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Observations and Modeling Forcing Mechanisms for the Coastal Dynamics of the Upper Gulf of Thailand

Suriyan Saramul
Old Dominion University

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OBSERVATIONS AND MODELING FORCING MECHANISMS FOR THE COASTAL DYNAMICS OF THE UPPER GULF OF THAILAND

by

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A Dissertation Submitted to the Faculty of Old Dominion University in Partial Fulfillment of the Requirements for the Degree of

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ABSTRACT

OBSERVATIONS AND MODELING FORCING MECHANISMS FOR THE COASTAL DYNAMICS OF THE UPPER GULF OF THAILAND

Suriyan Saramul
Old Dominion University, 2013
Director: Dr. Tal Ezer

A numerical model based on the Princeton Ocean Model (POM) with ~1 km horizontal grid and 21 vertical layers has been used to study the influence of wind stresses, river discharges, surface heat fluxes and tides on the three-dimensional circulation of the upper Gulf of Thailand (UGOT). Analysis of observations, including sea level data, provided additional information to support the model simulations. The UGOT is a shallow coastal system, ~1°×1° in size with the average depth of only 15 m. Sensitivity studies evaluate how the dynamics is affected by surface wind stresses, river discharges, and surface heat fluxes. The impact of the low latitudes (~13 °N) on the dynamics is tested by comparing the simulations to results obtained when simulations are modified to mid-latitudes (~45 °N). The model results show that the circulation in the UGOT is strongly affected by the seasonal monsoon winds, changing from counter-clockwise during the northeast monsoon to clockwise during the southwest monsoon. The vertical distribution of velocities is found to follow the classical wind-driven Ekman dynamics, even though the UGOT is shallow and low latitude. The flow tends to be more barotropic and with the wind direction in the shallow area where the water depth is less than the Ekman layer, while in the deeper channels a two-layer seems to develop. The results from the three-dimensional numerical model are compared well with a simple wind-driven analytical model. The impact of river discharges is the formation of a coastal jet of freshwater along the west coast, whereas the effect of Coriolis parameter is to advect the freshwater plume.
farther south along the west coast in mid-latitudes compared to low latitude simulations. Satellite sea surface temperature data generally agrees with the simulated spatial structure. The influence of ENSO resulted in up to 2–4°C inter-annual variations in water temperatures in the GOT. Analysis of sea level data shows in general a rate of relative sea level rise in Thailand of 6 mm yr⁻¹, which is ~3 times faster than the global mean rate; even higher rates (>10 mm yr⁻¹) are found near Bangkok, where the land is sinking. The seasonal sea level cycle shows differences between the stations in the GOT and Andaman Sea, indicating influence from large scale wind and pressure patterns. A positive/negative correlation between mean sea level anomaly and sea level atmospheric pressure is found at the GOT/Andaman Sea. The results from this research will help to better understand the dynamics of the UGOT and improve future modeling efforts.
To my wife, Thitiya Saramul and my daughter, Insuk Saramul
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CHAPTER 1

INTRODUCTION AND RESEARCH QUESTIONS

Many big cities around the globe, especially in developing countries, are facing environmental impact problems. Many of them are trying to produce both agricultural and industrial goods, concerning how environmental issues may affect the production. Bangkok, the capital city of Thailand, is one such big city that is faced with this problem. Highly polluted water from the industrial sector and untreated sewage from urbanization have been discharged into the upper Gulf of Thailand (hereafter UGOT) through four main rivers that situated along the northern part of the UGOT. According to Chongprasith and Srinetr [1998], on-land human activities, for example fish farming, pig farming and other agricultural activities are the main reasons that high organic substances have been transported into the UGOT. This leads to eutrophication that often occur in the UGOT. In order to solve or to manage such problems, the hydrodynamic features, such as the flow circulation, stratification, etc., of the UGOT are needed to be understood better. The hydrodynamics features in the UGOT can be studied through the available observed data or by numerical model simulations. Since field measurements are costly and time consuming, observed data in the UGOT are very sparse [Buranapratheprat, 2008]. But some data, such as temperature, salinity, current velocity, etc., are still available [Neelasri, 1981; Michida et al., 2006; Matsumura et al., 2006; Buranapratheprat et al., 2006, 2008a, 2009] through projects such as the SEAWATCH Thailand program. Therefore numerical modeling of the UGOT is another way that has been taking into account in studying the area. The first circulation features obtained from simple model simulations and analysis of year-long salinity data were shown in NEDECO [1965]. Buranapratheprat [2008] reviewed the previous investigations both field measurement and numerical
modeling results (two-dimensional) and concluded that the counter-clockwise circulation clearly exists during northeast monsoon (as shown by both model simulations and observed data), while during southwest monsoon, both clockwise and counter-clockwise circulations are presented in the UGOT (model shows only clockwise, while observed shows both). So, the mechanisms of the existence of the counter-clockwise circulation during southwest monsoon are not clearly understood and needed to be further investigated. Consequently, Buranapratheprat et al. [2009] developed a three-dimensional model for the UGOT and their results show similar patterns as those obtained by two dimensional models. They concluded that the variations of the wind field, the interaction with water coming from the lower Gulf of Thailand, the influence of river discharges and bottom morphology all might play important roles.

