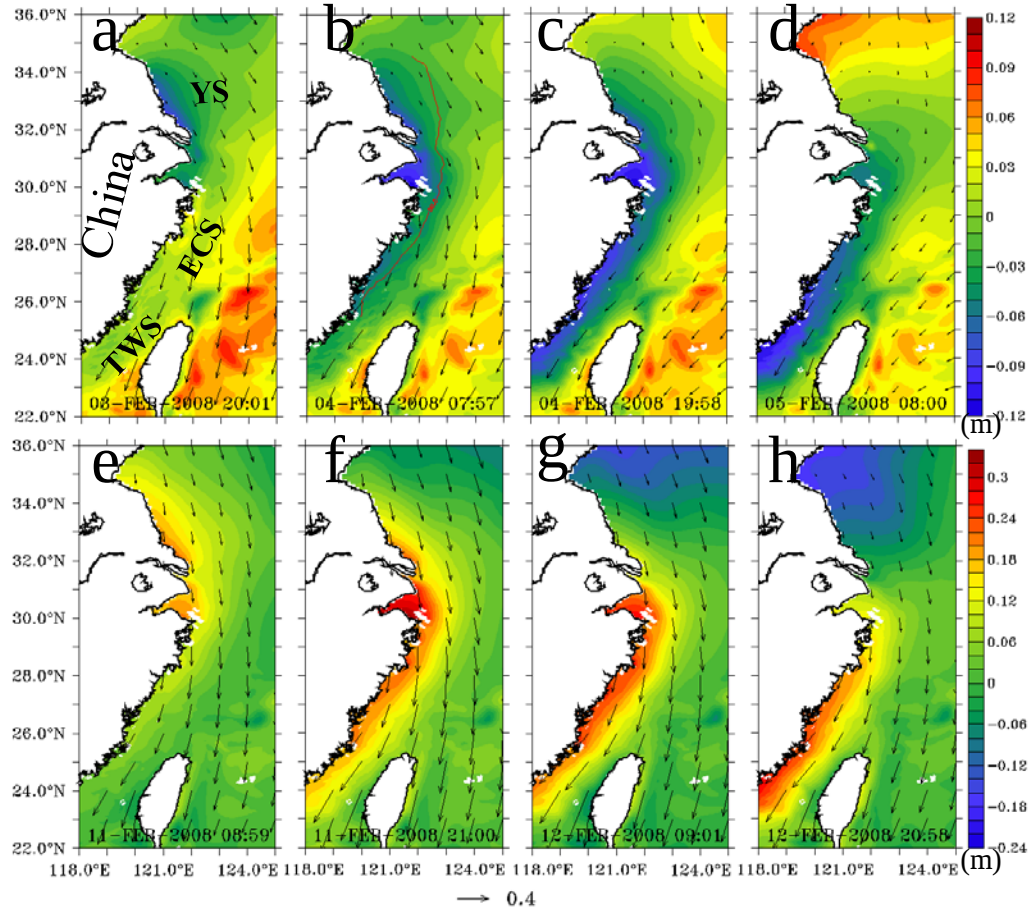


Figure 3.2. The simulated 12-hourly sea level anomaly (tidal filtered sea level minus the monthly mean sea level) along the China coast.

The low and high sea level propagated southward on February 3 (upper panel) and February 11 (lower panel), respectively. The MERCATOR wind stress vectors (units: N/m^2) are superimposed. In the panel a, the YS, ECS, and TWS represent Yellow Sea, East China Sea, and Taiwan Strait, respectively. In the panel b, the red line is the isobaths of 30 m along the China coast.

The 3.3b displays Hilbert spectrum (calculated by EEMD) of the velocities monitored by the buoy in the TWS. The lines represent the high spectra with time. The variation suggests a fluctuation of frequency for nonlinear effect. The spectrum

indicated two primary frequencies of about 0.4 and 0.1 cycles per day (cpd) between February 3 and 5 (the arrival of the CKW trough), and three primary frequencies of about 0.3, 0.2 and 0.1 cpd between February 11 and 12 (the arrival of the CKW crest). These frequencies are similar to the frequencies of previous CKWs studies [Li, 1989; Yin *et al.*, 2014], e.g. the frequencies range from 0.09 to 0.5 cpd. The transport direction, estimated phase speed, and wave frequencies are consistent with previous CKWs studies along the China coast. This indicates the waves can be considered as barotropic coastal-trapped Kelvin waves.

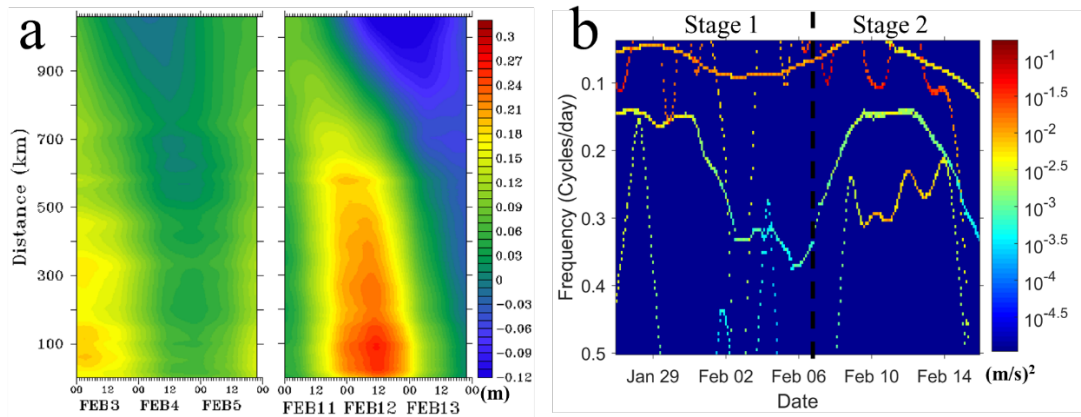


Figure 3.3. The time-distance plot of the alongshore sea level anomaly (tidal filtered sea level minus the monthly mean sea level, a) and Hilbert spectrum of vertical averaged velocity of the buoy (b).

In the panel a, the distance is from Taiwan Strait toward the Yellow Sea along the 30 m isobaths (plotted in 3.2). The low and high sea level originated in the Yellow sea and propagated southward along the isobaths.

Time series of cross-strait sea level difference between the control and non-CKWs cases is illustrated in 3.4a. The sea level difference can be treated as the sea level caused by the CKWs. The negative and positive values appeared alternatively

with time. This means wave trough and crest propagated from the north toward the TWS continually (3.4a). The Rossby deformation radius (280 km) in this region is larger than the strait width (about 200 km) in the TWS. This suggests the sea level, generated by the CKWs, can modify the cross-strait sea level gradient. The strongest negative (positive) values emerged on February 4 (February 12), which matched the arrival time of the wave trough (crest) shown in Figures 3.2 and 3.3.

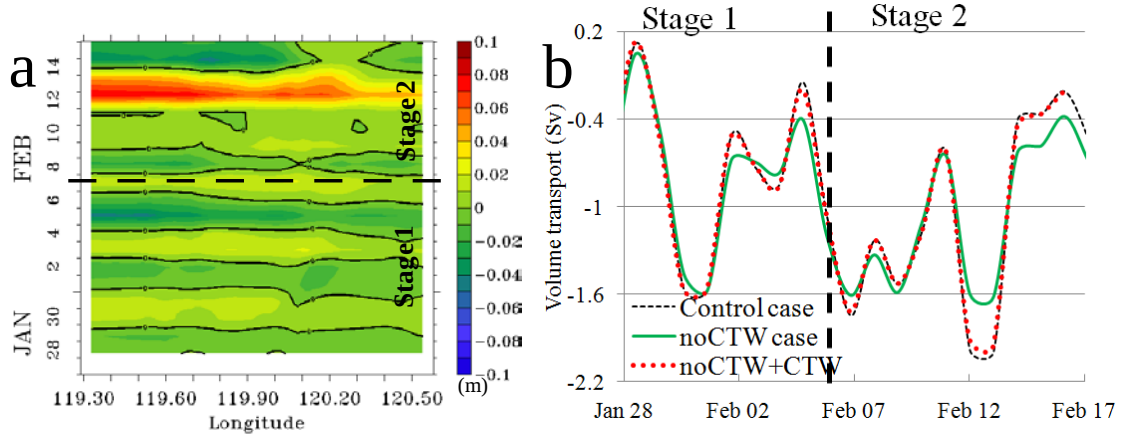


Figure 3.4. The cross-strait sea level difference (the control case minus non-CKWs case, a) and volume transport (b, units: Sv, $\times 10^6 \text{ m}^3/\text{s}$) through a section in the Taiwan Strait.

The position of the section is shown in the 3.1. The red dot line in the panel b is the sum of non-CKWs case and the transport induced by the CKWs (estimated by eq. 1).

Ko et al. [2003] showed the dominant dynamics, governing the volume transport through the TWS in the winter, is in geostrophic balance when the water is well mixed. Although the phase speed of CKWs is very fast (16.8 m/s), the wave length of CKWs is very long (~ 1000 km). Therefore, the passage time is about 20 hours, which is larger than the geostrophic adjustment time ($\sim f^{-1}$, ~ 4.5 hours) in the TWS. This suggests the high and low sea levels can remain in the TWS for enough

time to induce geostrophic flow. Therefore, the volume transport caused by the CKWs can be estimated by the cross-strait sea level difference. The along-strait geostrophic current can be calculated by $v = \frac{g}{f} \frac{\partial \eta}{\partial x}$, and then the volume transport can be estimated by

$$S_v = \frac{10^{-6} g h \Delta \eta}{f} \quad (3.1)$$

where f is Coriolis coefficient ($6.048 \times 10^{-5} \text{ s}^{-1}$), g is gravity acceleration constant, h is averaged depth along the cross-strait section (50 m), $\Delta \eta$ is cross-strait sea level differences (can be estimated from 3.4a), Δx is width of the cross-strait section (175 km) and v is along-strait velocity.

The 3.4b shows time series of the volume transport through the TWS. The red dot line (non-CKWs + CKWs) is the sum of the transport in the non-CKWs case (green dashed line) and the transport estimated by eq. 1 (generated by the CKWs). The sum transport (non-CKWs + CKWs) is almost similar to the transport of the control case (black dashed line). This suggests the transport differences between these two cases were caused by the CKWs. Specifically, the wave crest (trough) could induce a negative (positive) cross-strait sea level gradient through increasing (decreasing) the sea level in the western strait. The negative (positive) cross-strait sea level gradient can strengthen a southward (northward) geostrophic current leading to an increased southward (northward) volume transport.

The large transport differences between the control and non-CKWs cases emerged on February 4 and 12, which are consistent with the arrival times of the wave trough and crest, respectively (3.4b). On February 4, the CKW trough decreased the southward transport from -0.40 Sv to -0.14 Sv through weakening the mean southward

current by 0.03 m/s. In other words, the wave trough increased the northward transport by 0.26 Sv through strengthening the northward flow by 0.03 m/s. On February 12, the wave crest changed the volume transport from -1.61 Sv to -1.99 Sv through intensifying southward geostrophic current from 0.18 m/s to 0.23 m/s. This means the CKW crest could increase the southward transport by 0.38 Sv on February 12.

In summary, a large CKW trough (crest), occurred in the first (second) stage, strengthened the northward (southward) geostrophic current and northward (southward) volume transport through changing the cross-strait sea level gradient. Note that there was a huge impact of the CKWs on the volume transport in 2008, and similar impact can be found in other years as well [Ko *et al.*, 2003]. This indicates the CKWs may play an important role in short-term variation of volume transport in the TWS.

3.5 The Coastal-Kelvin Waves and Intensified Cold Water Intrusion

The PHI, located in the southeastern TWS, is surrounded by warm water from the south almost all through the year [Jan *et al.*, 2006; Liao *et al.*, 2013a]. The huge temperature decrease around the PHI in 2008 was induced by the unique dynamic processes [Liao *et al.*, 2013a], which are affected not only by the local wind stress but by the CKWs. 3.5 illustrates the time series of temperature (Zhangyun Ridge and PHI) and the CKWs propagation. The temperature at the Zhangyun Ridge decreased largely from January 28 to February 7 (the first stage) and increased on February 7-14 (the second stage). The temperature at PHI remained constant around February 4, and decreased significantly after February 7. In the first stage, the decreased temperature at Zhangyun Ridge and constant temperature at PHI coincided with eastward movement of cold water that the separated cold water was moved to the Zhangyun Ridge and did

not affect the PHI. In the second stage, the increased temperature at Zhangyun Ridge and decreased temperature at PHI matched the southward movement of cold water that the separated cold water left Zhangyun Ridge and transported to PHI. Comparison between the control and non-CKWs cases displays the temperature in the first case decreased more than that in the second case. The temperature difference at the Zhangyun Ridge (PHI) happened in February 4-6 (February 12-14) which was the arrival time of the CKW trough (crest). This indicates the temperature differences at Zhangyun Ridge and PHI are related to the CKWs.

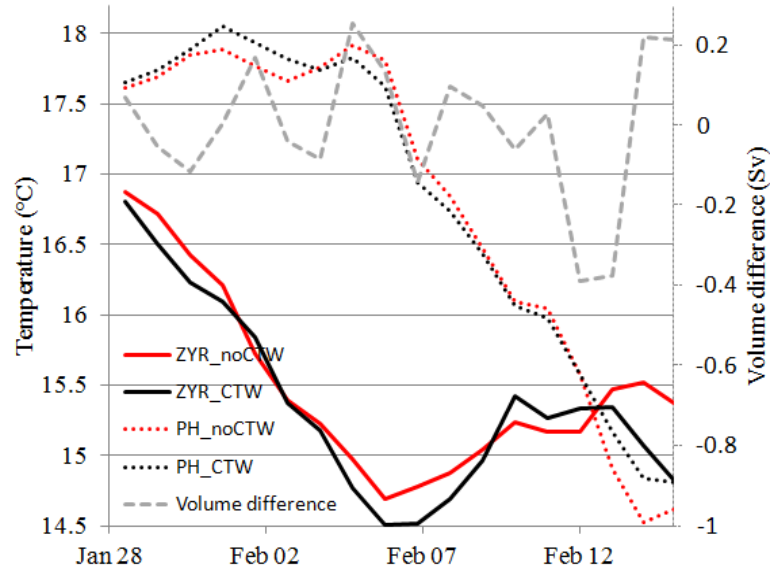


Figure 3.5. The time series of temperature at the Zhangyun Ridge and Penghu Island from model results.

The black (red) line is the control (non-CKW) case. The solid (dashed) line is the temperature at the Zhangyun Ridge (Penghu Island). The grey dashed line is the volume difference between the control case and non-CKW cases. The volume difference is positive (negative) when the wave trough (crest) arrives.

In order to show detailed relation between the CKWs and the temperature differences, the time series of the CKWs propagation and the boundaries of separated cold water are displayed in the 3.6. The boundary of separated cold water is determined by 14.5 °C isotherm. The movement of east (south) boundary can indicate the strength of cross flow (southward flow) which can be affected by the CKW trough (crest). Before the arrival of the CKWs, the east (south) boundaries between the control and non-CKWs cases were almost located at the same longitude (latitude, 3.6). Around February 4, the transport difference (the control case minus the non-CKWs case) increased from -0.01 Sv to 0.25 Sv, suggesting the arrival of the CKW trough. Meanwhile, the east boundary of cold water was moved eastward (120.3 °E) in the control case, while it nearly remained at the same place (120.22 °E) in the non-CKWs case. This indicates the further eastward movement (~8.0 km) in the control case is related to the CKW trough. In February 11-14, the volume difference decreased from -0.01 Sv to -0.39 Sv suggesting the arrival of the CKW crest. The south boundary only reached as south as 23.78 °N in the non-CKWs case, while it arrived at 23.61 °N in the control case. This suggests the further southward movement (~18.9 km) of the cold water was caused by the CKWs crest.

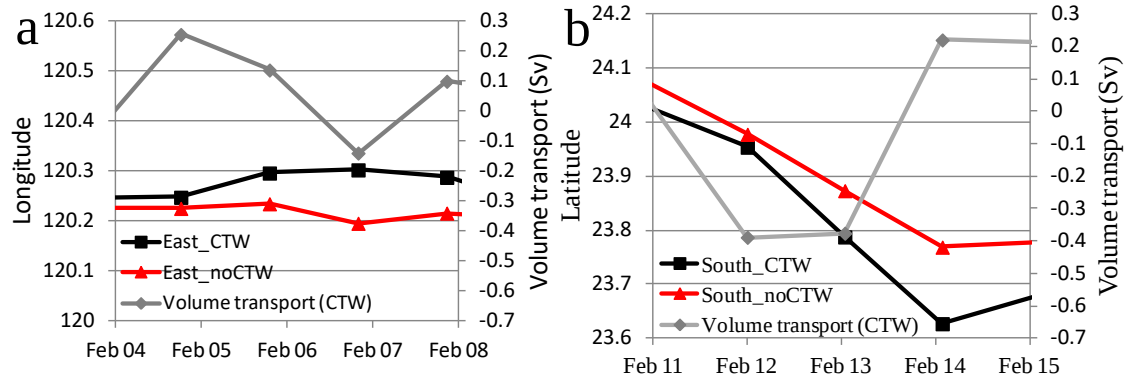


Figure 3.6. The time series of east (a) and south (b) boundary of the separated cold water (indicated by 14.5°C) in the Taiwan Strait. The grey line is the volume difference between the control case and non-CKWs cases. The volume difference is positive (negative) when the wave trough (crest) arrives. The black line is the boundary position in the control case and the red line is the boundary position in the non-CKWs case.

Figures 3.7a and 3.7b reveal the positions of 14.5°C isotherm on February 7 (the end of the first stage) and 14 (the end of the second stage) after the arrival of the CKW trough and crest, respectively. In the 3.7a, the black line (control case) is ~ 8.0 km east of the red line (non-CKWs case) meaning an enhanced eastward movement of cold water induced by the CKW trough. In the 3.7b, the black line is ~ 18.9 km south of the red line, indicating an intensified southward movement of the cold water caused by the CKW crest.

In order to study the dynamic mechanism of the additional movement, the sea level and current patterns induced by the CKWs (control case minus non-CKW case) are plotted in the Figures 3.7c and 3.7d. During the first stage, the dominant flow was a cross-strait flow under moderate northeasterly winds [Liao *et al.*, 2013a]. The arrival of the CKW trough caused a positive cross-strait sea level gradient by decreasing the

sea level in the western strait (3.7c). The positive cross-strait sea level gradient induced an additional northward flow which turned around Zhangyun Ridge forming a cyclone in the north of Zhangyun Ridge (3.7c). Oey *et al.* [2014] shown that a cyclone here contribute to a strongly negative vorticity which results in a cross flow. Therefore, it is the additional cross flow induced by the CKW trough and related cyclone that intensified the eastward (offshore) movement of the cold water. Note that the condition for the additional cross flow caused by the CKW trough is a moderate wind ($0.068\text{--}0.12\text{ N/m}^2$, referred to Oey *et al.*, [2014]). The impact of the CKW trough on the eastward (offshore) movement of the cold water may be not so significant if the northeasterly wind is beyond the bounds of moderate wind (the corresponding flow pattern is a strong northward/southward flow).

During the second stage, the flow was southward with strong northeasterly winds [Liao *et al.*, 2013a]. The arrival of the CKW crest on February 12 generated a negative cross-strait sea level gradient through increasing the sea level in the western strait (3.7d). The negative cross-strait sea level gradient induced an additional southward flow through geostrophic balance (3.7d). Therefore, it is the additional southward current caused by the CKW crest that intensified the southward movement of cold water.

In a word, the CKW trough made contribution to the cross flow and the CKW crest could enhance the southward flow during the first and second stage, respectively. The additional cross flow and southward flow intensified the eastward and southward movements of cold water, respectively. A schematic figure of the dynamic processes is shown in the 3.8. From the model results in the 3.7, if the CKWs were removed, the cold water ($14.5\text{ }^{\circ}\text{C}$) cannot reach the PHI and the cold disaster may be avoided. This

suggests the additional movements caused by the proceeding CKWs trough and crest are critical for the occurrence of cold disaster.

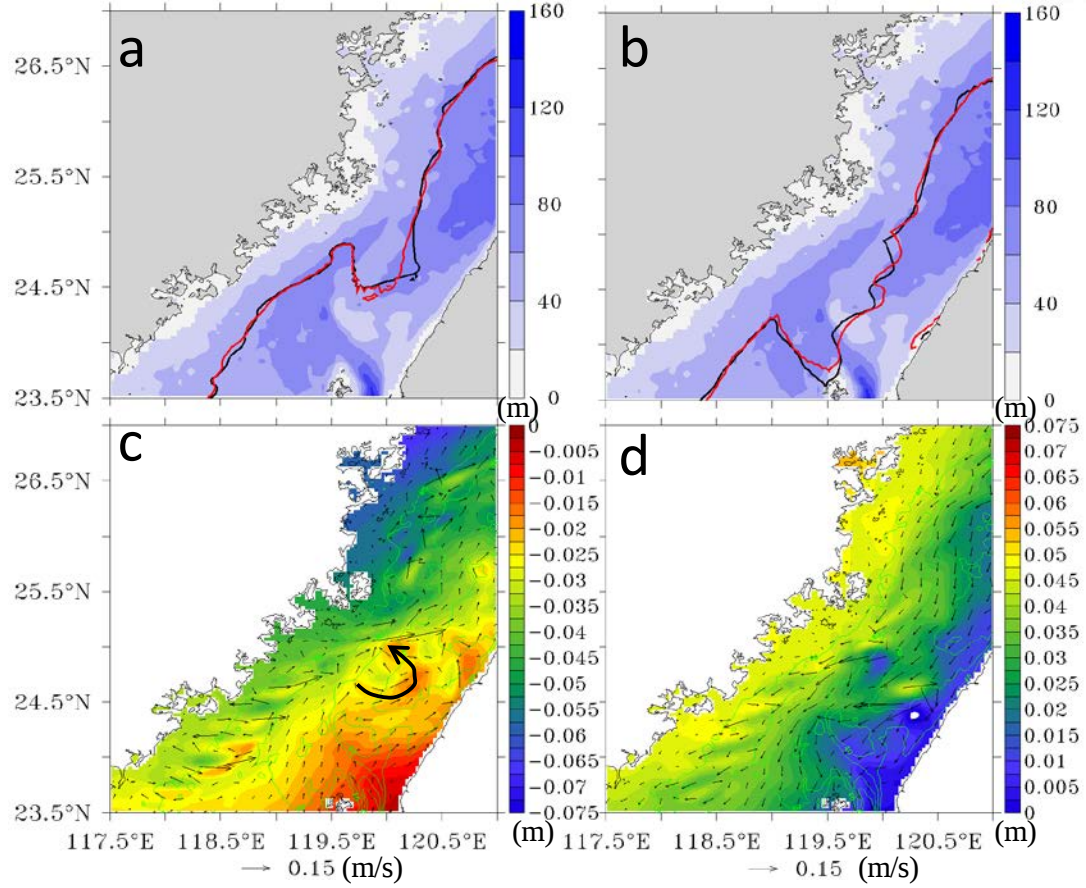


Figure 3.7. The position of 14.5 °C isotherm (a and b) and the sea level and flow differences (c and d) after the arrival of the CKW trough (a and c) and crest (b and d). The difference is the result in the control case minus the result in the non-CKW case. The black and red lines in the panel a and b are the control case and non-CKWs case, respectively. The background shadings (a and b) and thin green lines (c and d) show the bottom topography (units: m). The anti-clockwise arrow denotes the cyclonic eddy caused by northward flow.

3.6 Conclusion

In winter, the CKWs, as the third factor, constantly propagating toward the TWS from north, can impact flow patterns in the TWS. The arrival of a CKW trough (crest) in the western coast can strengthen a northward (southward) geostrophic current through changing the cross-strait sea level gradient. The strengthened northward (southward) current can amplify a northward (southward) volume transport.

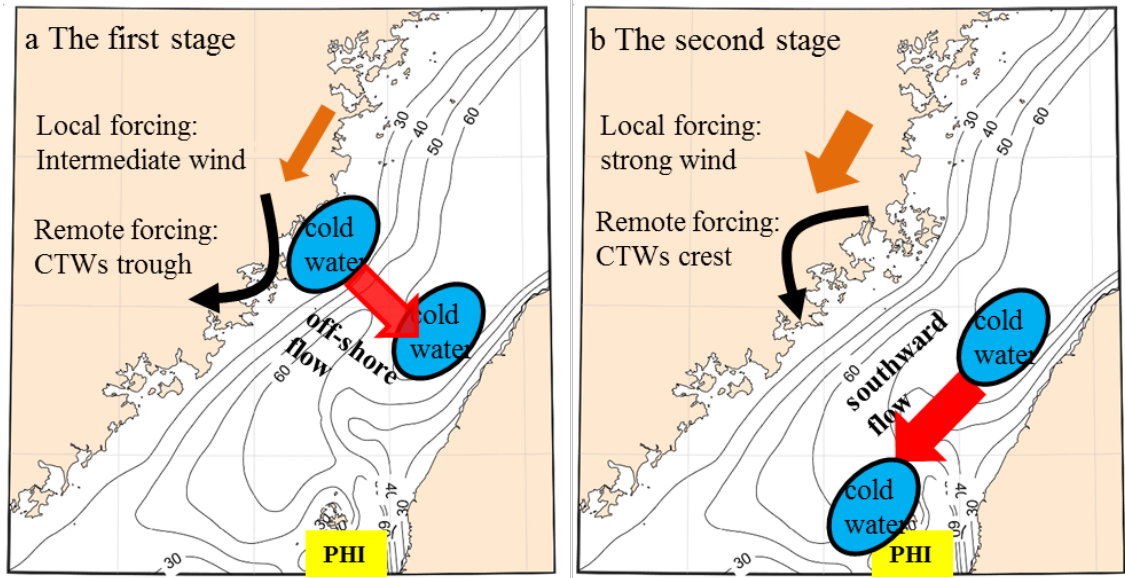


Figure 3.8. The schematic diagrams illustrating the processes of cold disaster. a is the off-shore movement caused by the intermediate wind and CKWs trough, b is the southward movement caused by the strong wind and CKWs crest. The wind is referred to as the local forcing and the CKWs is remote forcing.

During the first stage, the northeasterly wind was moderate resulting in a cross-strait flow pattern. Around February 4, the arrival of a CKW trough induced an additional northward flow. When the northward flow turned around the Zhangyun Ridge, a cyclone was formed in the north of Zhangyun Ridge. This cyclone led to an additional cross flow, which intensified the eastward (offshore) movement of cold

water. In the second stage, the northeasterly wind was very strong leading to a southward flow pattern. The reach of a CKW crest generated an additional southward current which increased the southward movement of cold water. The additional eastward and southward flow caused by the CKWs played an important role in transporting the cold water toward PHI killing the tropical fishes there.

The impact of volume transport variations caused by CKWs in the TWS on other ocean phenomena is also worthy to investigate. For example, the volume transport in the TWS is southward (northward) in winter (summer) for northeasterly (southwesterly) monsoon. *Moon and Hirose* [2014] showed the seasonal transport variation in the TWS can impact the Kuroshio water intrusion in the East China Sea. Therefore, it is worth to study that if the short-term transports variation caused by CKWs can impact the Kuroshio water intrusion in the East China Sea.

Chapter 4

THE COASTAL OCEAN RESPONSE TO THE GLOBAL WARMING ACCELERATION AND HIATUS

4.1 Abstract

Coastlines are fundamental to humans for habitation, commerce, and natural resources. Many coastal ecosystem disasters, caused by extreme sea surface temperature (SST), were reported when the global climate shifted from global warming to global surface warming hiatus after 1998. The task of understanding the coastal SST variations within the global context is an urgent matter. Our study on the global coastal SST from 1982 to 2013 revealed a significant cooling trend in the low and mid latitudes (31.4% of the global coastlines) after 1998, while 17.9% of the global coastlines changed from a cooling trend to a warming trend concurrently. The trend reversals in the Northern Pacific and Atlantic coincided with the phase shift of Pacific Decadal Oscillation and North Atlantic Oscillation, respectively. These coastal SST changes are larger than the changes of the global mean and open ocean, resulting in a fast increase of extremely hot/cold days, and thus extremely hot/cold events. Meanwhile, a continuous increase of SST was detected for a considerable portion of coastlines (46.7%) with a strengthened warming along the coastlines in the high northern latitudes. This suggests the warming still continued and strengthened in some regions after 1998, but with a weaker pattern in the low and mid latitudes.

4.2 Introduction

Despite the continued increase of atmospheric greenhouse gases, the global surface mean temperature has remained flat since 1998[Easterling and Wehner, 2009; Kosaka and Xie, 2013]. The global climate exhibited a shift from a rapid global surface warming to an unexpected deceleration in the global surface warming with a notable cooling in the eastern tropical Pacific Ocean (a La-Niña-like pattern) and a strengthening of the trade winds[England et al., 2014; Kosaka and Xie, 2013; Trenberth et al., 2014a]. This change in the acceleration of the global surface warming has received much notice with the title of "global surface warming hiatus"[Easterling and Wehner, 2009; England et al., 2014; Kosaka and Xie, 2013; Trenberth et al., 2014b]. To be consistent with previous studies, we refer to this time period (after 1998) as the hiatus period. Due to heterogeneously distributed aerosols and changes in the atmospheric/oceanic circulation, the regional ocean responses (warming/cooling) are not uniform[Levitus, 2005; Ramanathan et al., 2001]. Recent studies reported that the main locations of the warming hiatus are in the central and eastern Pacific Ocean and over the northern continents, especially Eurasia. This indicates that there is a regionality to the impact of the global climate shift [Cohen et al., 2012; Trenberth and Fasullo, 2013; Trenberth et al., 2014a]. Lima and Wethey [2012]reported a high warming rate for 71.6% of the global coastlines during the last three decades (1982-2010). However, there is limited information on the variations of warming/cooling for the global coastlines during the hiatus period. Since strong scientific interest has focused on the mechanism of the global surface warming hiatus[Kaufmann et al., 2011; Kosaka and Xie, 2013; Meehl et al., 2011; Solomon et al., 2011] and few studies have paid attention to the coastal effect, it is imperative to study the corresponding coastal effects of the global surface warming hiatus.

Approximately 50% of the world's population lives within 200 kilometers of coast and nearly half of the ocean's ecosystem goods and services come from coastal marine ecosystems [Costanza *et al.*, 1997; Hinrichsen, 1999]. This vital region is highly influenced by a multitude of forcing mechanisms (i.e. anthropogenic processes, sediment transport, ocean circulation, wind stress, and large scale climate variability) [Shearman and Lentz, 2010]. Recently, extreme sea surface temperatures (SST) occurred in many coastal areas and caused serious physiological stress and large casualties to various coastal ecosystems [Chen *et al.*, 2014a; Colella *et al.*, 2012; Eakin *et al.*, 2010; Feng *et al.*, 2013; Garrabou *et al.*, 2009; Hsieh *et al.*, 2008; Liao *et al.*, 2013a; Mills *et al.*, 2013a]. These extreme SSTs (i.e., extremely cold and hot events) were proven to be difficult for temperature-sensitive coastal coral reef ecosystems, some of the most diverse and productive communities on the earth [Selig *et al.*, 2010]. For instance, an extremely hot event was observed in the Caribbean Sea (2005), and extremely cold events were witnessed in the Taiwan Strait (the China coast, 2008) and the Florida Keys (the eastern U.S. coast, 2010) [Colella *et al.*, 2012; Eakin *et al.*, 2010; Hsieh *et al.*, 2008; Liao *et al.*, 2013a]. Additionally, the extreme SSTs can also cause disastrous impacts on non-coral-reef ecosystems (e.g., the western coast of Australia in 2011 and the northeastern U.S. coast in 2012) [Feng *et al.*, 2013; Mills *et al.*, 2013a] and rocky-reef communities (e.g., the Mediterranean in 2003) [Garrabou *et al.*, 2009]. Therefore, the understanding and prediction of the effects of the global surface warming hiatus on the coast are beneficial to the scientific field, policy makers, and general population. By studying the variability of coastal SST, extremely hot days (EHDs), and extremely cold days (ECDs) in a global context, this study details how global surface warming hiatus translates into regional patterns.

4.3 Results

Time series of yearly SST. The SST data is separated into two different periods (1982-1997, hereafter called warming period; 1998-2013, hereafter called hiatus period) for analysis. Globally, the trend of global mean coastal SST was $0.17 \pm 0.11^\circ\text{C}/\text{decade}$ (95% confidence interval) in the warming period and reduced to $0.11 \pm 0.09^\circ\text{C}/\text{decade}$ during the hiatus period. In order to better depict variations in these two time periods, yearly SSTs are filtered using a 10-year moving average (4.1). Specifically, temporal variations in these two periods can be classified into four patterns (4.1) that are **(a)** SST increased in the warming period ($0.26 \pm 0.10^\circ\text{C}/\text{decade}$, mean trend), and then decreased in the hiatus period ($-0.24 \pm 0.11^\circ\text{C}/\text{decade}$), which is P1 in the 4.1a and blue star mark in the 4.1b; **(b)** SST first decreased ($-0.11 \pm 0.08^\circ\text{C}/\text{decade}$), and then increased ($0.27 \pm 0.09^\circ\text{C}/\text{decade}$), which is P2 in the 4.1a and black plus mark in the 4.1b; **(c)** SST continued decreasing (-0.07 ± 0.06 and $-0.09 \pm 0.11^\circ\text{C}/\text{decade}$), which is P3 in the 4.1a and green circle mark in the 4.1b; **(d)** SST continued increasing (0.20 ± 0.11 and $0.31 \pm 0.11^\circ\text{C}/\text{decade}$), which is P4 in the 4.1a and red x mark in the 4.1b. The first pattern occupied 31.4% of the global coastlines (4.1b) and primarily distributed in the low and mid latitudes (60°S - 60°N , defined in the Method section). This pattern changing from warming to cooling is compatible with the present global climate shift. The second pattern is discontinuously distributed in 17.9% of the global coastlines (e.g., the western tropical Pacific and Indian Ocean, 4.1b). The third pattern primarily appeared in the Polar Regions. The fourth pattern (continued warming), corresponding with the global warming, is located in nearly half of the global coastlines (46.7%, 4.1b). These findings suggest that, while coastal warming was still the major pattern during the hiatus period, cooling was observed

along numerous coastlines (35.4%), with a majority of the cooling coastlines occurring in the low and mid latitudes.

Global patterns. The patterns of the global coastal SST trends (4.2) demonstrate a notable warming in 68.78% of the coastlines in the low and mid latitudes in the warming period. In contrast, during the hiatus period, a significant coastal cooling was detected in nearly half (48.33%) of the coastlines in the low and mid latitudes, while a significant warming occurred in 82.42% of the coastlines in the high northern latitudes (60°N-90°N). Compared with warming period, the area ratio of global coastal warming in the hiatus period reduced from 73.19% to 64.60%.

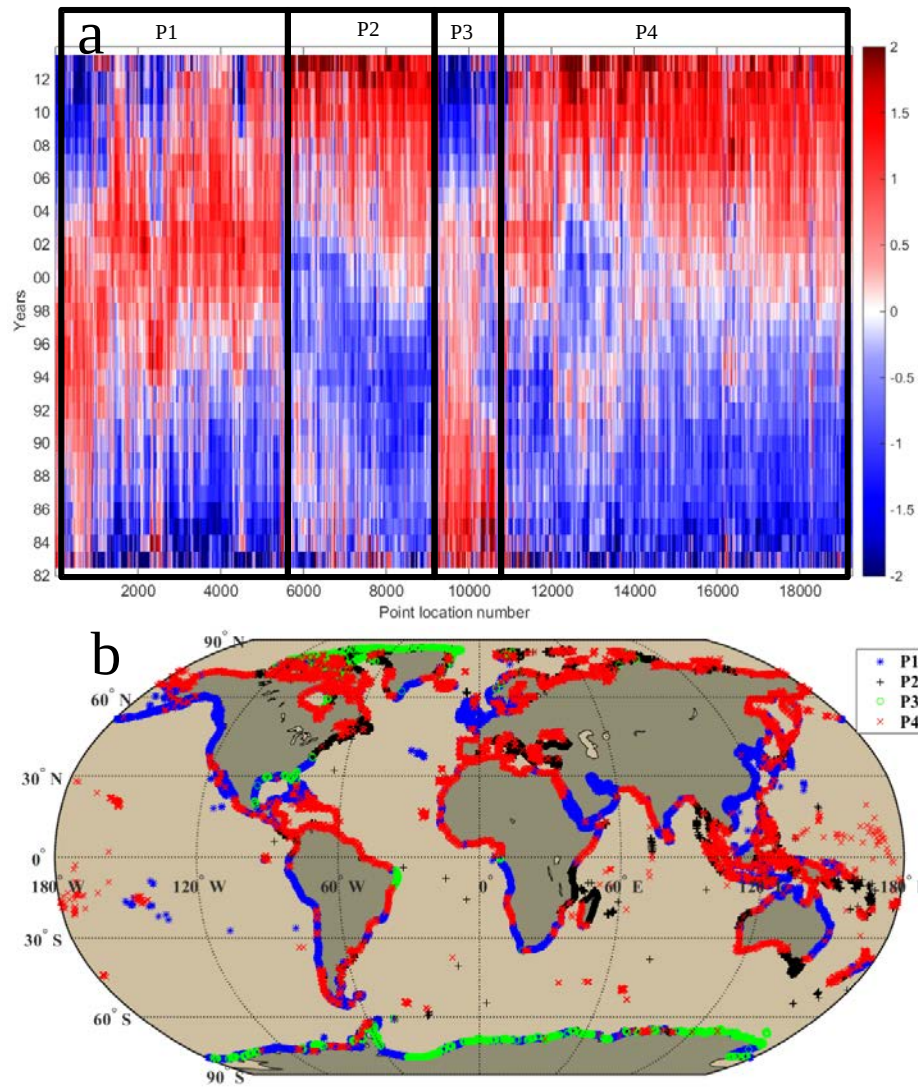


Figure 4.1. The time series of standardized yearly SST and four changing patterns. (a) The time series of standardized yearly SST at each coastal location (total 19276), sorted by four changing patterns. (b) The distribution of four changing patterns along the world's coastlines. The first pattern is that the SST increased in the warming period (1982-1997), then decreased in the hiatus period (1998-2013), which is P1 in the panel a and blue star mark in the panel b. The second pattern is SST decreased in the first period, then increased in the second period, which is P2 in the panel a and black plus mark in the panel b. The third pattern is SST continued decreasing, which is P3 in the panel a and green circle mark in the panel b. The fourth pattern is SST continued increasing, which is P4

in the panel a and red x mark in the panel b. See Method section for the methods to obtain the yearly SST time series.