In addition, the UGOT is facing not only environmental problems but also a coastal erosion problem which is a severe problem in many deltaic coastal systems. Sediment in the UGOT is classified as mud [Uehara et al., 2010; Emery and Niino, 1963]. From the study of Uehara et al. [2010], the coastal erosion or the variation of seabed elevation at the uppermost of the UGOT (breakwater pilot study project) is primarily caused by variations of wave height and wave direction. From this study, they concluded that to best understand coastal erosion processes, the across-shore sediment transport might need to be studied. But from the numerical study of shoreline change at the same location caused by wave conditions, Ekphisutsuntorn et al. [2010a,b] stated that the coastal erosion in this area is mainly controlled by alongshore sediment transport (across-shore sediment transport is neglected). The two-dimensional [Guan et al., 1998] and full three-dimensional hydrodynamics incorporated with sediment transport has been studied both in idealized estuaries [Wang, 2002] and realistic estuaries [Byun et al., 2004; Byun and Wang, 2005; Wang, 2000; Wang et al., 2005; Guan et al., 2005]. Furthermore, Whitehouse et al. [2000] suggested that on intertidal areas, wetting and drying processes is a key factor that
control the hydrodynamics. According to the suggestion of Whitehouse et al. [2000],
the hydrodynamics sediment transport model incorporated with wetting and drying
scheme has been applied for study sediment transport in San Francisco Bay
[Uchiyama, 2005] and Shirakawa intertidal flat, Japan [Uzaki and Kuriyama, 2007; Uzaki et al., 2010]. Coastal erosion can be strongly affected by sea level rise, especially in the UGOT, a low-lying area. Coastal erosion together with sea level rise can cause severe damage to the city of Bangkok. In addition, the flooding that may appear during the summer monsoon season can intensively increase the damage to the population of Bangkok and its suburb. Recent study has shown that an absolute long-term sea level rise at Ko Lak station, where the benchmark is stable and has a long tidal record, has a value of $3.0 \pm 1.5 \text{ mm yr}^{-1}$ [Trisiratsatayawong et al., 2011]. They also mentioned two other stations that also have almost the same rate. It is clearly shown that sea level in Thailand is rising faster than the global mean, which is $1.7 \pm 0.3 \text{ mm yr}^{-1}$ [Church and White, 2006]. The study presented here thus analyzed sea level data and show that at some locations the sea level is rising now as much as 10 times faster than the global mean due to land sinking.

The goals of this study are to investigate forcing mechanisms that control hydrodynamics features in the UGOT using a three-dimensional hydrodynamic model. The forcing parameters are composed of tides, wind fields, river discharges, and surface heat fluxes. The tidal record from stations in the Gulf of Thailand (hereafter GOT) and Andaman Sea are also included in the analysis of tides and sea level rise in Thailand. There are more than 20 tide gauge stations in Thailand covering both the GOT and Andaman Sea. What is the type of tides and what is the rate of sea level rise in Thailand? Are there any concerns or restrictions when calculating the rate of sea level rise in Thailand? Since Thailand is under the influence of monsoon system, how does the monsoon affect the seasonal sea level cycle and circulation pattern? Since the UGOT is shallow and located at low latitude (the Coriolis effect is expected to be
small), what are the mechanisms that control the clockwise and counter-clockwise circulation and the roles of friction (bottom and surface), buoyancy, Coriolis and external tides? Preliminary result show coastal plumes of warm/cool water disperse along the coast and small-scale spatial patterns, even when heat flux, wind stress and sea surface temperature are spatially constant and river discharge is neglected, so what is the mechanism that controls these characteristic? The observed and model sea surface temperatures show inter-annual differences of up to 2–4 °C, so the role of El Niño-Southern Oscillations (ENSO) due to large-scale Pacific variations are investigated as well.

This dissertation is organized in the following. A background on the GOT and the UGOT is presented in Chapter 2. The analysis of tidal records, harmonic analysis, sea level rise, and seasonal sea level cycle in the GOT and Andaman Sea is discussed in Chapter 3. Chapter 4 concentrates on the simulation of tides in the UGOT. This will include model calibration and model validation. Then in Chapter 5, sensitivity studies with the tidal model described in Chapter 4 focuses on three different forcings, wind stresses, surface heat fluxes, and freshwater discharges. The conclusions of this dissertation research are summarized at the last chapter, Chapter 6.
CHAPTER 2