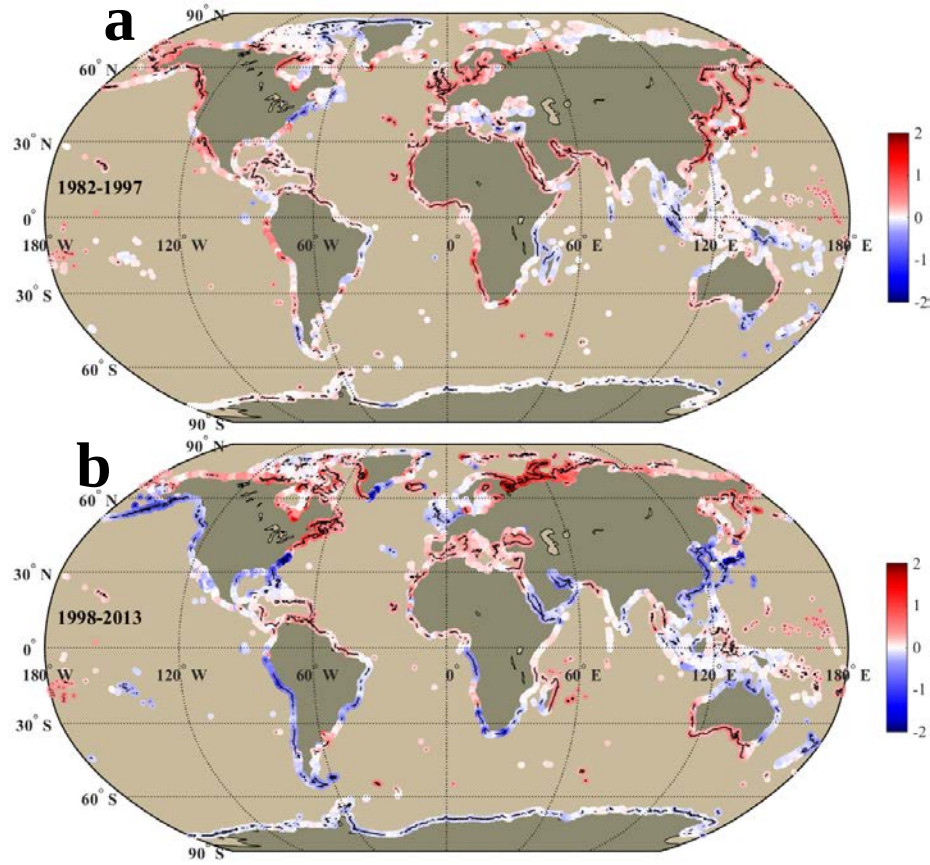


Figure 4.2 The linear SST trends (unit: °C/decade) along the world's coastlines in the warming (1982-1997, a) and hiatus periods (1998-2013, b). Black points/lines in the shading color indicate the trends in those locations are significant in statistics ($P < 0.05$).

Figures. 4.3 and 4.4 exhibit linear trends of annual frequency of EHDs and ECDs, which are positively and negatively correlated with SST trends, respectively. The correlation coefficients between SST trends and trends of EHDs and ECDs were

0.53 (EHDs) and -0.77 (ECDs) in the warming period and 0.73 (EHDs) and -0.62 (ECDs) in the hiatus period respectively ($P < 0.01$). Therefore, the region with higher warming (cooling) rate tends to have more EHDs (ECDs), suggesting a rising probability of occurrence of extremely hot (cold) event. However, the moderate correlations (0.53-0.77) indicate the trends of EHDs and ECDs along some coastlines may be influenced by other factors (e.g., SST variance) in addition to the SST trends. The positive trends of EHDs occupied almost half of the global coastlines with similar area ratios (51.8% and 51.9%) in these two periods, but different distributions. Compared with the warming period, the positive trends of EHDs in the hiatus period were mainly observed in the high northern latitudes (4.3). The area ratio of ECDs with positive trend in the global coast increased from 23.9% in the warming period to 44.6% in the hiatus period. The increase in the number of ECDs primarily occurred in the low and mid latitudes, notably in the Pacific Ocean (4.4) during the hiatus period. This observation intimates a modified regional pattern of warming and cooling after 1998: a significant cooling along the coast of the low and mid latitudes, principally in the Pacific Ocean, and a notable warming along the coasts of the high northern latitudes even though there was a hiatus in the increase of the global surface mean temperature. As a result, the corresponding numbers of EHDs and ECDs increased rapidly along these warming (high northern latitudes) and cooling (low and mid latitudes) coasts, leading to a higher probability of extremely hot and cold events, respectively.

Pacific. In the Pacific Ocean, the coasts (excluding the Antarctica region) grew warmer (~ 0.17 °C/decade, trend of mean SST in this area) in the warming period and colder (~ -0.05 °C/decade) in the hiatus period (4.2). The changes in the SST trends in

the Pacific Ocean are the largest of all the ocean basins. These variations are in keeping with the present report that the hiatus is associated with the Pacific Ocean [England *et al.*, 2014; Kosaka and Xie, 2013]. The overarching distribution of the coastal SST trends (4.2b) in the hiatus period were an eastern coastal cooling and a western coastal warming in the North Pacific (except for the China and Japan), and a coastal cooling in the South Pacific (except for the Maritime Continent and South Pacific Islands). The pattern in the Pacific may be related to the present climate features (the negative PDO phase and eastern tropical Pacific cooling) [Kosaka and Xie, 2013; Trenberth and Fasullo, 2013]. In the western Pacific, the cooling trend along the China and Japan coast (-0.69 ± 0.44 °C/decade), opposing to the overarching distribution, implies other influencing factors (e.g., the East Asian Monsoon) [Ding *et al.*, 2014].

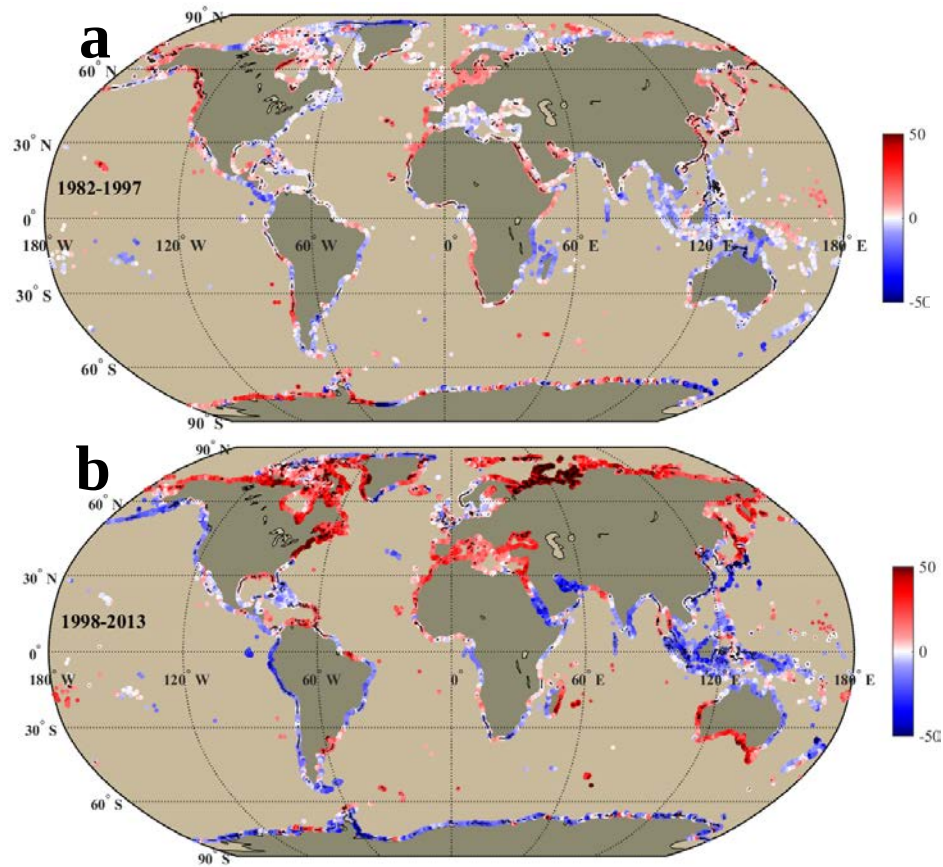


Figure 4.3 Same as 4.2, but shows the linear trend (days/decade) of annual frequency of extremely hot days in the warming (1982-1997, a) and hiatus periods (1998-2013, b).

Black points/lines in the shading color indicate the trends in those locations are significant in statistics ($P < 0.05$).

The number of EHDs (4.3) in the hiatus period increased significantly for warming coastlines (e.g., the Northwestern Pacific, 21.46 ± 7.90 days/decade). The number of ECDs (4.4) increased along cooling coastlines, particularly for the coastlines of China and Japan (25.30 ± 11.75 days/decade), the Alaskan Peninsula (21.13 ± 19.74 days/decade), and the western coastlines of South America (19.27 ± 11.98 days/decade). The China coast witnessed a distinct reversal, in that the

trend of EHDs decreased from 8.33 ± 7.22 to -6.53 ± 5.01 days/decade (4.3) and ECDs increased from -36.57 ± 23.29 to 25.30 ± 11.75 days/decade (4.4). The consequence of the increasing trend of ECDs is a higher probability of an extreme cold event (e.g., the Taiwan Strait cold disaster in 2008[Liao *et al.*, 2013a]). If this trend persists, the frequency and intensity of extreme events may increase, leading to ecological and economic instability for these regions. The Alaskan Peninsula coast is another example of a trend reversal of EHDs and ECDs (see detailed trend values in Table B1 in Appendix B).

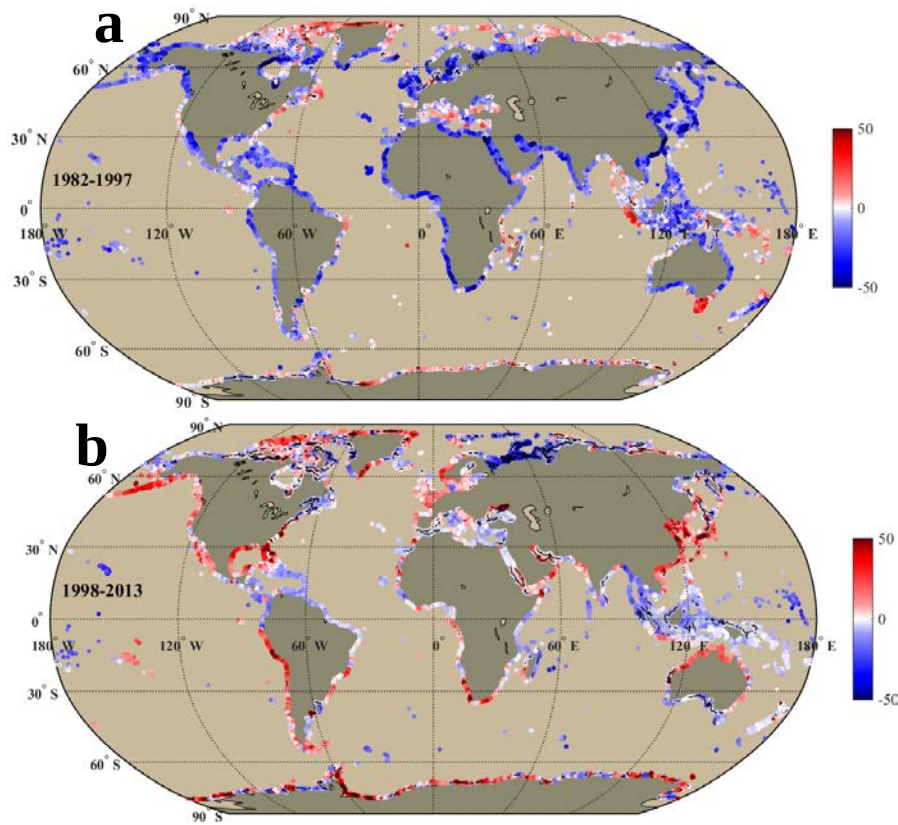


Figure 4.4 Same as 4.2, but shows the linear trend (days/decade) of annual frequency of extremely cold days in the warming (1982-1997, a) and hiatus periods (1998-2013, b). Black points/lines in the shading color indicate the trends in those locations are significant in statistics ($P < 0.05$).

Atlantic. In the Atlantic Ocean (excluding the Antarctic region), during the warming period, coastal warming (~ 0.16 °C/decade) was the characteristic feature (4.2a). In the hiatus period, although coastal warming (~ 0.09 °C/decade) was still the dominant feature, coastal cooling was observed in part of the North Atlantic and most of the South Atlantic. During the hiatus period, the North Atlantic pattern of the coastal SST trends (4.2b) was a tri-polar pattern, with a warming SST trend (0.54 ± 0.24 °C/decade) in the north (north of Cape Hatteras and the Labrador Sea), a cooling SST trend (-0.37 ± 0.33 °C/decade) in the middle (south of Cape Hatteras, the southeastern Greenland Island, and the North Sea), and a warming SST trend (~ 0.15 °C/decade) in the south (the tropical Atlantic). The tri-polar pattern may be associated with the present negative phase of the North Atlantic Oscillation (NAO) [Goldenberg *et al.*, 2001; Grossmann and Klotzbach, 2009; Li *et al.*, 2014; Marshall *et al.*, 2001]. In the South Atlantic, the distribution of coastal warming and cooling areas were discontinuous, indicating different influencing factors in different coastal areas.

The corresponding number of EHDs in the North Atlantic increased significantly in the eastern part of North America (north of Cape Hatteras, 35.41 ± 17.94 days/decade), the Labrador Sea (26.33 ± 16.26 days/decade), the Caribbean Sea (~ 17.5 days/decade), and the Mediterranean Sea (18.52 ± 10.39 days/decade, 4.3). The increase of ECDs was exhibited in the eastern part of North America (south of Cape Hatteras, 27.2 ± 24.7 days/decade), and along most of the South Atlantic coast. The trends of EHDs and ECDs along the eastern North American coast reversed at the Cape Hatteras, where the EHDs increased from -3.19 ± 4.19 to 35.41 ± 17.94 days/decade to the north of Cape Hatteras (4.3), and the ECDs increased from -11.4 ± 14.1 to 27.2 ± 24.7 days/decade to the south of Cape Hatteras during the

hiatus period (4.4). Coinciding with the increase of EHDs and ECDs, an extremely hot (cold) event occurred to the north (south) Cape Hatteras during the hiatus period [Chen *et al.*, 2014a; Colella *et al.*, 2012; Mills *et al.*, 2013a]. In the Caribbean Sea, the trend of EHDs increased from ~ -0.2 to ~ 17.5 days/decade, inducing a rising probability of extremely hot events (e.g., the 2005 coral reef bleaching event in the Caribbean Sea[Eakin *et al.*, 2010]). In the South Atlantic, the largest increase of EHDs appeared along the Argentina coast (20.4 ± 7.4 days/decade) and the largest increase of ECDs occurred along the western coast of South Africa (29.6 ± 14.6 days/decade).

Indian Ocean. In the Indian Ocean, the coastlines (excluding the Antarctic region) revealed reverse SST trends between the warming and hiatus periods (4.2). In the warming period, coastal warming rate was high (~ 0.14 °C/decade) in the Northern Indian Ocean, and low (~ 0.01 °C/decade) in the South Indian Ocean. However, in the hiatus period, this pattern was reversed. The coastal warming rate was low (negative, ~ -0.03 °C/decade) in the North Indian Ocean and high (~ 0.07 °C/decade) in the South Indian Ocean. Specifically, in the hiatus period, the most significant coastal cooling occurred in the Red Sea (~ -0.11 °C/decade) and Persian Gulf (-0.36 ± 0.29 °C/decade), while the most significant coastal warming happened along the eastern portion of Madagascar Island (0.33 ± 0.16 °C/decade) and the western and southern part of Australian coast (0.33 ± 0.21 °C/decade). The significant coastal warming in the western part of Australia is likely influenced by the recent swing to the negative phase of the Interdecadal Pacific Oscillation (IPO) and enhanced El Niño Southern Oscillation variance since 1970 [Doi *et al.*, 2015; Feng *et al.*, 2015]. However, in the North Indian Ocean, there may be other factors influencing the coastal SST. Raitsos *et al.* [2011] indicated air temperature is the key parameter that influenced the Red Sea

SST, which implied the SST cooling during the hiatus period in the Red Sea may be related to atmospheric forcing.

During the hiatus period, the corresponding number of EHDs increased along the coastlines around the South Indian Ocean, especially for the eastern shore of Madagascar Island (22.8 ± 18.7 days/decade) and the western and southern part of Australian coast (~ 28 days/decade, 4.3). As a result, there is an increased probability of an extremely hot event in these coastal areas (e.g., the 2011 extremely hot event in the western part of Australia [Feng *et al.*, 2013]). However, the mean number of ECDs in the North Indian Ocean showed a negative trend (~ -0.9 days/decade). This reduction in ECDs is inconsistent with the cooling trend (~ -0.03 °C/decade). The inconsistencies suggest a decreased SST variance or standard deviation in the hiatus period (see FigureB 1 in Appendix B). However, further studies are needed to explain the decreased SST variance or standard deviation. The significant increase in the number of ECDs appeared along the northern coast of Australia (~ 17.7 days/decade) where the cooling trend was ~ -0.14 °C/decade (4.4).

Polar Regions. In the Arctic Ocean, SST trends (0.37 ± 0.30 °C/decade) and trends of EHDs (24.8 ± 5.4 days/decade) increased very significantly in the hiatus period, which is consistent with reports of remarkably high temperature trends and ice reduction trends [Comiso *et al.*, 2008; Tingley and Huybers, 2013]. The rapid coastal warming in the Arctic Ocean indicates a significant warming signal in the high northern latitudes [Miller *et al.*, 2013]. This signifies an extremely hot event might happen if the current trend persists. In the Southern Ocean, the Antarctic coast showed a moderate cooling (-0.08 ± 0.04 °C/decade) which corresponds the recent report that the sea ice expanded slightly over the last decade [Comiso and Nishio, 2008]. The

number of ECDs increased (10.1 ± 5.9 days/decade) in the Antarctic coast during the hiatus period.

4.4 Discussion

Lima and Wetthey [2012] reported that the last three decades (1982-2010) witnessed a high warming rate in 71.6% of the global coastlines. However, with the same data set, the coastal SST trends illustrate exceptional trend reversals when the last three decades are separated into the warming (1982-1997) and hiatus (1998-2013) periods. In the warming period, most of the global coastlines (68.19%) showed a warming trend consistent with global warming. However, during the hiatus period, the warming trend reversed in many coastlines of the low and mid latitudes (31.4% of the global coastlines). The distribution of these trend reversals matches with multiple reports that the global surface warming hiatus is mainly concentrated in the low and mid latitudes, especially in the Pacific Ocean [*England et al.*, 2014; *Kosaka and Xie*, 2013].

Additionally, 17.9% of the global coastlines changed from cooling to warming in the hiatus period. These trend reversals, distributed in the low and mid latitudes, may be attributed to natural climate variability instead of anthropogenic-induced warming. There may be two reverse SST variability patterns (i.e., from warming to cooling and from cooling to warming) in response to the global surface warming hiatus. However, further study is needed to distinguish the exact causes of these coastal trends.

Although SST trend reversals are apparent for many coastlines (49.3%, $31.4\% + 17.9\%$), a continuous increase of SST could be detected for a considerable portion of the global coastlines (46.7%). Notably, the high northern latitudes

demonstrated an accelerated warming even though in the hiatus period. The continued warming was concentrated in the high northern latitudes, while the effect of global climate hiatus was focused in the low and mid latitudes.

The timing of the warming/cooling trend reversal in the North Pacific and North Atlantic was in conjunction with the recent phase shifts of the PDO and the NAO. This suggests the coastal cooling around the Eastern Pacific and the coastal warming along the coast of the western part of Australia are likely influenced by the recent swing to the negative phase of the PDO [Ding *et al.*, 2014; Doi *et al.*, 2015; Feng *et al.*, 2015; Kosaka and Xie, 2013; Trenberth and Fasullo, 2013]. The tri-polar pattern (warming SST trends in the north and south, and a cooling SST trend in the middle) in the North Atlantic may be associated with the present negative phase of the North Atlantic Oscillation (NAO) [Goldenberg *et al.*, 2001; Grossmann and Klotzbach, 2009; Li *et al.*, 2014; Marshall *et al.*, 2001]. Additionally, other factors (e.g., ocean circulation and monsoons) may play a significant role in the SST variability [Li *et al.*, 2015]. For example, the cooling of the China and Japan coasts does not match with the negative phase of the PDO and may be related to the recent strengthening of East Asian Winter Monsoon [Ding *et al.*, 2014]. Therefore, further study is needed to detail the mechanisms of SST variation for different coastlines.

The corresponding EHDs and ECDs trends have similar variations with the SST trend spatially and temporally. Spatially, the increase of EHDs was observed in the warming coast (e.g., the eastern portion of Madagascar Island, western and southern part of Australia, Caribbean Sea, Mediterranean Sea, eastern part of North America (north of Cape Hatteras), and much of the Arctic Ocean) in the hiatus period (4.3). The number of ECDs increased in the cooling coast (the Eastern Pacific, China

and Japan, northern part of Australia, Red sea, Persian Gulf, South Africa, eastern part of North America (north of Cape Hatteras), and some coastlines in both polar regions) in the hiatus period (4.4). Temporally, the trend of global mean EHDs (ECDs) was ~ -0.74 (-10.5 ± 8.89) days/decade in the warming period and increased to ~ 3.96 (~ 1.65) days/decade in the hiatus period. In contrast to the warming period, the sum of EHDs and ECDs increased globally in the hiatus period (Figures. 4.3 and 4.4). This means the probability of extreme events (hot or cold) might increase if the global surface warming hiatus continues. The increased sum of EHDs and ECDs may be related to a rising coastal area of large positive and negative SST trends from the warming period to the hiatus period (see FigureB 2 in Appendix B).

In addition to the SST trend, SST variance may influence the variation of EHDs (ECDs). For example, the mean number of ECDs showed a negative trend (~ -0.9 days/decade) when the coast grew colder (~ -0.03 °C/decade) in the North Indian Ocean. These inconsistencies along the coastlines around the North Indian Ocean illustrate a decreased standard deviation from the warming period to hiatus period (see FigureB 1 in Appendix B). This suggests the reduced number of ECDs for cooling coastlines is associated with a decreased standard deviation. Due to the same reason, there may be a reduced number of EHDs for coastlines with an increased SST (see FigureB 1 in Appendix B). The decreased standard deviation indicates a less varied SST, which can be shown by the change in the probability density functions between these two periods (see FigureB 3 in Appendix B). If the standard deviation increases, the ECDs (EHDs) might increase even if the SST increases (decreases) along the inconsistent coastlines (see FigureB 4 in Appendix B). The standard deviation mainly decreased in the low latitude and increased in the mid and high latitudes from the

warming period to the hiatus period (see FigureB 5 in Appendix B). These coastlines with increased standard deviation are susceptible to extreme hot (cold) events, even if these coastlines are cooling (warming).

The global mean surface air temperature shows an increase of approximately 0.05 ± 0.1 °C/decade from 1998 to 2012, when described by a linear trend [Hartmann *et al.*, 2013]. The eastern tropical Pacific Ocean warms by approximately 0.08-0.1 °C/decade from 1900 to 2008, similar in magnitude to the tropical Indian Ocean and central tropical Atlantic Ocean [Deser *et al.*, 2010]. However, the SST trend was -0.69 ± 0.44 °C/decade along the China Coast, 0.54 ± 0.24 °C/decade along the eastern North American coast, 0.33 ± 0.21 °C/decade along the western coast of Australia in the hiatus period. In contrast to the global mean and open ocean, coastal areas have large and heterogeneous responses, which can have a significant, complicated impact on coastal ecosystems. For example, the temperature-sensitive coral reefs, which normally live above 16 °C, experienced temperatures as low as 11.73 °C during a cold event in the Taiwan Strait (China Coast) in 2008. This event resulted in large casualties to the entire coastal ecosystem [Hsieh *et al.*, 2008]. A rise of 0.1 °C in the Caribbean Sea can trigger 35% and 42% increases in the geographic extent and intensity of coral bleaching, respectively [McWilliams *et al.*, 2005]. By providing an understanding of the SST variability and its consequences in a global context, this study is significant to understanding spatial pattern of extreme event, designing conservation strategies, and mitigating negative ecosystem responses.

Chapter 5

A STRENGTHENED SOUTHWARD HEAT TRANSPORT IN THE INDIAN OCEAN DURING THE GLOBAL SURFACE WARMING HIATUS

5.1 Abstract

Since the end of the twentieth century, the global mean surface temperature (GMST) exhibited a shift from a rapid warming to an unexpected deceleration [Easterling and Wehner, 2009]. The GMST phenomenon as a surface characteristic does not represent a slowdown in warming of the climate system but rather is an energy redistribution within the oceans [Yan *et al.*, 2016]. One possible explanation is the change was induced by the heat sequestration from the atmosphere into the upper layer of equatorial Pacific via strengthened trade winds, and then the sequestered heat flowed westward into upper ocean in the Indian Ocean through strengthened Indonesian Throughflow (ITF) [England *et al.*, 2014; Kosaka and Xie, 2013; Lee *et al.*, 2015; Liu *et al.*, 2016; Nieves *et al.*, 2015; Trenberth and Fasullo, 2013]. It is essential to track the dispersion of the anomalous heat arriving in the Indian Ocean to form a comprehensive picture of the global ocean energy redistribution. The anomalous heat may continue flowing westward into the Atlantic Ocean via the South Equatorial Current (SEC) and Agulhas Current as noted by Lee *et al.* [2015]. However, here we examine an alternate pathway: a southward heat transport in conjunction with a weakened SEC, diverting the canonical westward transport. This additional transport pathway causes an increase in heat content in the South Indian Ocean mid-latitudes (30 °S), contributes to the Southern Ocean warming, and

intensifies hemispheric asymmetry of oceanic heat content. The heat increase has important climate impacts such as changes to rainfall over the western coast of Australia and increased coral bleaching. The new path discovered here may be an essential route linking the tropical Indo-Pacific Ocean and the Southern Ocean in 2003-2012.

5.2 Introduction

The global surface warming hiatus, originally thought to indicate a reduction in global warming, is a widely-used phrase used to describe the deceleration in global surface warming. This nomenclature is continued herein to be consistent with previous studies. The surface warming hiatus does not represent a slowdown in the overall warming of the climate system, but rather is an energy redistribution within the global oceans[Yan *et al.*, 2016]. As a possible response to the surface warming hiatus, the recent global ocean heat trend pattern features an intensifying hemispheric asymmetry with 67-98% heat gain occurred in the southern hemisphere extra-tropical ocean and a large accumulation of heat in the mid-latitudes of Southeast Indian Ocean (SEIO)[Levitus *et al.*, 2012; Roemmich *et al.*, 2015; Wijffels *et al.*, 2016]. The Indian Ocean accounts for most of the global ocean (0-700m) heat gain via the strengthened ITF[Lee *et al.*, 2015] and may have an important contribution in the hemispheric asymmetry during the global energy redistribution. The detailed heat variability in the Indian Ocean and related relationship with the anomalous heat inflowing from strengthened ITF is still not well-understood. In this study, we investigate the OHC₇₀₀ variability in the Indian Ocean and use a climate model (CESM) to simulate and interpret the physical processes involved.

5.3 Results

In the SEIO, the OHC_{700} trend in 2003-2012 shows a prominent meridional band with a positive trend in heat content and the Southwest Indian Ocean (SWIO) has a negatively trended zonal band (5.1a; Levitus data). In order to study the transfer of heat in the Indian Ocean, we divide the Indian Ocean into the North Indian Ocean (NIO, 0-25°N, 25-125°E), the SEIO (0-34°S, 25-76°E), and the SWIO (0-34°S, 76-125°E), which are defined in FigureC 1 in Appendix C. The positively trended meridional OHC band extends along the western coast of Australia from the equator to 40°S and the negatively trended zonal OHC band spreads along 15°S from Africa to the middle of the Indian Ocean (85°E). When the heat flows from the Pacific into the Indian Ocean, we expect a heat transport along the equator from the ITF region to the eastern coast of Africa [Lee *et al.*, 2015], but, instead, there is an anomalous heat closer to Western Australia, implying some southward transport mechanism within the Indian Ocean. The OHC trends in the NIO are weak, but show an energy balance with a cooling Arabian Sea and a warming Bay of Bengal. The heat content trend is not sensitive to the end points of the trend analysis indicating the observed trend in the Indian Ocean is a decadal timescale change (FigureC 2 in Appendix C). The heat trend patterns simulated by the CESM (5.1b) are in good agreement with Levitus data [Levitus *et al.*, 2012], suggesting the CESM has captured major physical processes that determine the OHC_{700} variability. The time series of the OHC_{700} changes in the SEIO reveal an unusual rise in 2000s, in contrast to OHC_{700} in the NIO and SWIO (FigureC 3 in Appendix C). The time series of the OHC_{700} in the SEIO grew persistently in the last decade, i.e., 2003-2012, while the OHC_{700} in the SWIO demonstrated a negative trend from 2008 to 2012. The spatial pattern and time series results imply most of the

anomalous heat carried by the ITF is accumulated in the SEIO and spreads southward off the western coast of Australia toward the Southern Ocean.

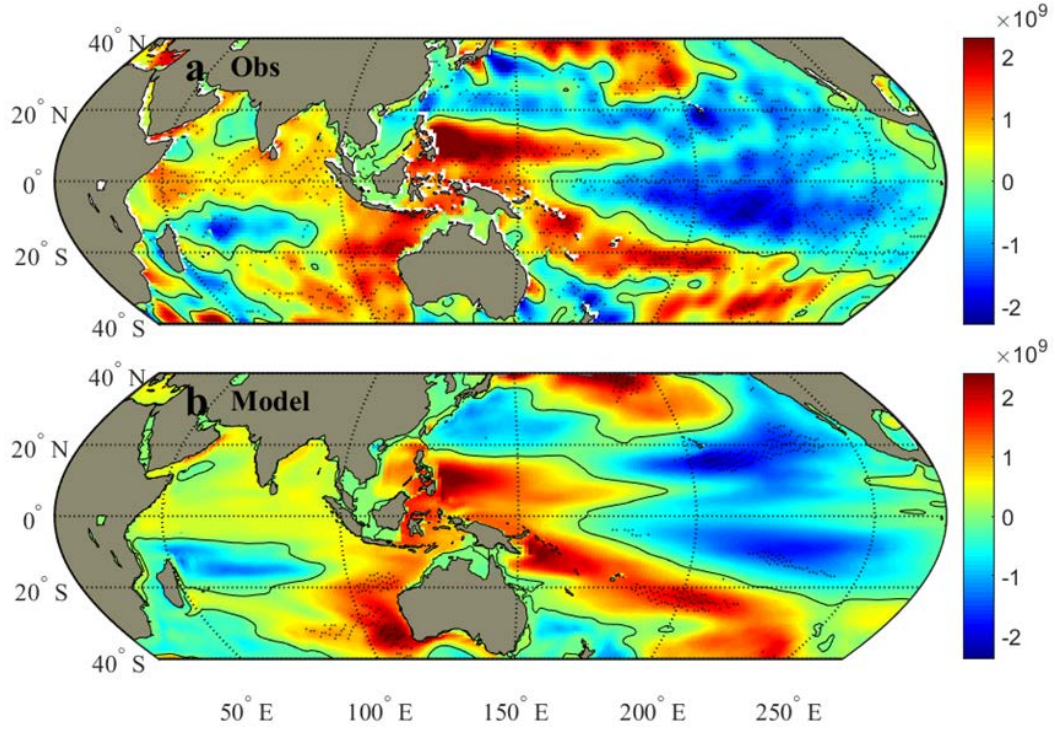


Figure 5.1. The OHC_{700} linear trends (units: J/m²/Decade) during the warming hiatus (2003-2012) in the upper ocean (0-700m). The black contour line is shown for 0. a, the observation data from Levitus et al., 2012. b, the model result from the CESM model. Black points in the shading color indicate the trends in those locations are significant in statistics ($P < 0.05$).

The mechanisms for the anomalous heat gain in the SEIO can be evaluated by comparing the heat budget terms in 2003-2012 with period computed as anomalies relative to the reference period (1990-2002). The heat budget includes four terms: storage rate, horizontal heat advection, vertical heat advection, and net surface heat flux. The heat budget anomalies (5.2a) indicate the simulated increase of OHC_{700} was

majorly dominated by an intensified horizontal heat advection. This horizontal advection anomaly can be as high as 0.005PW in 2003-2012 and is damped slightly by both the vertical heat advection and net surface heat flux. The overall heat budget is in the supplementary FigureC 4 (Appendix C). Further heat advection analysis (5.2b) indicates that the growing horizontal heat advection is due to an increased heat inflow through the eastern (0.098 ± 0.056 PW/decade, positive trend means heat inflow in the box) and the western (0.13 ± 0.057 PW/decade) boundaries in the SEIO. The sections for the heat transport are plotted as white box in the 5.3a. The northern (-0.17 ± 0.05 PW/decade, negative trend means heat outflow in the box) and southern (-0.032 ± 0.029 PW/decade) borders show a rising heat outflow, but cannot compensate for the inflow in the other two borders.

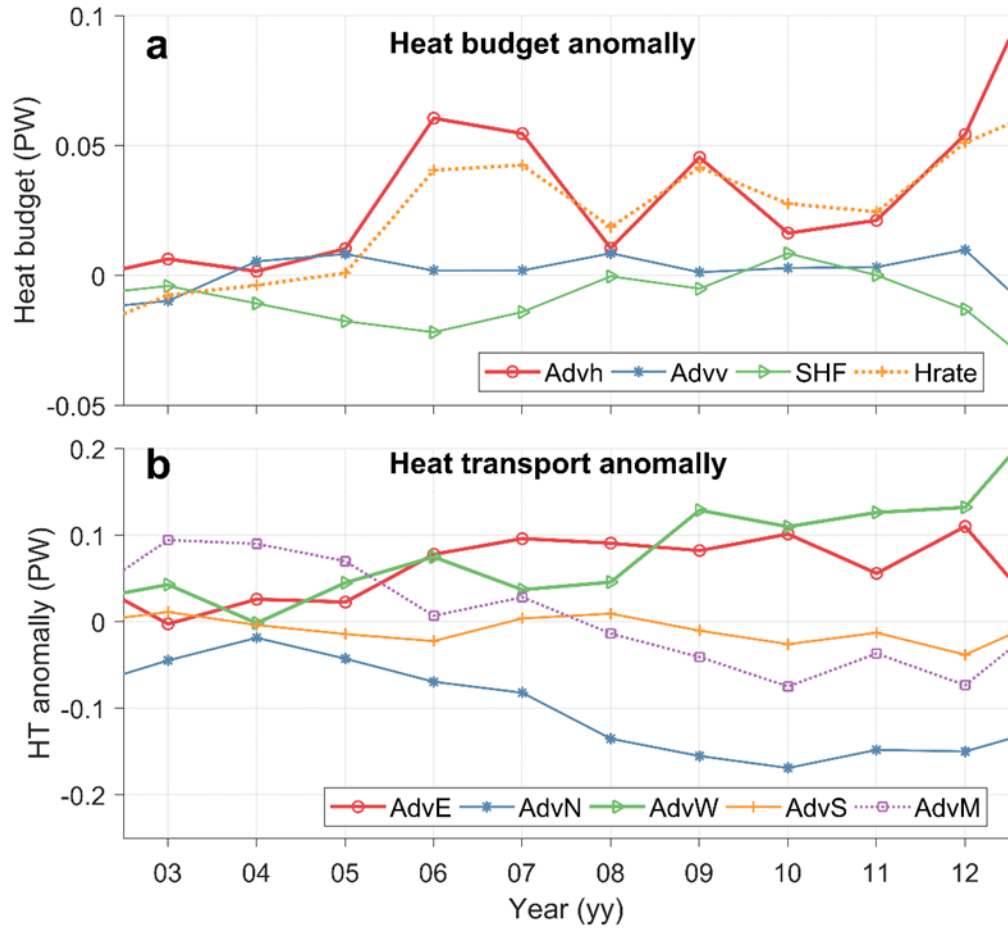


Figure 5.2. The time series of heat budget anomalies (a) and heat transport anomalies (b) in the SEIO.

a, storage rate (Hrate), horizontal heat advections (Advh), vertical heat advections (Advv), and net surface heat flux (SHF) for the SEIO. b, the heat transport in the east (AdvE), north (AdvN), west (AdvW), south (AdvS) boundary of the SEIO and the AdvM is the heat transport cross the middle latitude in the SEIO. Anomalies relate to the reference period in 1990-2002. Heat increase (positive trend) is inflowing to the box and heat decrease (negative trend) is outflowing of the box. The section locations are plotted as white box in the 5.3a. PW is 10^{15} J/s.

The transports along the eastern and western borders are related to the ITF and SEC system respectively. The intensified heat inflow in the eastern border evidently

relates to the recently strengthened ITF agreed with observations[*Lee et al.*, 2015; *Liu et al.*, 2015; *Susanto et al.*, 2012]. The significant heat growing in the western boundary coincides with a reduced volume transport of SEC (Supplementary FigureC 3 in Appendix C). This suggests a reduced heat outflow by a weakened SEC. The corresponding flow pattern will be analyzed in the following. A middle section (advM in 5.2b) between tropical and subtropical Indian Ocean (section location is in the 5.3a) is selected to illustrate the strength of southward transport from the tropic to the mid-latitudes in the SEIO. Note that the middle section (advM in 5.2b) indicates a strengthened southward heat flow with a rate of -0.2 ± 0.046 PW/decade (negative trend means heat outflow from the tropic to the mid-latitudes), which is stronger than the northern border (-0.17 ± 0.05 PW/decade) and the southern border (-0.032 ± 0.029 PW/decade). This suggests a large portion of the heat inflow from the eastern border moves southward and remains in the mid-latitudes. The heat transport variations are consistent with the water volume transport changes (Supplementary FigureC 3 in Appendix C). The strengthened ITF moves plenty of anomalous heat into the SEIO, but a weakened outflow cannot deliver these anomalous heats out through the western side. Most of the anomalous heat shifts southward and remains in the mid-latitudes for a strengthened southward flow in the middle section and a slightly varied volume transport along the southern border.

Given the important role played by the ITF and SEC systems, closer examinations of the depth-averaged velocity (0-700m) and sea level are warranted (5.3a). Eastward and westward current trends emerge in the western and eastern tropical Indian Ocean, respectively. These two anomalous flows converge around 90°E , then most of the anomalous flows turn southward and part of them turn

northward to the NIO. The location of anomalous eastward current coincides with the SEC which implies the SEC is weakened in the western part. Stemmed from the southward turning, a strengthened southward flow is witnessed from the equator to the 34°S in the SEIO, which is consistent with a recent increase in the strength of poleward Leeuwin Current[Feng *et al.*, 2015].