BACKGROUND OF THE STUDY

In order to clearly understand the hydrodynamic features in the UGOT, first it requires knowledge of the dynamic features of the whole gulf, the GOT. Geographically, the GOT lies between longitudes 99°E and 105°E and between latitudes 6°N and 14°N (see Figure 1). This area has been called the Sundra Shelf. The GOT is a large shallow semi-enclosed sea that connects to the South China Sea through the bay mouth. Its area is approximately $3.5 \times 10^5 \text{ km}^2$ (one-tenth of the South China Sea area) and receives the total freshwater discharge (between $59 \times 10^6$ and $150 \times 10^6 \text{ m}^3 \text{ d}^{-1}$) through the four major rivers located at the upper part of the GOT [Suvapepun, 1991]. The bottom topography is relatively flat with an average depth of approximately 45 m; the maximum depth of 80 m can be found at the center of the GOT. At the bay mouth, a sill with a depth of 65 m is roughly presented at the mid-way between Point Ca Mau, Vietnam and Kota Bharu, Malaysia (approximately 460 km width) [Faughn and Taft, 1967; Standsfield and Garrett, 1997]. Another sill with a depth roughly of 25 m is found 95 km from Point of Ca Mau [Faughn and Taft, 1967; Robinson, 1974]. These are two locations through which water masses in and outside the GOT (cold, saline water mass from South China Sea; freshwater from Mekong River, etc.) can be thoroughly exchanged [Robinson, 1974; Laongmanee et al., 2002; Aschariyaphotha et al., 2008]. Wyrtki [1961] was probably the first to investigate in great detail both the physical and biological resources in East Asian region, included the GOT during the Joint Thailand–Vietnam–U.S. NAGA expedition (1959–1961). According to the NAGA expedition data, Robinson [1974] concluded that the GOT might be classified as a classical two-layered, shallow water estuary and the general circulation and physical properties experience either longer period (seasonal) or
shorter period variability. *Robinson* [1974] and *Standsfield and Garrett* [1997] also concluded that during northeast monsoon if the river flood from Mekong River does not enter the South China Sea until November or later it might play a crucial role as a source of freshwater at the lower GOT (eastern coast). The circulation in the GOT is mainly controlled by monsoonal winds, density gradients, and tides [Wyrtki, 1961; Robinson, 1974; Fuh, 1977; Yanagi and Takao, 1998a; Aschariyaphotha et al., 2008]. Based on the 5 m depth current data from SEAWATCH Thailand program during April 1993 to August 1994, [Wattayakorn et al., 1998] concluded that the circulation in the GOT is mainly controlled by the water from the South China Sea. Tides in the GOT are the result of two tidal waves that move in opposite directions, one from the South China Sea and another from the GOT itself (reflected wave) [Wyrtki, 1961; Fuh, 1977]. In general, the amplitude of diurnal tide is larger than that of semi-diurnal tide [Robinson, 1974; Yanagi and Takao, 1998b]. Wyrtki [1961] analyzed observed sea level records in the South China Sea region and presented the co-tidal and co-range lines of the 4 major tidal components which are M2, S2, K1, and O1. The velocity of the semi-diurnal tidal wave in the South China Sea is decreased when reaching the Sundra Shelf, its refraction causing high amplitude to be developed. After it passes the edge of the shelf, it travels 8 hours across the shelf reaching the coast of Malaya and splits northward to the GOT and southward. The northward movement causing the clockwise circulation around the amphidromic point near the east coast with an amplitude of less than 20 cm. The strong tide with amplitude of 82 cm is observed in the northern part of the GOT. Unlike the semi-diurnal tide, the high diurnal tidal wave reaches the edge of the Sundra Shelf with a phase of 13 hours relative to the phase in Pacific Ocean (between 140°E and 150°E). When the fast tidal wave almost reaches the coast of Malaya, it splits northward and southward like the semi-diurnal wave. The northward movement causes the counter-clockwise circulation of small amplitude around the amphidromic system centered
close to the coast of Malaya, while the large amplitude of 113 cm is observed in the northern part of the GOT. Based on the three-dimensional numerical model of the GOT [Yanagi and Takao, 1998a; Fang et al., 1999; Singhruck, 2002; Tomkrutoke and Sirisup, 2010; Tangang et al., 2011], the amphidromic systems of the semi-diurnal and diurnal tides are in good reproduced and they are well agreement with the observation shown in Wyrtki [1961]. A geostrophic flow study based on temperature and salinity and satellite altimetry data during 1995-2001 shows at the entrance of the GOT a strong southwestward flow. In addition, basin-wide clockwise and counter-clockwise circulations inferred from seasonal geostrophic flows were found in the GOT during northeast and southwest monsoons, respectively [Sajisuporn et al., 2010].

Results from an intensive hydrographic cruise survey in the western part of the GOT in September 1995 and April to May 1996 and NAGA cruises report [Robinson, 1974; Yanagi et al., 2001] point out that strong northeast monsoon wind and surface cooling in January caused vertically well-mixed condition in the GOT to be developed. As a result, an inverse estuarine-like circulation is developed during northeast monsoon. Unlike January, large surface heating and weak southwest monsoon wind produces strong stratification caused by the surface Ekman transport and estuarine circulation to be seen in April. During the southwest monsoon, the largest water exchange between the GOT and the South China Sea is expected. In September, large river discharge from the head of the GOT, moderate surface heating and moderate southwest monsoon wind create moderate stratification.
Figure 1. Map of South China Sea. The area with color is the GOT. The bottom topography of the GOT is based on ETOPO1 data obtained from NOAA National Geophysical Data Center [Amante and Eakins, 2009] (http://www.ngdc.noaa.gov/mgg/global/global.html). On the western side of the GOT is Andaman Sea. The small blue box represents the UGOT.

Located at the head of the GOT is the UGOT (sometimes also called the Gulf of Bangkok or the Bight of Bangkok), which is nearly a square semi enclosed sea covering the area between latitudes 12.5°N to 13.5°N and longitudes 100°E to 101°E (approximately 10^4 km^2; see Figure 1 and 13). In the south, it connects to the whole GOT at latitude 12.5°N. The sources of freshwater in the UGOT are from four main
rivers that situated along the head of the UGOT from west to east, namely Mae Klong, Tha Chin, Chao Phraya and Bang Pakong River (see more in section 5.3 and in Figure 13). These four rives carry eroded sediment from the highlands which is deposited at the river mouths causing lowlands or tidal flats to be developed along the head of the UGOT with average height of 2 m above mean sea level [Uehara et al., 2010]. Chao Phraya River is the third largest river in Southeast Asia after the Mekong and Irrawaddy Rivers [Tanabe et al., 2003]. The sediment layers in the UGOT contain fined grain of mud and clay [Winterwerp et al., 2005; Uehara et al., 2010]. The water depth gradually increases from north to south with a slope of less than 1/1000. Specifically, the slopes of 1/400 to 1/650 have been observed near the shore, which is consistent with a nautical chart [Winterwerp et al., 2005]. The average depth of the UGOT is approximately 15 m while the maximum depth is approximately 55 m, which can be found near the southern part of the east coast.