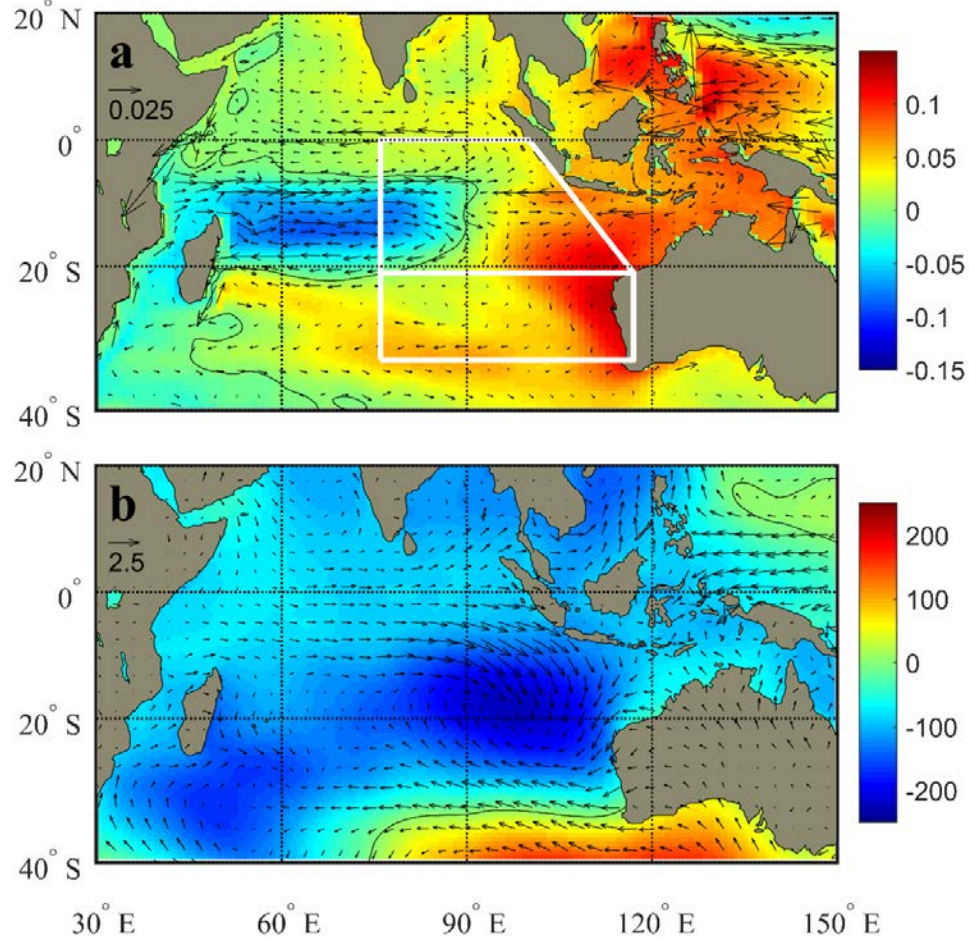


Figure 5.3. a, The linear trends for the depth-averaged (0-700m) velocity (a, vector, m/s/decade), sea level (a, shading, m/decade), 10 m-wind (b, vector), and sea level pressure (b, shading) during the warming hiatus (2003-2012). The velocity and sea level trends in a are from CESM model result and units are m/s/decade and m/decade respectively. The white box is the line for the computation of volume transport and heat transport in the 5.2. The wind and sea level pressure trends in b are from JRA55-GO atmospheric

reanalysis data that drive the CESM. The black contour line is shown for 0.

Corresponding to the flow trend, the sea level trend reveals zonal-negative and meridional-positive bands in the equatorial region in the SWIO and along the western coast of Australia in the SEIO respectively (5.3a) which match the OHC₇₀₀ pattern. The zonal-negative sea level trend results in a southward sea level gradient, covering a region from the eastern coast of Africa to the 90°E and from -5°S to -17°S. This suggests that the negative sea level trend may weaken the western part of SEC through a latitudinal sea level gradient in the SWIO. The meridional-positive sea level trend leads to a westward sea level gradient, spanning from the equator to the 40°S along the coast of Western Australia. The westward sea level gradient may strengthen a southward flow that connects the ITF and the mid-latitudes of South Indian Ocean. The velocity variabilities mostly emerge in the surface layer (supplementary FigureC 5 in Appendix C) which reaffirms that the weakened SEC and strengthened southward current are associated with the surface sea level variations.

The sea level variability at decadal time scale in the Indian Ocean is largely explained by the wind variation in the tropical and sub-tropical Indo-Pacific in terms of the Ekman pumping and the planetary waves[Feng *et al.*, 2010; Nidheesh *et al.*, 2013; Timmermann *et al.*, 2010]. The sea level and wind trends in the 5.3 illustrate similar responses with previous studies[Han *et al.*, 2010; Lee and McPhaden, 2008; Nidheesh *et al.*, 2013]. The location (5°S-20°S) of zonal drop of sea level is consistent with Southern Hemisphere Ekman mass transport to the left of the surface westerly wind trend (Fig. 3b). In addition, the time evolution of this zonal drop of sea level (Supplementary FigureC 6) shows a westward propagation of sea level drop

suggesting a contribution of oceanic Rossby waves to the development and propagation of the drop. The Rossby waves are driven by the wind stress curl through Ekman pumping velocity (Supplementary FigureC 7) which is specified in *Lee and McPhaden* [2008] and *Susanto et al.* [2012]. The sea level variations in the SWIO (5°S-20°S) and tropical Pacific are negatively correlated with each other as are trade wind variation across two basins. The sea level variations in the SWIO (5°S-20°S) and tropical Pacific are negatively correlated with each other as are trade wind variation across two basins. The negative correlation may be associated with a change of deep convection associated with the Walker circulation over the tropical Pacific and Indian Oceans[*Han et al.*, 2010], which is consistent with the recent La Niña-like state in the Pacific.

In the SEIO, the positive sea level trend is related to the northerly wind anomalies off the Western Australia coast (5.3b) and a transmission of the high sea level signals from the Pacific via the equatorial and coastal waveguides. The northerly wind anomalies as part of the cyclonic wind anomalies is caused by a negative sea level pressure anomalies (5.3b) due to a Gill response to the La Niña[*Feng et al.*, 2013] and the sea level anomalies result from recently strengthening of trades winds and associated wind stress curls in the Pacific[*England et al.*, 2014; *Nidheesh et al.*, 2013]. The transmission is accompanied with an increased ITF transport as indicated in the *Feng et al.* [2010]. As a result, the weakened SEC is related to the westerly wind anomalies while the strengthened southward current is linked to the cyclonic wind anomalies and sea level anomalies transmission from Pacific as a response to the La Niña like-state in the Pacific.

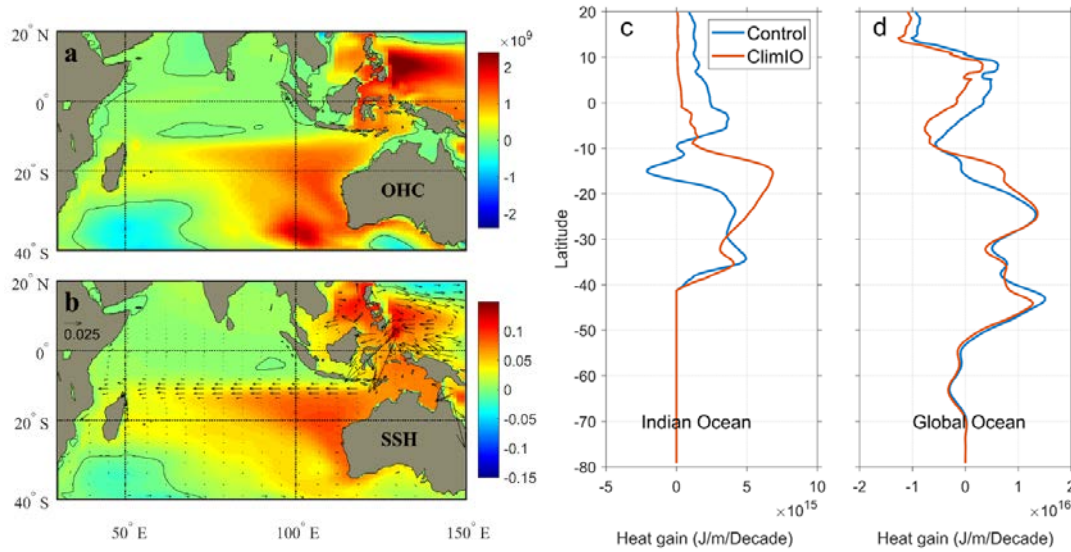


Figure 5.4. a, The The OHC_{700} linear trends (units: $J/m^2/Decade$) for the IOclim case during the warming hiatus (2003-2012) that the Indian Ocean is driven by climatological wind. b, depth-averaged velocity (0-700m, m/s/decade) and sea level (m/decade) linear trends for the IOclim case (2003-2012). c and d are zonally integrated heat gain versus latitude in the Indian Ocean and global ocean respectively. The results are from the IOclim case. The black contour lines in a and b are shown for 0.

The conjunction between flow and wind trend patterns indicates the wind variability is responsible for the flow variation and heat redistribution in 2003-2012. In order to further verify the role of wind and sea level variability, we conduct a numerical experiment, named as IOclim that the Indian Ocean is driven by the climatological wind while the other ocean basins are still driven by the realistic wind. In contrast to the control case (all realistic wind), the OHC_{700} in the IOclim (5.4a) is characterized by a zonal band instead of a meridional band. The anomalous OHC_{700} extends from the eastern tropical Indian Ocean to the western tropical Indian Ocean along the path of the SEC. The sea level trend (5.4b) has a similar westward expansion

with the OHC₇₀₀. An intensified westward flow emerges in the tropical region connecting the eastern and western tropical Indian Ocean. This is an expected heat pathway via the strengthened SEC and Agulhas current. Note that the IOclim case still has a warming trend in the mid-latitudes (western coast of Australia), but a less rate than that of the control case. The reason is the existence of high sea level transmission from the Pacific, even though wind anomalies vanish in the IOclim case. The test result confirms the above conclusion that the weakened westward heat transport and strengthened southward heat transport observed are largely dominated by the wind and sea level variability during the hiatus period. The zonally integrated heat gain in the Indian Ocean (5.4c) indicates most of the anomalous heat is accumulated in the tropical Indian Ocean (15°S) in the IOclim case instead of mid-latitudes (32°S) in the control case. For the global energy redistribution, the heat gain versus latitude (5.4d) implies that the IOclim case tends to keep the heat into the Indian Ocean while the control case moves some of the anomalous heat southward to the Southern Ocean. As a consequence, the Southern Ocean experiences a slight higher warming rate in the control case ($1.57 \times 10^{22} \pm 1.20 \times 10^{21}$ J/decade) than that in the IOclim case ($1.50 \times 10^{22} \pm 1.02 \times 10^{21}$ J/decade, supplementary FigureC 8 in Appendix C). Although relatively small variability with a rate of 5% in the experiment, this may partly contribute a recently rapid warming of the Southern Ocean which is reported by many studies[Riser *et al.*, 2016; Sutton and Roemmich, 2011]. This reaffirms that the strengthened-southward heat transport not only builds up heat in the middle latitudes of South Indian Ocean but also contributes to the warming Southern Ocean and global energy redistribution within the oceans.

5.4 Discussion and Conclusion

The consistent changes of observed/simulated sea level and atmospheric circulation show that the strengthened southward heat transport in the Indian Ocean is a consequence of the La Niña state in the Pacific during the warming hiatus. The mechanism and significance are summarized in 5.5. Under the influence of the La Niña-like state in the Pacific, according to Gill response, the westerly wind anomalies are induced in the SWIO and cyclonic wind anomalies and a high sea level anomaly are formed in the SEIO during the surface warming hiatus. The westerly wind anomalies weaken the SEC while the cyclonic wind anomalies and a high sea level anomaly strengthens a southward flow. As a result, the anomalous heat from the Pacific moves southward through the strengthened southward flow instead of flowing westward by the weakened SEC.

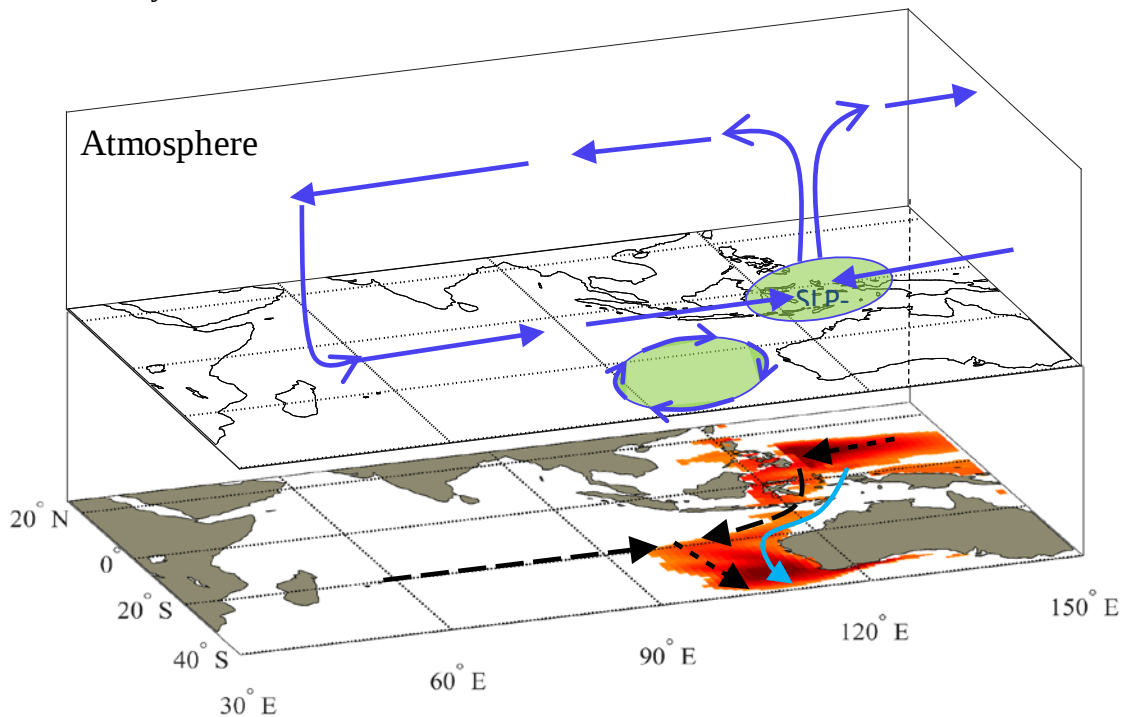


Figure 5.5. The schematic graph of the trends in OHC700 and ocean-atmosphere circulation in the Indo-Pacific over 2003-2012.

The color shading represent the OHC700 trend (red represents warming). In the atmosphere part, the blue solid vector are the wind anomaly under the influence of La Niña-like state in the Pacific Ocean. The “SLP-“ means the negative sea level pressure trend. In the ocean part, the dashed black vectors are the flow trend and the green solid vector is the transmission of high sea level from the Pacific to the Indian Ocean.

The heat path detailed here may be an important path linking the tropical Indo-Pacific and the Southern Ocean during the La Niña. Compared with the traditional path (SEC and Agulhas Current), this path is short indicating a fast poleward heat transport. The response between the tropical Indo-Pacific and Southern Ocean through oceanic-teleconnection may be quicker than we thought before. The strengthened southward heat transport has important climate impacts such as changed rainfall over the western coast of Australia and increased coral bleaching[Feng *et al.*, 2013; Zinke *et al.*, 2015]. The anomalous heat not only moves southward to the Southern Ocean, but also flows eastward along the southern coast of Australia through with the Leeuwin Current (5.1). A rapid surface warming was observed in this coastal region during the hiatus period[Liao *et al.*, 2015] suggesting a rising probability of occurrence of hot event in the future. Therefore, monitoring and predicting the continuing changes of the strengthened southward heat transport in the Indian Ocean and links to other regions, remains a key priority.

Chapter 6

CONCLUSIONS

Starting with a cold event, I have studied the dynamics of two off-shore flows and then the global coastal variabilities and heat redistribution in the Indian Ocean. Some of the main results are: (i) a detailed mechanism and wind criterion of the off-shore flow in the southern strait; (ii) a thorough reason for the cold event of 2008; (iii) the global-scale coastlines show significant SST variabilities in two periods (1982-1998 and 1999-2013); (iv) the southward spread of anomalous heat in the Indian Ocean indicate a new route linking the equatorial region and Southern Ocean.

6.1 The Off-Shore Flow in the Southern Strait

The cold and fresh coastal current leaves the 30m isobath at 23.5°N, crosses the West Waterway, and spreads from the coast to the Central Bank. The chlorophyll shows high concentration in the same region, in agreement with the temperature and salinity distributions. The separated, nutrient-rich coastal water is important to the marine ecosystem and fishery. Model simulation and analyses are conducted to understand the deflection process and associated mechanism.

Three circulation regimes are derived depending on the balance of wind stress and pressure gradient associated with the poleward tilt of sea level along the Taiwan Strait. The flow is controlled by a southward (northward) current when the wind stress is stronger (weaker) than an upper (lower) bound. The two opposite flows occur

simultaneously and converge, and the coastal current deflects when the wind stress is moderate.

There is a striking along-shore variation around 23.5°N along the 30m isobath. The southward coastal current is weakened in the TWB and a reversed (northward) current is generated over this bank. Associated with this two-flow concurrence, an off-shore transport is strengthened and large temperature and salinity gradients are formed over the bank.

Vorticity analysis and numerical experiments indicate that the off-shore transport is associated with the topography of the TWB and two opposite external forces: wind and sea-level gradient. Over the TWB, the northeasterly monsoon drives a southward coastal current along the coastline and the open ocean forcing induces a northward warm current (SCSWC) on the shelf. These two currents meet at the western bank with the former limited in the inner shelf and latter located outer, generating a strong relative vorticity. The high relative vorticity trigger a negative bottom stress curl which is reinforced in the bank by the gentle slope over the Bank. This reinforced negative bottom stress curl dominates the other terms in Eq. (2.7) and leads to an amplified off-shore flow over the bank.

6.2 The Role of Coastal-Kelvin Waves on the 2008 Cold Disaster in the Taiwan Strait

In winter, the CKWs, as the third factor, constantly propagating toward the TWS from north, can impact flow patterns in the TWS. The arrival of a CKW trough (crest) in the western coast can strengthen a northward (southward) geostrophic current through changing the cross-strait sea level gradient. The strengthened northward (southward) current can amplify a northward (southward) volume transport.

During the first stage, the northeasterly wind was moderate resulting in a cross-strait flow pattern. Around February 4, the arrival of a CKW trough induced an additional northward flow. When the northward flow turned around the Zhangyun Ridge, a cyclone was formed in the north of Zhangyun Ridge. This cyclone led to an additional cross flow, which intensified the eastward (offshore) movement of cold water. In the second stage, the northeasterly wind was very strong leading to a southward flow pattern. The reach of a CKW crest generated an additional southward current which increased the southward movement of cold water. The additional eastward and southward flow caused by the CKWs played an important role in transporting the cold water toward PHI killing the tropical fishes there.

The impact of volume transport variations caused by CKWs in the TWS on other ocean phenomena is also worthy to investigate. For example, the volume transport in the TWS is southward (northward) in winter (summer) for northeasterly (southwesterly) monsoon. *Moon and Hirose* [2014] showed the seasonal transport variation in the TWS can impact the Kuroshio water intrusion in the East China Sea. Therefore, it is worth to study that if the short-term transports variation caused by CKWs can impact the Kuroshio water intrusion in the East China Sea.