The climate of the whole GOT or of Southeast Asian region is largely influenced by monsoonal winds. In general the southwest monsoon is set up in May and ends around September. The southwest winds prevail across the Indian Ocean and the Bay of Bengal, bringing considerable rainfall to Thailand between July and October. In general, an increasing of rainfall over Thailand results in river floods that peak one month later; these floods bring abundant freshwater to the UGOT causing seawater dilution near the coastal areas [Robinson, 1974]. The northeast monsoon (dry season) normally starts in November and lasts until February, but occasionally can last until March or early April. It is a period of variable moderate wind over the GOT. It brings cold and dry air from Siberia. There are two other transition periods of the opposing monsoon winds: southerly wind in between March and April and northerly wind in October.
Temperature and salinity data in the UGOT from 6 cruises between October 2003 and July 2005 are shown in Matsumura et al. [2006] together with other parameters. Buranapratheprat et al. [2008a] analyzed these data and found that stratified conditions in the UGOT are expected for the whole year except December, when surface heat loss (cooling), low river discharge and high wind speed (northeast monsoon wind) exist. Strong stratification is present between September and October due to the existence of large river discharge and moderate surface heating. Even when high surface heating occurs in April and May, the stratification is not that strong because of low river discharge. The weak to moderate stratification develops during January to March and June to August is largely influenced by river discharge and surface heating while tidal and wind stirring is insufficient to completely mix the water column. The surface temperatures range from \(~26^\circ C\) in January 2004 (winter time) to \(~31^\circ C\) in May 2004 (summer time). The low surface salinity of 16–17 psu and high surface salinity of 33–34 psu have been observed close to the west coast in October 2003 and near the mouth of the UGOT in July 2005, respectively. In terms of horizontal gradient between the head and the mouth of the UGOT, \(~7–9\) psu salinity gradients could be observed in October for both years, while temperature gradients larger than \(~1^\circ C\) are rarely observed [Buranapratheprat et al., 2008a].

The freshwater discharge through the four major rivers has been reported both monthly averaged and annual averaged over several years [Robinson, 1974; Snidvongs, 1993; Yanagi et al., 2001; Buranapratheprat et al., 2008b]. From these results, it is clearly seen that September to October is the peak of freshwater discharge, while during March to April small river discharge occurs [Singhrattana et al., 2005a]. In terms of yearly averaged, Snidvongs [1993] stated that Chao Phraya River is the largest river with the average discharge rate approximately of 430 m$^3$ s$^{-1}$. Mae Klong, Bang Pakong, and Tha Chin Rivers are the second, third, and fourth with the average discharge of 320, 220, and 50 m$^3$ s$^{-1}$, respectively.
Similar to the GOT, the circulation pattern of the UGOT is controlled by the combination of monsoonal winds, effects of freshwater runoff (density-driven), and tides [Buranapratheprat et al., 2006, 2008b] which can be inferred from the horizontal distribution of salinity [NEDECO, 1965; Buranapratheprat et al., 2008b]. The distribution of the surface salinity from field measurements [Matsumura et al., 2006] is in good agreement with the seasonal mean flow based on model simulations (two-dimensional model of Buranapratheprat et al. [2002]). The model shows clockwise and counter-clockwise circulations during southwest and northwest monsoons, respectively. In general, the counter-clockwise circulation during the northeast monsoon from the model result is consistent with the observed data [Buranapratheprat et al., 2002, 2008b], but during the southwest monsoon, disagreement of the circulation pattern between model result and field measurement in particular months has been observed [Buranapratheprat et al., 2008b]. The analysis of current measurement data at 5 m depth obtained from the SEAWATCH Thailand program during 1996 to 1998 by Booncherm [1999], found that the monthly mean residual flow pattern in the UGOT corresponds to the wind directions and the counter-clockwise and clockwise circulations both exist during southwest monsoon; clockwise circulation has been observed especially in July. A similar circulation pattern has also been reported in Wattayakorn et al. [1998]. They used the current measurement data from SEAWATCH Thailand program but in different period (during April 1993 and August 1994). According to their result, they suggested that during March to August the clockwise circulation in the UGOT was a result of the sweeping of the clockwise circulation in the GOT. The reverse circulation in the GOT occurred by September which created the counter-clockwise circulation in the UGOT with no flow separation at the eastern corner of bay mouth. In November, the clockwise circulation in the UGOT caused by the flow separation at the eastern corner of bay mouth has been observed.
Based on one year tidal record in the head of the UGOT and at the bay mouth, tides in the head of the UGOT are in the range of 2-3 m and are generally classified as mixed, mainly semi-diurnal tide. Unlike at the head of the UGOT, tides at the bay mouth are defined as mixed, mainly diurnal tide (see Chapter 3). In the UGOT, the residual north-south current (order of $10^{-2}$ cm s$^{-1}$) is less than that of the tidal current (2 orders of magnitude smaller) [Snidvongs, 1993].