6.3 The Coastal Ocean Response to the Global Warming Acceleration and Hiatus

Lima and Wetthey [2012] reported that the last three decades (1982-2010) witnessed a high warming rate in 71.6% of the global coastlines. However, with the same data set, the coastal SST trends illustrate exceptional trend reversals when the last three decades are separated into the warming (1982-1997) and hiatus (1998-2013) periods. In the warming period, most of the global coastlines (68.19%) showed a

warming trend consistent with global warming. However, during the hiatus period, the warming trend reversed in many coastlines of the low and mid latitudes (31.4% of the global coastlines). The distribution of these trend reversals matches with multiple reports that the global surface warming hiatus is mainly concentrated in the low and mid latitudes, especially in the Pacific Ocean[*England et al.*, 2014; *Kosaka and Xie*, 2013].

Additionally, 17.9% of the global coastlines changed from cooling to warming in the hiatus period. These trend reversals, distributed in the low and mid latitudes, may be attributed to natural climate variability instead of anthropogenic-induced warming. Although SST trend reversals are apparent for many coastlines (49.3%, 31.4%+17.9%), a continuous increase of SST could be detected for a considerable portion of the global coastlines (46.7%). Notably, the high northern latitudes demonstrated an accelerated warming even though in the hiatus period. The continued warming was concentrated in the high northern latitudes, while the effect of global climate hiatus was focused in the low and mid latitudes.

The corresponding EHDs and ECDs trends have similar variations with the SST trend spatially and temporally. Spatially, the increase of EHDs was observed in the warming coast (e.g., the eastern portion of Madagascar Island, western and southern part of Australia, Caribbean Sea, Mediterranean Sea, eastern part of North America (north of Cape Hatteras), and much of the Arctic Ocean) in the hiatus period (4.3). The number of ECDs increased in the cooling coast (the Eastern Pacific, China and Japan, northern part of Australia, Red sea, Persian Gulf, South Africa, eastern part of North America (north of Cape Hatteras), and some coastlines in both polar regions) in the hiatus period (4.4). Temporally, the trend of global mean EHDs (ECDs) was ~-

0.74 (-10.5 ± 8.89) days/decade in the warming period and increased to ~ 3.96 (~ 1.65) days/decade in the hiatus period. In contrast to the warming period, the sum of EHDs and ECDs increased globally in the hiatus period (Figures. 4.3 and 4.4). This means the probability of extreme events (hot or cold) might increase if the global surface warming hiatus continues. The increased sum of EHDs and ECDs may be related to a rising coastal area of large positive and negative SST trends from the warming period to the hiatus period (see FigureB 2 in Appendix B).

6.4 A Strengthened Southward Heat Transport in the Indian Ocean during the Global Surface Warming Hiatus.

The consistent changes of observed/simulated sea level and atmospheric circulation show that the strengthened southward heat transport in the Indian Ocean is a consequence of the La Niña state in the Pacific during the warming hiatus. The mechanism and significance are summarized in 5.5. Under the influence of the La Niña-like state in the Pacific, according to Gill response, the westerly wind anomalies are induced in the SWIO and cyclonic wind anomalies and a high sea level anomaly are formed in the SEIO during the surface warming hiatus. The westerly wind anomalies weaken the SEC while the cyclonic wind anomalies and a high sea level anomaly strengthens a southward flow. As a result, the anomalous heat from the Pacific moves southward through the strengthened southward flow instead of flowing westward by the weakened SEC.

The heat path detailed here may be an important path linking the tropical Indo-Pacific and the Southern Ocean during the La Niña. Compared with the traditional path (SEC and Agulhas Current), this path is short indicating a fast poleward heat transport. The response between the tropical Indo-Pacific and Southern Ocean through

oceanic-teleconnection may be quicker than we thought before. The strengthened southward heat transport has important climate impacts such as changed rainfall over the western coast of Australia and increased coral bleaching[Feng *et al.*, 2013; Zinke *et al.*, 2015]. The anomalous heat not only moves southward to the Southern Ocean, but also flows eastward along the southern coast of Australia through with the Leeuwin Current (5.1). A rapid surface warming was observed in this coastal region during the hiatus period[Liao *et al.*, 2015] suggesting a rising probability of occurrence of hot event in the future. Therefore, monitoring and predicting the continuing changes of the strengthened southward heat transport in the Indian Ocean and links to other regions, remains a key priority.

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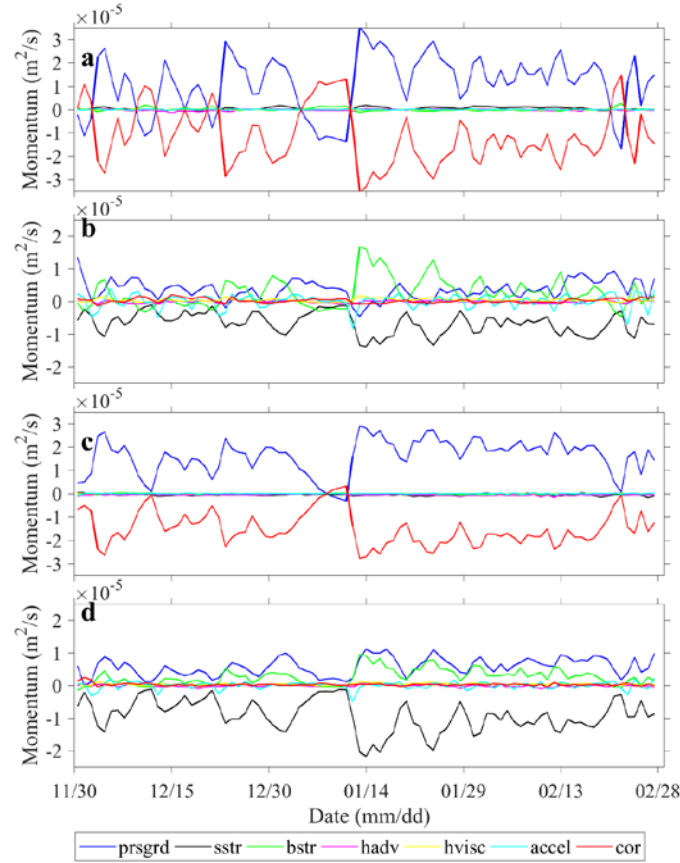
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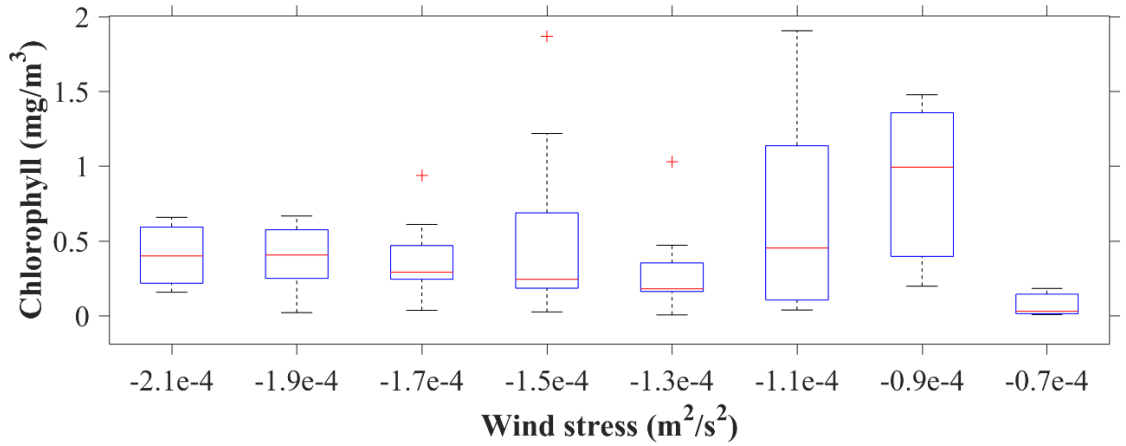
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Appendix A

SUPPLEMENTARY MATERIAL FOR CHAPTER 2



FigureA 1. The terms of depth-averaged momentum equation in the cross-shore (a and c) and along-shore (b and d) directions. a and b are regional-mean values in the region Rtwb where is the green rectangle in the 2.6b, c and d are regional-mean values in the region Rntwb where is the red rectangle in the 2.6b. This figure indicates the Coriolis term is negligible in the along-shore momentum equation (Eq. 2.1) in both regions where is close to the coastline. prsgrd, sstr, bstr, hadv, hvisc, accel, and cor represent pressure gradient, surface wind stress, bottom stress, horizontal advection, horizontal viscosity, acceleration, and Coriolis term respectively.



FigureA 2. The distribution of regional-mean chlorophyll in different ranges of wind stress. The region is defined as the black rectangle in 2.2a. The each range of wind stress is $0.2 \times 10^{-4} m^2/s^2$ and the tick label is the center of each range. The lower and upper edges of each blue box are the first quartile and third quartile respectively. The red line in the blue box is the median value. The whiskers (black line outside of blue box) is the smallest and largest non-extreme values in the data. The red plus symbol is the extreme value which is the value lying farther from the box edge than 1.5 times the interquartile (blue box). The chlorophyll is from MODIS data from 2002 to 2015. This figure indicates the chlorophyll concentration tends to be the largest in the wind stress range between 0.9×10^{-4} and $1.36 \times 10^{-4} m^2/s^2$.

The Relation Between Bottom Stress Curl and Along-shore Pressure Gradient

Here we derive a vorticity equation of integrated momentum equation to show the relation between relative vorticity and along-shore pressure gradient. For a steady flow, assuming $D \approx H$, the equation is expressed as:

$$\bar{\bar{u}} \cdot \nabla \bar{\xi} = J(P^b, H) + \nabla \times \bar{\tau}^s - \nabla \times \bar{\tau}^b \quad \text{A1.1}$$

where the overbar denotes a vertically integrated quantity. $\bar{\xi} = \nabla \times \bar{\bar{u}}$, $\bar{\bar{u}} = \int_{-H}^{\eta} \bar{u} dz$, and P^b is bottom pressure. If we rewrite eq. A1.1 along an isobath according to *Gan et al.* [2012],

$$-\frac{1}{\rho} P_p^b = \frac{1}{H_n} \left[\bar{\bar{u}} \cdot \nabla \bar{\xi} - \nabla \times \bar{\tau}^s + \nabla \times \bar{\tau}^b \right] \quad \text{A1.2}$$

where P_p^b is the along-isobath gradient of bottom pressure. P_p^b is equivalent to the depth averaged along-isobath pressure gradient force when stratification is weak (Gan et al. 2012). The vorticity dynamical processes (i.e., nonlinear, wind stress and bottom stress) can cause an along-isobath pressure gradient. The third term in eq. S1.2 can be split into two terms which are

$$\frac{1}{H_n} \nabla \times \bar{\tau}^b = \frac{1}{H_n} \left[C_d \left| \bar{u}'_b \right| \nabla \times \bar{u}'_b + C_d \nabla \left| \bar{u}'_b \right| \times \bar{u}'_b \right] \quad \text{A1.3}$$

This equation states that the along-isobath pressure gradient force can be induced by two components of bottom stress curl. If we let this equation (eq. A1.3) divided by f , these two terms are exactly same expression with terms I and III in eq. 4.5. This can explain why terms I and III are geostrophic. Therefore, the cross flow can also be explained by an off-shore geostrophic current induced by a positive along-isobath pressure gradient which is related to vorticity dynamics.

Reference:

Gan, J., H. San Ho, and L. Liang, 2012: Dynamics of intensified downwelling circulation over a widened shelf in the northeastern South China Sea. *J. Phys. Oceanogr.*, 43, 8-94.

The Calculation of Vorticity Budget

The different terms in the vorticity equation (Eq. 2.7 and 2.9) is obtained using ∇ cross product the momentum terms in the diagnostic results of the ROMS. The vorticity locates at the cell corners (psi point in the ROMS). The budget residual is smaller than the time rate (TR) term indicating a negligible TR term. The calculation of different terms in Eq. 2.9 and 2.10 uses flow field in the ROMS average results. The estimated bottom stress curl by the flow field agrees with the results calculated by the momentum term

Appendix B

SUPPLEMENTARY MATERIAL FOR CHAPTER 4

Methods

Data. The SST data used in this study is NOAA's Optimum Interpolation (OI) $\frac{1}{4}$ Degree Daily SST (also known as Reynolds 0.25 v.2), which is obtained from NOAA's National Climatic Data Center (<http://www.ncdc.noaa.gov/oisst/data-access>). There are two SST analysis products developed using OI. The product with only AVHRR infrared satellite SST data is used for long-term time series. A large-scale adjustment of satellite biases with respect to the in situ data is included in the data. Many studies of decadal SST variability, based on OISST indicate, there is no evident decadal scale error [Wang *et al.*, 2012; Zhu *et al.*, 2015]. The OISST shows very small biases in comparison with coastal in-situ SST data [Lima and Wethey 2012; Seabra *et al.*, 2011]. The coastal pixel used for analysis is the pixel which is closest to land but with a land contamination of less than 50% [Lima and Wethey 2012].

The Yearly SST. The yearly SST, computed by taking annual mean of daily SST data, is first smoothed by a 10-year moving average, and then is transformed into standardized data ($z = \frac{x - \mu}{\sigma}$, z is the standardized data, x is the smoothed yearly SST data, μ is the mean value, and σ is the variance).

Trend Calculation. The SST trend in the study is calculated as the slope of the linear regression of seasonally detrended daily SST. In order to decrease errors in the SST OI estimates (e.g., random, sampling, and bias error), the slope is computed through a weighted least squares method, whose weights are inversely proportional to

the variance of the SST OI (i.e., $\propto 1/SD^2$). The degrees of freedom were adjusted using the Quenouille procedure to account for temporal autocorrelation [Santer *et al.*, 2008].

Confidence Interval. The confidence interval in the study is stated at the 95% confidence level. The standard error of the trend is computed by $s_b = \sqrt{\frac{s_e^2}{\sum_{t=1}^{n_t} t - \bar{t}}}$,

$s_e^2 = \frac{1}{n_e - 2} \sum_{t=1}^{n_t} e(t)^2$, $n_e = n_t \frac{1-r_1}{1+r_1}$, where s_b is the standard error, s_e^2 is the variance of the linear regression residuals, t is the time index, $\bar{t}$ is the average time index, $e(t)$ is the linear regression residual, n_t is the total number of time series, n_e is the adjusted number of time series after removing temporal autocorrelation, r_1 is the lag-1 temporal autocorrelation coefficient of $e(t)$ [Santer *et al.*, 2008]. In the study, the confidence interval is shown as 0.17 ± 0.11 . If the margin of error (“radius”) of a confidence interval is larger than the trend value (e.g., -0.13 ± 0.25), this means the result is not significant. However, the trend can still indicate a decreasing or increasing tendency, and the trend value is shown as ~ 0.13 .

Annual Frequency of Extremely Cold and Hot Days. In order to analyze the probability of variations of hot and cold events, the linear trends of annual frequency of extremely hot and cold days are given. The thresholds of extremely low and high temperature are defined as the 5th and 95th percentiles of the standardized anomalies of the raw SST (1982-2013) at each location separately. Then, the annual frequencies of daily anomalies exceeding the threshold values were computed. In the linear regression, the Quenouille procedure is adopted to adjust the degrees of freedom like SST trend calculation. The unit of trend of annual frequency of extremely hot and cold days is the number of days per decade $\pm$ trend confidence interval.

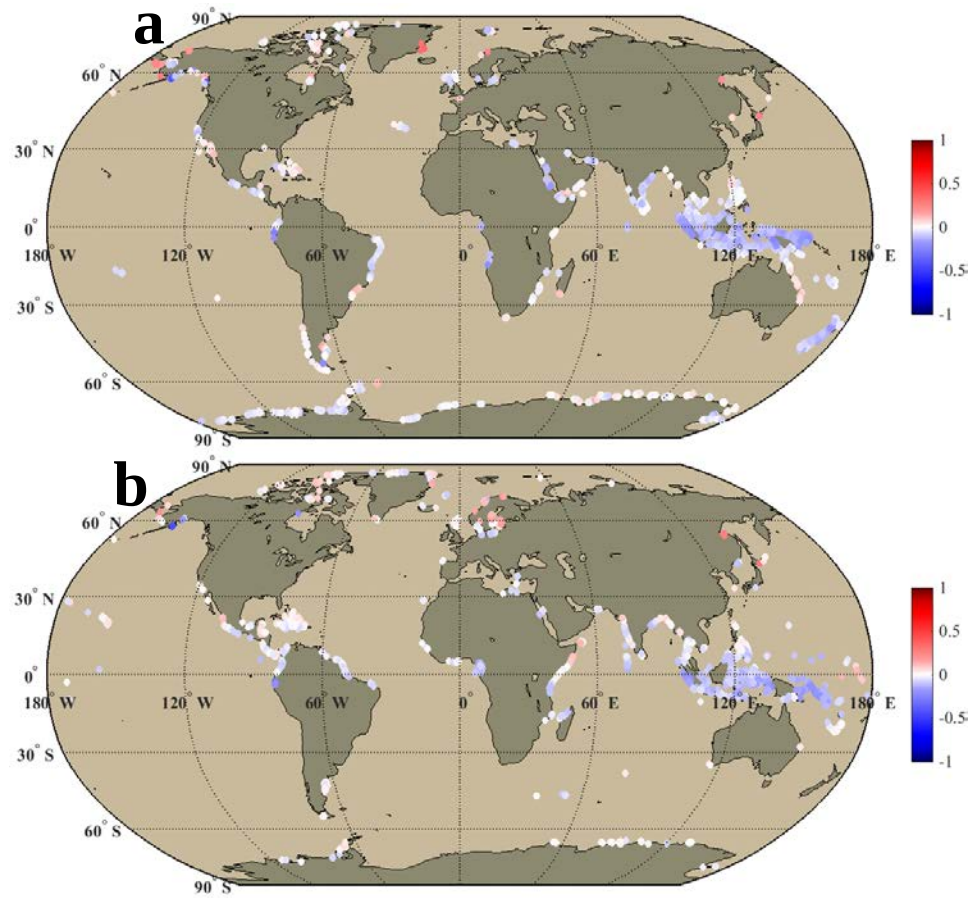
Mean Trend. The global (regional) mean warming/cooling rate is computed based on global (regional) mean data (area-weighted average). The area ratio in this study is also based on area-weighted average.

Standard Deviation. At each location, the standard deviation of SST in the warming (hiatus) period is calculated by $sd = \sqrt{\frac{1}{N-1} \sum_{i=1}^N |A_i - \mu|^2}$, where A is the time series of SST at each location, N is the number of variable A (SST), μ is the mean of A .

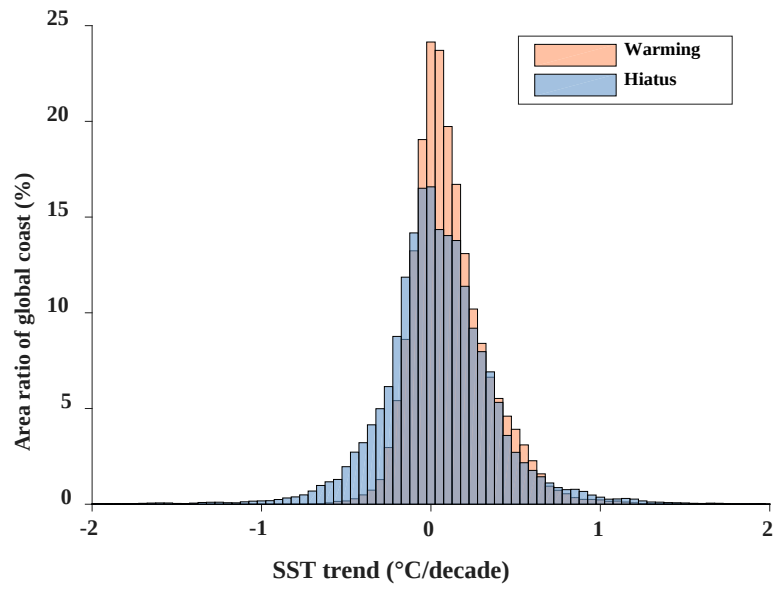
Low, Mid, and High latitudes. In this study, low latitude is defined as 30°S-30°N, mid latitude is 30°S-60°S and 30°N-60°N, high latitude is 60°S-90°S and 60°N-90°N.

References

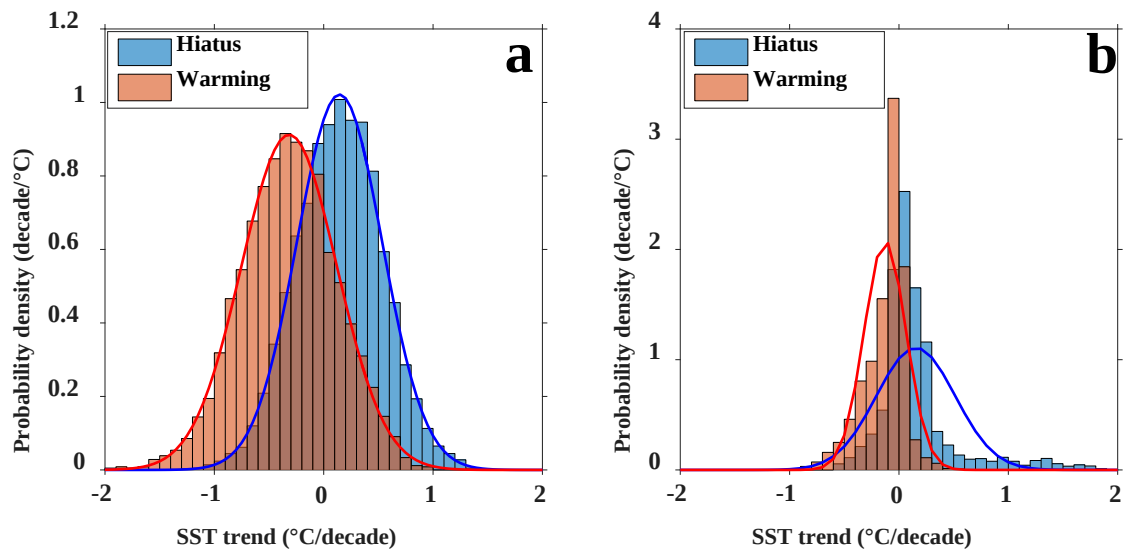
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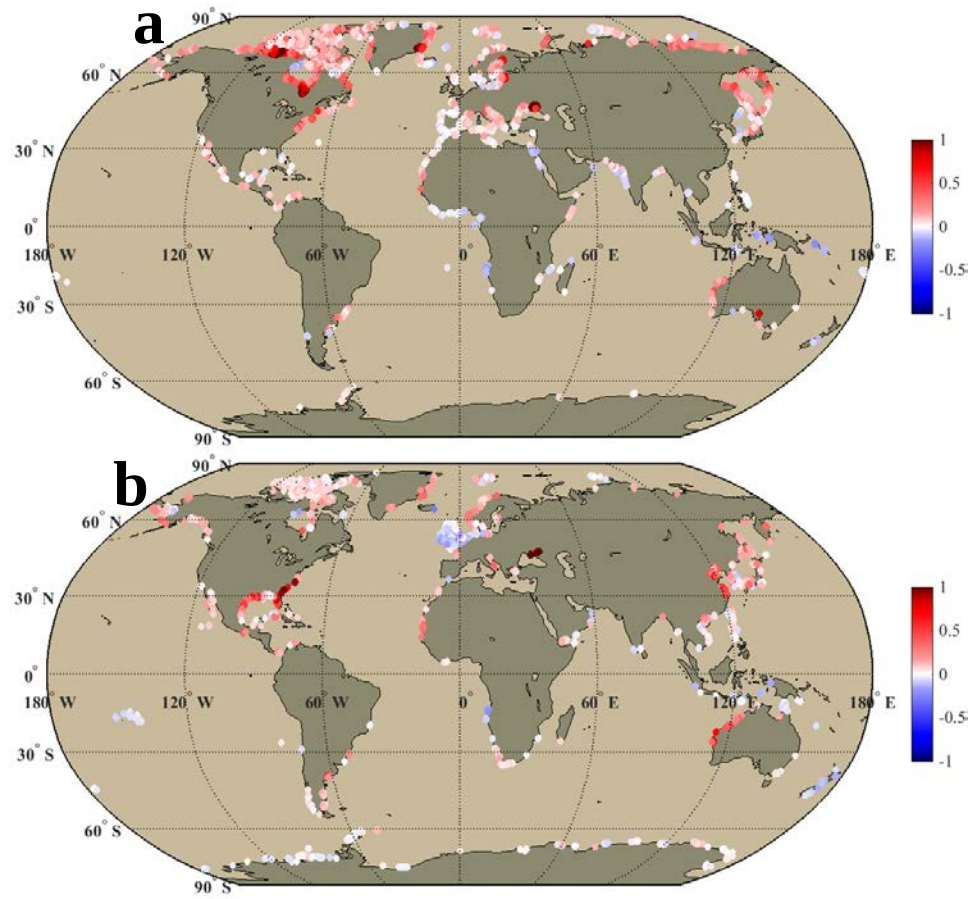
FigureB 1. The inconsistent points between SST trends and trends of extremely cold (hot) days. a, the SST decreased and the extremely cold days decreased; b, the SST increased and the extremely hot days decreased. The shading color is the standard deviation difference between warming period and hiatus period. The positive (negative) value means the standard deviation is larger (smaller) in the hiatus period than the warming period.



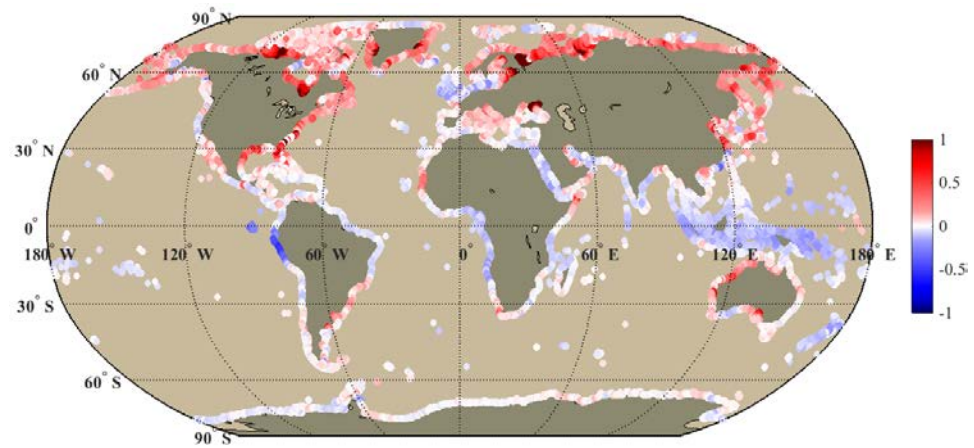
FigureB 2. The histogram of SST trends for percentage of global coastal area between the warming and hiatus periods.



FigureB 3. The probability density functions between the warming and hiatus periods in the Red Sea (a) and Arctic Ocean (b). The histogram is normalized with a uniform bin width (0.1 °C/decade). The line is calculated by probability density function with mean and standard value.



FigureB 4. Same as FigureB 1., the inconsistent points between SST trends and trends of extremely cold (hot), a, the SST increased and the extremely cold days increased; b, the SST decreased and the extremely hot days increased.



FigureB 5. The standard deviation difference between warming and hiatus periods.
The positive (negative) value means the standard deviation in the hiatus period is larger (smaller) than that in the warming period.

Table B1 The summary table for the SST trends, trends of extremely hot days (EHDs) and extremely cold days (ECDs) in the key regions

Region name	SST trend (°C/decade)		EHDs trend (days/decade)		ECDs trends (days/decade)	
	Warming	Hiatus	Warming	Hiatus	Warming	Hiatus
China and Japan coast	0.65±0.27	-0.69±0.44	8.33±7.13	-6.53±5.01	-36.57±23.29	25.30±11.57
Alaskan Peninsula	0.21±0.29	-0.49±0.46	2.64±15.26	-14.88±45.41	-14.71±10.02	21.13±19.74
Western South America	0.18±0.62	-0.36±0.36	4.15±37.33	-11.92±15.66	-11.16±6.81	19.27±11.98
Eastern North America (North of Cape Hatteras)	-0.13±0.25	0.54±0.24	-3.19±4.19	35.41±17.94	0.91±5.08	-3.04±8.74
Eastern North America (South of Cape Hatteras)	0.16±0.47	-0.37±0.33	-2.0±7.71	7.35±9.72	-11.42±14.1	27.2±24.7
Labrador Sea	0.19±0.19	0.38±0.22	3.29±2.66	26.33±16.26	-16.98±21.67	-2.89±5.36
North Sea	0.4±0.53	-0.17±0.46	10.66±4.58	1.45±27.75	-10.49±37.25	10.64±8.83
Caribbean Sea	0.12±0.14	0.24±0.17	-0.19±6.64	11.6±30.21	-17.08±12.52	-8.46±8.41
Mediterranean Sea	0.09±0.24	0.22±0.24	1.67±4.33	18.52±10.39	-6.82±11.44	-0.15±2.78
Argentina coast	0.01±0.36	0.19±0.35	-3.77±14.33	20.4±7.45	-8.10±15.35	1.82±13.12
Western South Africa	0.49±0.26	-0.49±0.25	8.67±12.33	-3.31±12.98	-31.15±41.82	29.6±14.6
Red Sea	0.29±0.14	-0.11±0.14	8.01±7.07	-9.05±13.36	-22.54±17.67	-3.14±5.03
Persian Gulf	0.40±0.35	-0.36±0.29	7.87±11.0	-27.56±20.07	-25.78±34.87	8.64±8.57
Madagascar Island	-0.09±0.16	0.33±0.16	-8.64±10.32	22.79±18.71	-4.74±25.65	-10.82±7.43
Western and southern part of Australia	0.07±0.20	0.33±0.21	-1.97±10.40	28.08±30.46	-12.9±12.9	-2.26±10.11
Arctic Ocean	0.12±0.18	0.38±0.30	2.45±2.34	24.79±5.43	-6.7±6.7	-3.16±7.66
Southern Ocean	0.01±0.03	-0.08±0.04	1.69±8.23	-9.55±11.74	0.34±5.31	10.12±5.78

Appendix C

SUPPLEMENTARY MATERIAL FOR CHAPTER 5

Methods

Data Sets. The Levitus ocean heat content (OHC) data set was employed in this study to estimate the trend of the OHC₇₀₀ in the Indian and Pacific Ocean.

Model Experiments. The National Center for Atmospheric Research (NCAR) climate model, the Community Earth System Model (CESM1.2) [Gent *et al.*, 2011] is a state-of-art global climate model which fully couples the atmosphere, ocean, land, land-ice, and sea-ice. The ocean-sea ice model [Danabasoglu *et al.*, 2011] in the CESM system was employed in this study to investigate the heat transport in the Indian Ocean. The ocean-sea ice grid is on a displaced pole grid (GX1V6) at a resolution of approximate 1° in the ocean with 60 vertical levels. Descriptions and evaluations of the CESM and ocean component (CESM_POP) refers to [Danabasoglu *et al.*, 2011]. Initialized from the Levitus initial condition, the model is forced by the JRA-55 based surface data set for driving ocean-sea ice models (JRA55-do v1.1). The dataset consists of nine atmospheric elements necessary for computing surface fluxes as well as fresh water runoff to the ocean. All of the atmospheric elements are derived from the forecast phase of JRA-55 and are three hourly. Temporal coverage is from 1958 to 2015.

The control case shown in the study is the ensemble mean of six member runs after spin-up. To spin up, a GIAF-configured case is run for 290 model years (1-296) when during the course of this 290-year period, the 58-year JRA55-do interannual

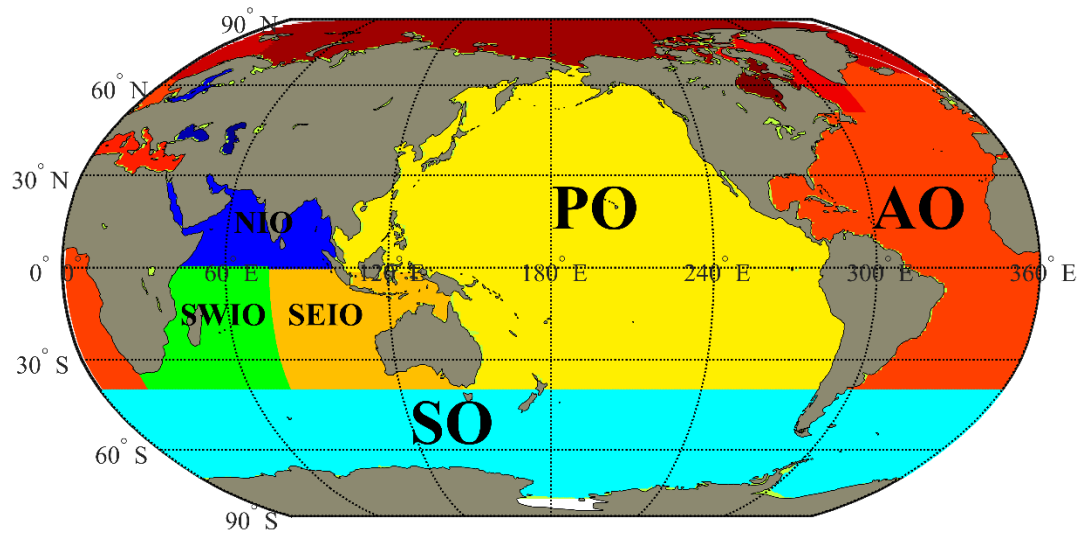
forcing data record (1958-2015) cycles five times. The simulated global OHC700 showed no sign of drift after about 2 cycles (120 years) indicating a quasi-equilibrium state in the spin-up run. Detailed set up refers to the user guide given by CESM development group (<http://www.cesm.ucar.edu/models/cesm1.2/pop2/doc/faq/>). After the spin-up run, the six member runs were initialized from the 233rd, 243rd, 253rd, 263rd, 273rd, and 283rd model year in the spin-up run. The ensemble mean of the six member runs is the results shown as the control case. The ClimIO case is identical to the control case except that the Indian Ocean is driven by a climatological wind which is defined as the mean wind in 1958-2015.

Trend Estimates. The trend in the study is calculated as the slope of the linear regression of seasonally detrended monthly data. The degrees of freedom were adjusted using the Quenouille procedure to account for temporal autocorrelation [Santer *et al.*, 2008]. The confidence interval in the study is stated at the 95% confidence level. To remove the temporal autocorrelation, the standard error of the trend is computed by $s_b = \sqrt{\frac{s_e^2}{\sum_{t=1}^{n_t} (t-\bar{t})^2}}$, $s_e^2 = \frac{1}{n_e - 2} \sum_{t=1}^{n_t} e(t)^2$, $n_e = n_t \frac{1-r_1}{1+r_1}$, where s_b is the standard error, s_e^2 is the variance of the linear regression residuals, t is the time index, $\bar{t}$ is the average time index, $e(t)$ is the linear regression residual, n_t is the total number of time series, n_e is the adjusted number of time series after removing temporal autocorrelation, r_1 is the lag-1 temporal autocorrelation coefficient of $e(t)$.

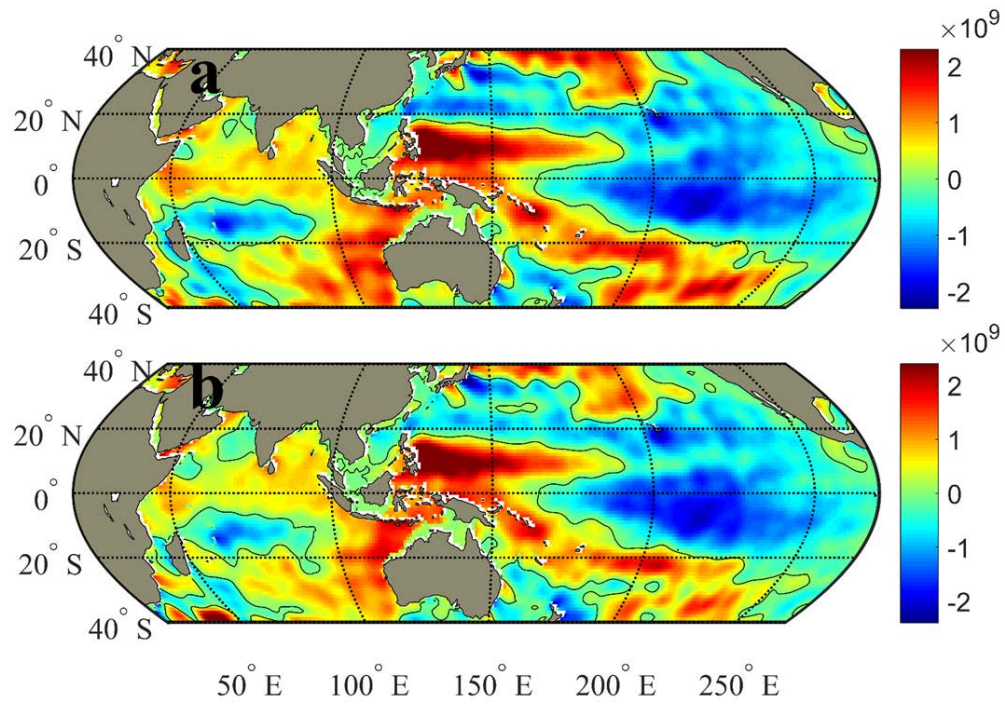
References

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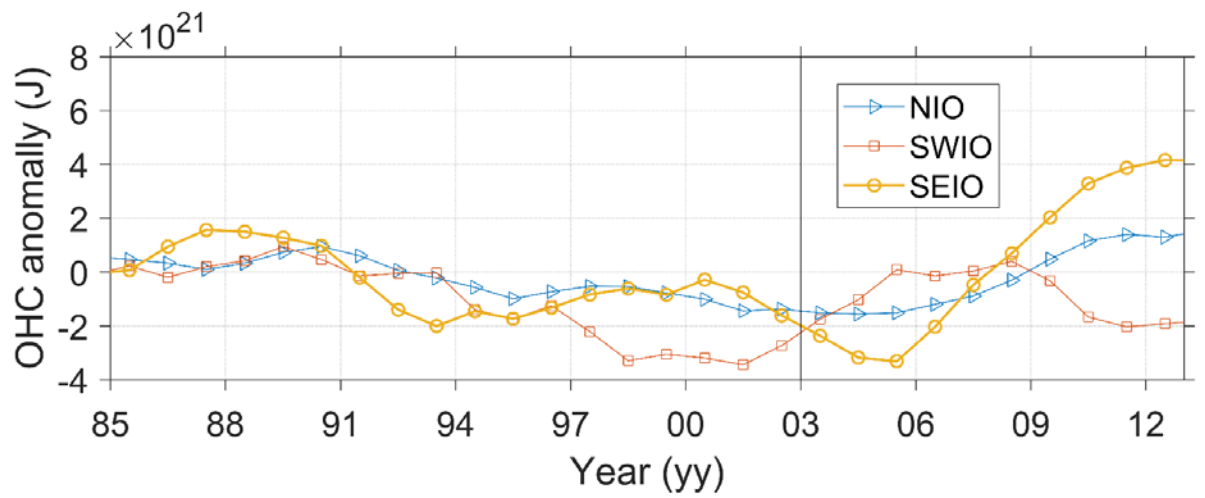
Santer, B. D., P. W. Thorne, L. Haimberger, K. E. Taylor, T. M. L. Wigley, et al. (2008), Consistency of modelled and observed temperature trends in the tropical troposphere, *Int. J. Climatol.* , 28(13), 1703-1722, doi:10.1002/joc.1756.