In the ocean or in the atmosphere, an important length scale is the so-called Rossby radius of deformation. It is, in general, the important length scale in rotating systems [Cushman-Roisin, 1994; McWilliams, 2006]. There are two types of Rossby radii, external or barotropic (homogeneous water) and internal or baroclinic (stratified water) Rossby radii, which can be expressed as $R_o = \sqrt{gH/f}$ and $R_i = NH/f = \sqrt{g'\bar{H}/f}$, respectively, where $g$, $g'$, $N$, $H$, and $f$ are gravitational acceleration, reduced gravitational acceleration, buoyancy frequency, water depth, and Coriolis parameter, respectively. For the GOT, Yanagi and Takao [1998b] calculated the barotropic Rossby radius at 9°N (where averaged depth was 40 m) to be 870 km. It is about two times the width of the mouth of the GOT as mentioned above (about 460 km). They concluded that the Coriolis force in the GOT might play a minimal role in tidal phenomena.

For the UGOT, the averaged depth $H \sim 15$ m and latitude $\sim 13°N$ (then $f = 3.27 \times 10^{-5}$ s$^{-1}$). This leads to the barotropic Rossby radius of approximately 370 km, which is almost four times larger than the width of the UGOT (a distance between western and eastern sides is about 100 km). Therefore, it can be concluded that for the barotropic flow, the Coriolis force is not the major force in the UGOT.

However, if stratification exists, the Coriolis force in the UGOT might or might not be important. The internal Rossby radius in the UGOT, where $g' = 0.145$ m s$^{-2}$, is roughly 45 km, which is about half the width of the UGOT. So, for baroclinic dynamics, the Coriolis force can play a role in the UGOT. The research will address
this possibility when discussing simulations of river plumes, since those cases involve stratification induced by the freshwater input.
CHAPTER 3

ANALYSIS OF TIDES IN THE GULF OF THAILAND

Based upon a mean tidal range, tides in the GOT are considered to be either micro- (tidal range < 2 m) or meso-tidal (tidal range is between 2 and 4 m) depending on geographic location. The UGOT and the Central GOT are classified as meso-tidal while other areas, for example the lower GOT, are micro-tidal system. The type of tides in the GOT is also dependent on geographic location which will be discussed in the following section. The analysis of tides and sea levels from short- and long-term tidal records will also be analyzed in this chapter.

3.1 BACKGROUND

There are approximately 27 tide gauge stations operated in Thai Waters (both in the GOT and in Andaman Sea). More specific, about 10 stations are situated along the coast of the UGOT. The Hydrographic Department, Marine Department, and Port Authority of Thailand are three main governmental agencies that operate tide gauge stations in Thailand. The Hydrographic Department as part of the Royal Thai Navy is responsible for publishing predicted tide tables and high–low water in some stations available to public. The predicted tide in Thai Waters is estimated based on 112 tidal harmonic constituents using software provided by Tidal Laboratory of the Flinders Institute of Atmospheric and Marine Sciences, Australia [Hydrographic Department, 2001].

Based on availability of quality data, only 18 tide gauge stations (Figure 2) will be used in this study. The description of each tide gauge station for this study is shown in Table 1. The data are obtained from 4 sources of surface elevation
data which are Marine Department (MD), Hydrographic Department (HD), Permanent Service for Mean Sea Level (PSMSL; http://www.psmsl.org/) [Woodworth and Player, 2003; Holgate et al., 2013], and the University of Hawaii Sea Level Center (UHSLC; http://uhslc.soest.hawaii.edu/ and http://ilikai.soest.hawaii.edu/). Data provided by PSMSL is available only monthly and yearly (mean sea level), while others are hourly. Note that data obtained from UHSLC is available either hourly, daily, and monthly. The periods of available data from these four sources are different from one station to another. For example Hua Hin and Sattahip stations are available for only 5 and 8 years respectively though these 2 stations are used as tidal forcing at open boundary for the UGOT model (see later chapter). The longest records, ~70 years, are obtained from PSMSL though only monthly data are available. An hourly sea surface elevation plot for the whole month of January 2004 from 9 selected tide gauge stations is shown in Figure 3 (black solid line). From Figure 3, it is clearly seen that tides in Andaman Sea (Ko Taphoa Noi station) are different from tides in the GOT. However, in the GOT itself tides from each station differ in both magnitude and type. Tides in the GOT seem to be amplified as they propagate northward to the shallower UGOT. Tidal ranges are found to be approximately 1 m at Geting station (GT; southernmost) and to be about 4 m at Bang Pakong station (BK; up north). A plot of sea surface elevations between Hua Hin and Sattahip stations (stations at open boundary of the UGOT model) shows that tides at Hua Hin are larger than at Sattahip and tidal phase at Hua Hin lags behind at Sattahip approximately an hour. Hence, it might be concluded that tides in the UGOT propagate counter-clockwise along the coastline. The tidal pattern for each tide gauge station is described by harmonic analysis in Section 3.2.
Figure 2. Location and list of tide gauge stations used for this study. Note that station 16 and 18 belong to Malaysia.
<table>
<thead>
<tr>
<th>Station</th>
<th>Available data</th>
<th>Data type</th>
<th>Sources</th>
</tr>
</thead>
<tbody>
<tr>
<td>Klongyai</td>
<td>2004</td>
<td>Hourly</td>
<td>MD</td>
</tr>
<tr>
<td>Sattahip</td>
<td>1997–2004</td>
<td>Hourly</td>
<td>HD</td>
</tr>
<tr>
<td>Au Udom</td>
<td>2006–2012</td>
<td>Hourly</td>
<td>MD</td>
</tr>
<tr>
<td>Si Chang</td>
<td>1940–2002</td>
<td>Monthly</td>
<td>PSMSL</td>
</tr>
<tr>
<td>Bang Pakong</td>
<td>1981–2012</td>
<td>Hourly</td>
<td>MD</td>
</tr>
<tr>
<td>Phrachulachomkloa</td>
<td>1940–2011</td>
<td>Monthly</td>
<td>PSMSL</td>
</tr>
<tr>
<td>Mae Klong</td>
<td>1980–2012</td>
<td>Hourly</td>
<td>MD</td>
</tr>
<tr>
<td>Ban Lam</td>
<td>1997–2012</td>
<td>Hourly</td>
<td>MD</td>
</tr>
<tr>
<td>Hua Hin</td>
<td>2000–2004</td>
<td>Hourly</td>
<td>HD</td>
</tr>
<tr>
<td>Klongvale</td>
<td>2006–2012</td>
<td>Hourly</td>
<td>MD</td>
</tr>
<tr>
<td>Ko Mattaphon</td>
<td>1992–2012</td>
<td>Monthly</td>
<td>PSMSL</td>
</tr>
<tr>
<td>Langsuan</td>
<td>2004</td>
<td>Hourly</td>
<td>MD</td>
</tr>
</tbody>
</table>
In this chapter, the analysis of tides in the GOT (tides data obtained from Andaman Sea area, which are situated on the other side of the GOT are also included) was performed using the harmonic analysis technique that is demonstrated in Pawlowicz et al. [2002] and Leffler and Jay [2009]. The result of this analysis is shown in Section 3.2. The analysis of tides in the GOT will help us to understand the characteristics of tides in the GOT and Andaman Sea area. In Section 3.3, the analysis of long-term sea surface elevation will be demonstrated aiming to separate the seasonal and/or long-term trends from tidal oscillation (a fluctuating data).