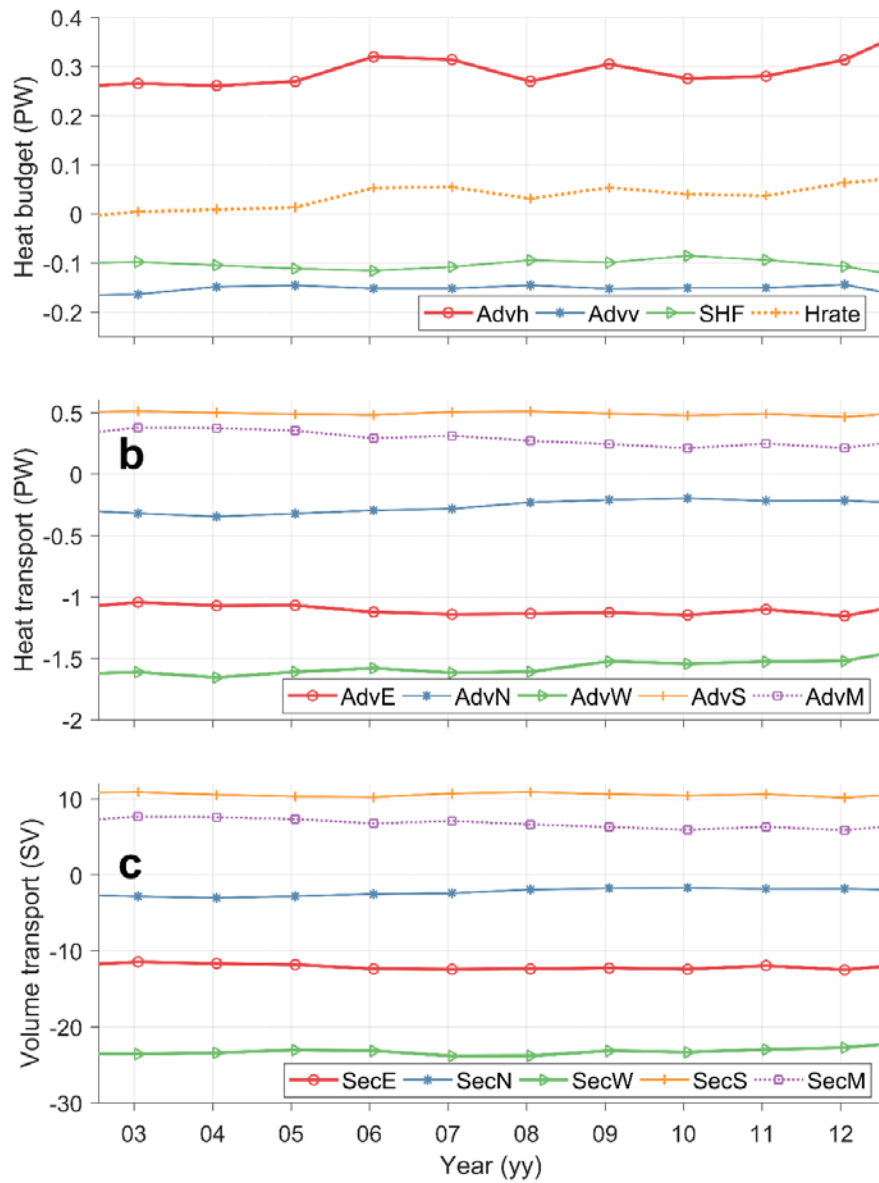
FigureC 1. The ocean basins defined in the study. NIO, SWIO, SEIO, PO, AO, and SO represent North Indian Ocean, Southwest Indian Ocean, Southeast Indian Ocean, Pacific Ocean, Atlantic Ocean, and Southern Ocean respectively. The Arctic Ocean is not discussed in the study.



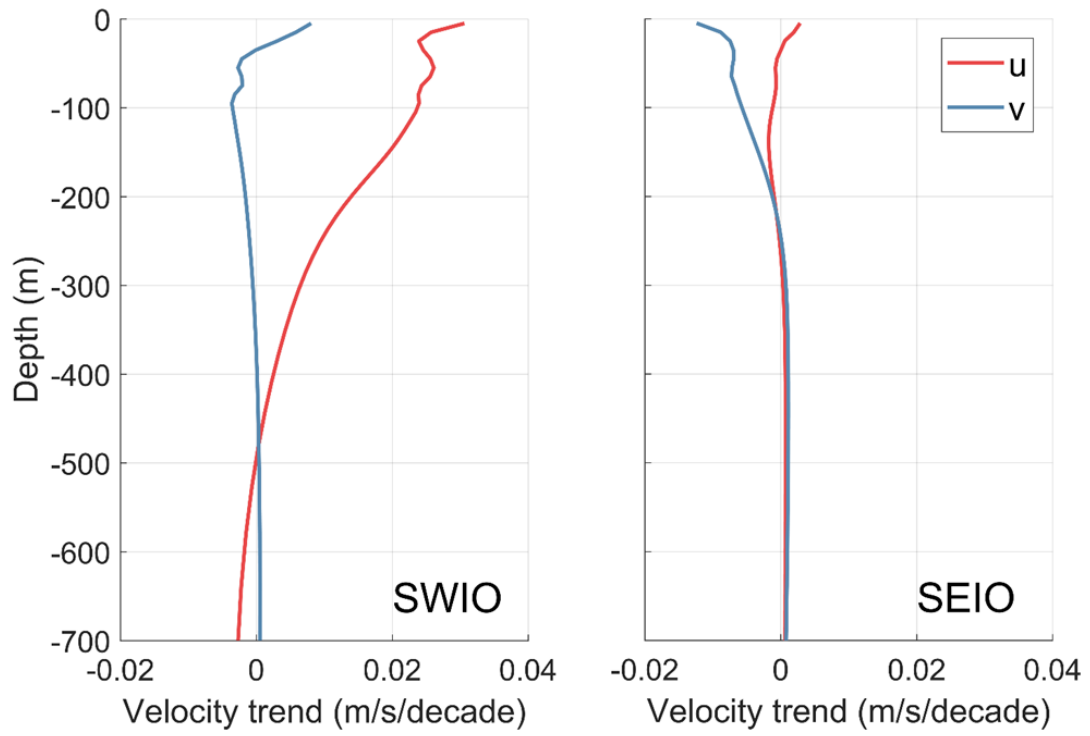
FigureC 2. The OHC₇₀₀ linear trends (units: $\text{J/m}^2/\text{decade}$) during the 2003-2013 (a), 2002-2012 (b) in the upper ocean (0-700m). The black contour line is shown for 0. The data is from Levitus. Black points in the shading color indicate the trends in those locations are significant in statistics ($P < 0.05$). a and b indicate the heat content trend is not sensitive to the end points of the trend analysis.



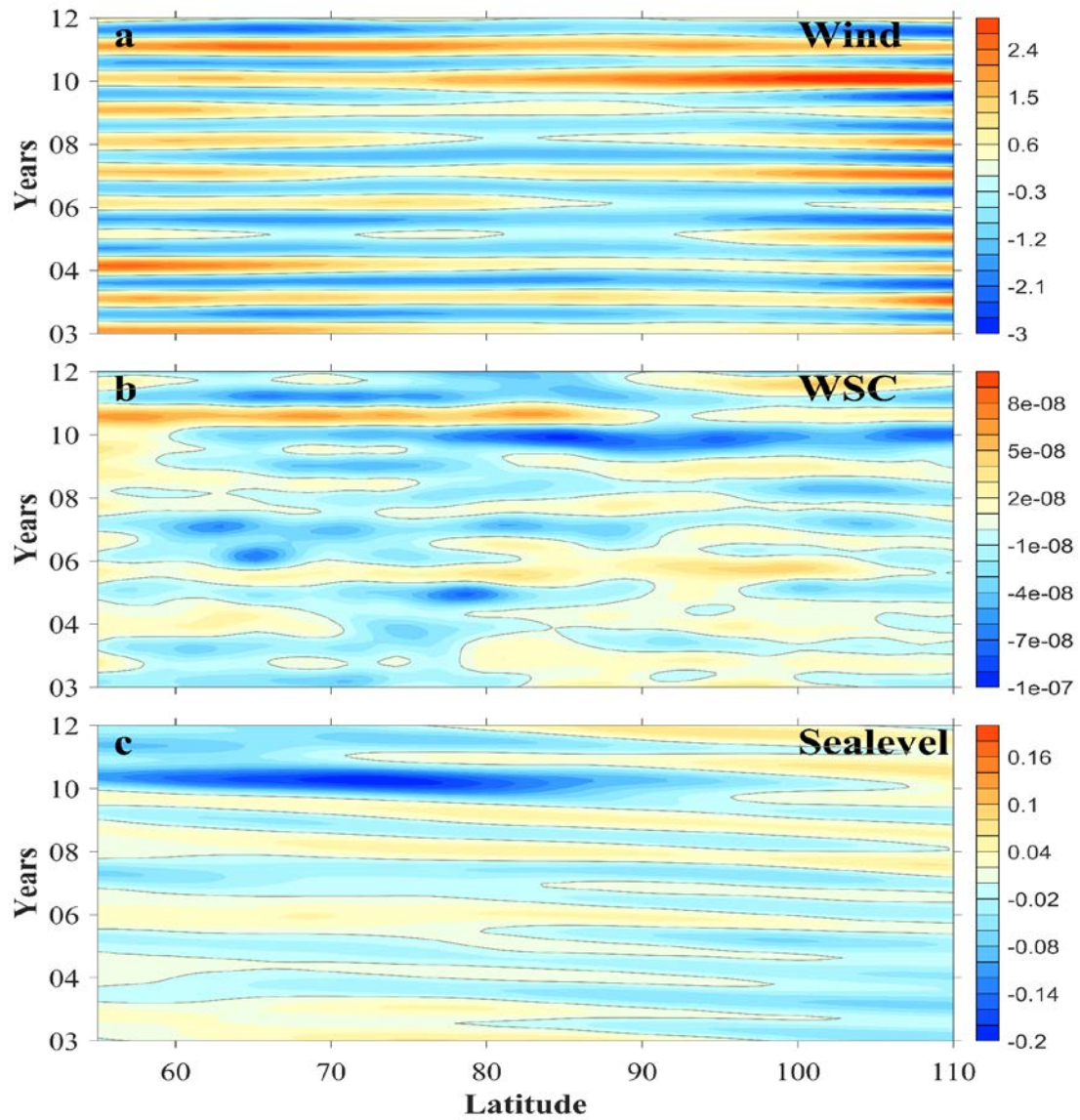
FigureC 3. The time series of OHC₇₀₀ anomalies in three regions of Indian Ocean. NIO, SWIO, and SEIO represent North Indian Ocean, Southwest Indian Ocean, and Southeast Indian Ocean respectively, which are defined in FigureC 1. The anomalies relative to the reference period 1990-2002.



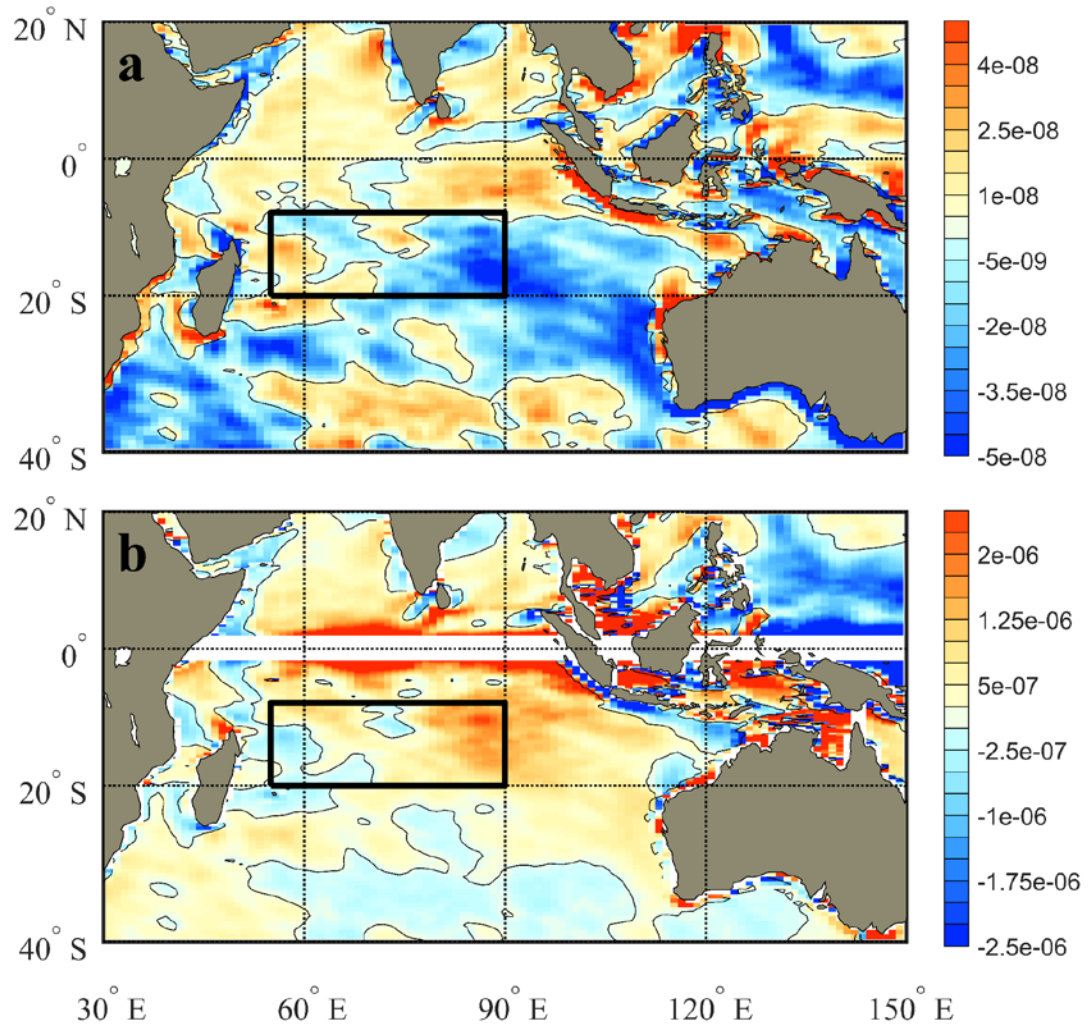
FigureC 4 The time series of heat budget (a), heat transport (b), and volume transport (c) in the SEIO. a, storage rate (Hrate), horizontal heat advctions (Advh), vertical heat advctions (Advv), and net surface heat flux (SHF) for the SEIO. b, the heat transport in the east (AdvE), north (AdvN), west (AdvW), south (AdvS) boundary of the SEIO and the AdvM is the heat transport cross the middle latitude in the SEIO. c, the volume transport in the five sections. The positive is northward and eastward and negative is southward and westward. The section locations are plotted as white box in the 5.3a. PW is 10^{15} J/s.



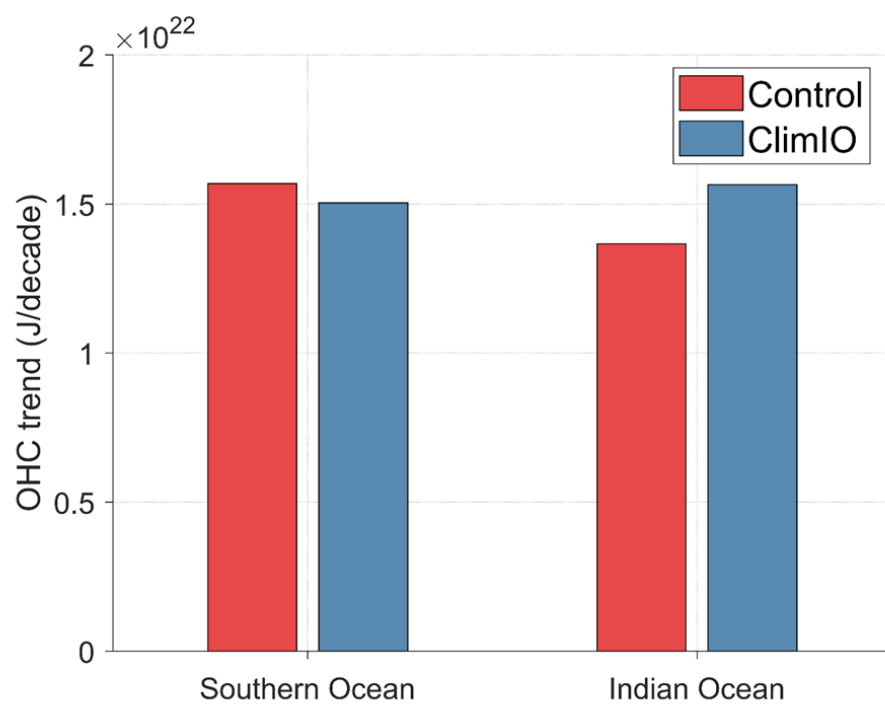
FigureC 5. The vertical profile of velocity trends in the Southwest Indian Ocean and Southeast Indian Ocean. u and v are meridional and zonal directions respectively. The velocity is regional-mean velocity in 60-85°E and -15 to -8°S and 95-110°E and -25 to -15°S.



FigureC 6. The time-longitude of the low-frequency (>12 months) sea level anomalies (a, units: m), wind stress curl anomalies (b, units: N/m^3), and zonal wind speed anomalies(c, m/s) along the 12°S-16 °S band. Since the zonal wind is easterly, the negative value in panel a indicates a strengthened easterly trade wind and positive value vice versa. The wind stress curl is negative in this region, the negative value in b represent a strengthened wind stress curl (enhanced positive Ekman upwelling velocity or sea level falls).



FigureC 7. The linear trends of wind stress curl (a, Pa/m/decade) and Ekman pumping velocity (b, m/s/decade) during the 2003-2013. The thick black box indicates the region where the sea level subsides in Figure 5.3a. In the black box, the negative value in panel a indicates a strengthened negative wind stress curl which induces an enhanced Ekman upwelling velocity in the black box of panel b.



FigureC 8. The trends of Southern Ocean and Indian Ocean for the control and IOclim cases. The trends are all significant in statistics ($P < 0.05$).

Appendix D

PUBLISHED PAPERS DURING MY PHD STUDY

In preparation:

Liao, E.H., X.-H. Yan, and Y.W. Jiang, The strengthened southward heat transport in the Indian Ocean during the warming hiatus. **Ready to submit.**

In revision:

Liao, E.H., Leo Oey, X.-H. Yan, L. Li, and Y.W. Jiang, The separation of the China Coastal Current over the Taiwan Bank in winter. **Under revision in Journal of Physical Oceanography.**

Li, L., X.G. Guo, **E.H. Liao**, and Y.W. Jiang, Subtidal variability in the Taiwan Strait induced by combined forcing of winter monsoon and topography. **Accepted in Science China Earth Science.**

Published:

Liao, E.H., X.-H. Yan, and Y.W. Jiang (2017), The role of coastal-trapped waves on the 2008 cold disaster in the Taiwan Strait, **Ocean Dynamics**, 1-9, doi:10.1007/s10236-017-1042-7.

Liao, E.H., W. F. Lu, X.-H. Yan, Y.W. Jiang, and A. Kidwell (2015), The coastal ocean response to the global warming acceleration and hiatus, **Scientific Reports**, 5, 16630, doi:10.1038/srep16630.

Appendix E

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