3.2 HARMONIC ANALYSIS OF TIDES IN THE GOT

The goal of tidal analysis is to separate the tidal signal from sub-tidal or super-tidal variation [Pawlowicz et al., 2002]. The separation of tidal oscillations from sub-tidal can be either by high-pass or band-pass filtering techniques, or it can be done by harmonic analysis; the latter is discussed in this section.

According to Boon [2004], harmonic analysis is a technique that separates sinusoidal signals at specific frequencies related to astronomical tide from the sum of a finite set of sinusoidal signals. Hence, each tidal component has its unique frequency. A least-square fit method is used to determine amplitudes and phases for each sinusoidal signal, which is the so-called tidal component or tidal constituent. In general, observed sea surface elevation is the sum of a set of sinusoidal signals plus non-tidal signal or residual.

How is harmonic analysis of observed surface elevation performed? First, set a predicted surface elevation at time \( t \), \( \eta(t) \) as the sum of \( k \) tidal components, \( \eta(t) = A_0 + \sum_{i=1}^{k} A_i \cos(\omega_i t - \phi_i) \), where \( A_i \), \( \omega_i \), and \( \phi_i \) are amplitude, frequency, and phase of tidal component \( i \), respectively, and \( A_0 \) is mean surface elevation for a series of observed surface elevation. Second, estimate the sum of squared differences (SSD) between predicted \( (\eta(t)) \) and observed surface elevation \( (\eta_t) \). SSD can be expressed as
\[ \sum_{j=1}^{N} [\eta_t - \eta(t)]^2, \] where \( j \) and \( N \) are index and total number of values in the tidal record, respectively. Finally, determine parameters \( A_i \) and \( \phi_i \) for each tidal component in the first step in such a way that SSD will be as small as possible. This is then the best-fit of the predicted to the observed surface elevation in term of least-square fit methods. A basic statistics to check the fit is Root–Mean Square Error (RMSE) and \( R^2 \). By definition, RMSE and correlation coefficient \( R \) can be written as

\[
\text{RMSE} = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (P_i - O_i)^2}
\]

\[
R = \frac{\left[ N \sum_{i=1}^{N} (O_iP_i) - \left( \sum_{i=1}^{N} O_i \right) \left( \sum_{i=1}^{N} P_i \right) \right]}{\sqrt{\left[ N \sum_{i=1}^{N} O_i^2 \right] \left[ N \sum_{i=1}^{N} P_i^2 \right] - \left( \sum_{i=1}^{N} O_i \right)^2 \left( \sum_{i=1}^{N} P_i \right)^2}},
\]

where \( P_i \) and \( O_i \) are observed and modeled values at time \( i \), respectively. From equation (2), \( R^2 \) is just the square of \( R \).

Harmonic analysis of tidal records obtained from stations situated in the GOT and Andaman Sea is illustrated with predicted surface elevation (see Figure 3, gray solid line). Tidal amplitudes and phases obtained from harmonic analysis will be discussed later. Visually predicted surface elevations from all stations shown in Figure 3 well captures the observed surface elevations. There are discrepancies between observed and predicted tides after day 20 at all stations except Ko Taphoa Noi, a station that located in Andaman Sea. The residual after day 20 shows a sinusoidal feature with a period of approximately 2–3 day. Klongyai is the station with largest errors; one explanation might be instrumental error. There is an increase of surface elevation around day 12. This elevation change cause predicted surface elevation to be over-estimated before day 12 and under-estimated afterwards. Please bear in mind that predicted surface elevation is dependent on how many tidal constituents are included.
in the tidal prediction. Predicted surface elevation shown in Figure 3 is based on approximately 30 tidal constituents, therefore it might miss some constituent causing the error. The error is also coming from sub-tidal or super-tidal variation merging in surface elevation signal. Hence, the analysis of residual, a difference between observed and predicted surface elevation, is necessary.

In order to investigate seasonal variation of residual, a year-long residual is plotted (Figure 4). The RMSE and $R^2$ of all stations depicted in Figure 4 are presented in Table 2. Since mean of residual is very small, RMSE and standard deviation of difference are almost the same. Hence only RMSE is depicted in Table 2. From Figure 4 and Table 2, a residual of Ko Taphoa Noi confirms how small the discrepancy is and its residual is totally different from that at the other stations. From Table 2, RMSE of all stations is in the range of 6 to 15 cm, while $R^2$ is in the range of 0.88 to 0.99. The residuals of all tide gauge stations in the GOT are almost the same, especially the peaks shown on day 25, 40, 70, 275, 300, 325, and 345 (approximated). The residual of Bang Pakong is quite variable. One explanation might due to Bang Pakong is situated at the river mouth and it has been affected by the river discharge. At Ko Taphoa Noi station approximately day 360, ~2 m range of residual shows the signal of 2004 Indian Ocean Tsunami. Note that this signal exists only at the Andaman Sea not in the GOT. In general, the residuals seen in the GOT and Andaman Sea are affected by large scale variability.
Figure 3. Surface elevation comparison between observed (black solid line) and predicted (gray solid line) during January 2004.
Figure 4. A difference between observed and predicted surface elevation (top panel; zero vertically shifted in each station) and wind vector (bottom panel) during year 2004.

At the bottom panel of Figure 4, wind vectors are also displayed. Visually, a positive residual seems to follow wind blowing from the south and vice versa for a negative residual; this is more apparent during the winter season (November to February), when winds are variable. Note that not only winds affect surface elevation in the GOT but other factors.
In order to further investigate what factors might cause the change of surface elevation, a plot of power spectrum density of the residual is shown in Figure 5. It appears that additional tidal components may be needed to be included in the prediction. A first peak of 3-4 days period is shown in all station except Ko Taphoa Noi. Other peaks shown in Figure 5 are at 19-21 hours, 11-12 hours, 8 hour, and 5-6 hours period. More interesting is a peak of 8.1042 hour found in all stations (this peak is referred to a non-tidal constituent that has a frequency in between tidal constituents MK3 and SP3).

**Figure 5.** A log-log plot of power spectrum density of residual against frequency for all nine stations in year 2004. A number shown in each station has a unit of hour.

A result of harmonic analysis of tidal records obtained from stations situated in the GOT and Andaman Sea is illustrated in Figure 6 in terms of tidal amplitude and tidal phase of 8 tidal constituents, which are Q1, O1, K1, P1, N2, M2, S2, and
K2. Tidal amplitude of Ko Taphoa Noi clearly demonstrates how tides in Andaman Sea’s station is different from tides in the GOT’s stations. At Ko Taphoa Noi, semi-diurnal tidal components are the major components, specifically M2 tide. An interesting tidal pattern, for example in K1, is that tidal amplitude of K1 is gradually increased from far east station (Klongyai) to the UGOT’s station (Bang Pakong) and gradually decreased from the UGOT to far south station (Geting). At Bang Pakong station, M2 and S2 tides are almost two times larger than that at Sattahip and Hua Hin stations (see Figure 2). From tidal amplitudes, one can calculate a Form number, $F$, a ratio of tidal amplitude of diurnal components K1+O1 to semi-diurnal components M2+S2 for all stations and it is shown in Table 2. Clearly, in the lower and middle GOT tides are diurnal, while in the UGOT they are mixed mainly diurnal except at the head of the UGOT. At the head of the UGOT, tides are mixed mainly semi-diurnal (all five stations as shown in Chapter 4). In addition, Aungsakul et al. [2007] performed harmonic analysis of tidal records in the GOT and Andaman Sea, they found that tides change from mixed mainly diurnal at station near Langsuan to mixed mainly semi-diurnal at stations farther south between Langsuan and Geting and to mixed mainly diurnal at station near Geting. Tidal phases of semi-diurnal components in the GOT propagate in the counter-clockwise fashion, while in the UGOT diurnal components seem to gradually increase in the same fashion as semi-diurnal components. This result might confirm that each tidal component in the GOT propagate in the counter-clockwise direction. This complex structure of tides needs to be taken into account later when tides are introduced in the numerical model. The tides will play a role in mixing processes in the UGOT.
Table 2. Statistics of the difference between observed and predicted surface elevation from some stations located in the GOT and Andaman Sea. Note Form number and type of tides at each station are also shown.

<table>
<thead>
<tr>
<th>Station</th>
<th>RMSE (cm)</th>
<th>R²</th>
<th>Form number, $F$</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Klongyai</td>
<td>15.15</td>
<td>0.90</td>
<td>3.81</td>
<td>Diurnal</td>
</tr>
<tr>
<td>Rayong</td>
<td>11.42</td>
<td>0.94</td>
<td>4.05</td>
<td>Diurnal</td>
</tr>
<tr>
<td>Sattahip</td>
<td>12.73</td>
<td>0.96</td>
<td>2.51</td>
<td>Mixed, diurnal</td>
</tr>
<tr>
<td>Bang Pakong</td>
<td>13.01</td>
<td>0.96</td>
<td>1.30</td>
<td>Mixed, semi-diurnal</td>
</tr>
<tr>
<td>Hua Hin</td>
<td>13.01</td>
<td>0.96</td>
<td>2.08</td>
<td>Mixed, diurnal</td>
</tr>
<tr>
<td>Ko Lak</td>
<td>12.68</td>
<td>0.96</td>
<td>11.31</td>
<td>Diurnal</td>
</tr>
<tr>
<td>Langsuan</td>
<td>14.30</td>
<td>0.93</td>
<td>3.13</td>
<td>Diurnal</td>
</tr>
<tr>
<td>Geting</td>
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<td>0.88</td>
<td>1.57</td>
<td>Mixed, diurnal</td>
</tr>
<tr>
<td>Ko Taphoa Noi</td>
<td>6.14</td>
<td>0.99</td>
<td>0.15</td>
<td>Semi-diurnal</td>
</tr>
</tbody>
</table>
Figure 6. Tidal amplitude in cm (left panel) and tidal phase in degree (right panel) based on harmonic analysis of tides from tide gauge stations situated in the GOT and Andaman Sea.
3.3 ANALYSIS OF LONG-TERM SEA LEVEL IN THE GOT

3.3.1 Sea level changes

The rate of sea level rise is very important information for the population that lives near the coast. Increasing sea level, storm surge, together with tides, can severely damage the properties located at the low-lying areas where people live [Douglas, 2001]. The global rate of sea level rise calculated based on reconstructed sea level data from tide gauge and satellite data between 1950 and 2000 using empirical orthogonal function (EOF) is found to be $1.8 \pm 0.3$ mm yr$^{-1}$ [Church et al., 2004]. Using the same technique but extending sea level data to cover the period of 1870 to 2004, the global rate of sea level rise of the 20th century is determined to be $1.7 \pm 0.3$ mm yr$^{-1}$ [Church and White, 2006].

There are only a few studies of sea level rise in Thailand. However, for the present sea level rise Yanagi and Akaki [1994] studied sea level variability in Eastern Asia using data obtained from PSMSL between 1951 and 1991 and they found the rate of sea level rise at Si Chang, Phrachulachomkloa, Ko Lak, and Ko Taphoa Noi are $0.60 \pm 0.39$, $16.4 \pm 0.85$, $-0.83 \pm 0.22$, and $2.3 \pm 1.1$ mm yr$^{-1}$, respectively. Others have also addressed the rate of sea level rise based on the same source of data at Ko Lak and Sattahip, but for a different period (1940 to 1996). The rate of sea level rise of $-0.36$ mm yr$^{-1}$ is found at both stations [Vongvisessomjai, 2006, 2010]. Si Chang and Ko Lak are approximately 150 km apart with presumably similar oceanographic, meteorological, and tectonic conditions. Therefore it is not clear why we see those differences. The opposite rate from that shown in Yanagi and Akaki [1994], may be caused by a vertical land movement at each station [Trisirisatayawong et al., 2011]. This leads to the study of rate of sea level rise taking a vertical movement at each station [Trisirisatayawong et al., 2011]. The rate of sea level rise they estimated is
based on precise GPS measurement correction. From their results, they conclude that sea level rise in the GOT is significantly faster than the global rate and it is in the range of 3.0–5.5 mm yr\(^{-1}\). Instead of falling sea level at Ko Lak as found in Yanagi and Akaki [1994] and Vongvisessomjai [2006, 2010], they found it a rising at the rate of 3.6 ± 0.7 mm yr\(^{-1}\) (absolute rate). Because of these conflicting and confusing results and the importance of sea level rise on the region, analysis of sea level is done here.

In this section, only 12 stations shown in Section 3.1 (with 15 year-long recorded) are analyzed to estimate the rate of relative sea level rise (including vertical land movement). Rate of sea level rise at each station will be estimated based on monthly mean sea level data. Hence, hourly data obtained from Marine Department and UHSLC are averaged to have a monthly values. If the monthly data from PSMSL's station has a longer period than that of Marine Department and UHSLC, the data from PSMSL's station will be used instead.

Rate of sea level rise at each station will be calculated based on 1) a least square linear fit of monthly mean sea level data and 2) an averaged slope calculated from the low frequency oscillations using last one or two modes of Empirical Mode Decomposition and Hilbert–Huang Transform (EMD/HHT) of monthly mean sea level [Huang et al., 1998]. EMD/HHT method can be used to separate oscillatory modes from trends of any nonstationary and nonlinear time series. Detail of this method can be seen in [Huang et al., 1998]. EMD/HHT for sea level data has been applied at the Chesapeake Bay and surrounded area to estimate trend and acceleration [Ezer and Corlett, 2012]. When compared with linear trends, the EMD trend can show if sea level rise rate changes in time, i.e., if there is acceleration or deceleration of sea level rise.

To deal with errors from each tide gauge data, the following steps have been performed [Jevrejeva et al., 2006]. First, the error of mean sea level caused by inverted
barometer is neglected because it has a small impact on the changes sea level and not all stations provide such data. Second, Glacial Isostatic Adjustment (GIA) corrections of Peltier [2004] (ICE-5G(VM2) updated 2012b) are taken into consideration. Unfortunately, not all tide gauge stations have GIA corrections. Therefore, the value from nearest station (within 100 km) will be applied to those stations that have no GIA correction. If the distance to the nearest station is greater than 100 km but less than 1,000 km, interpolation will be used. Note, however, that significant vertical land movement due to water extractions, earthquakes and other processes exist in the GOT in addition to the GIA effects.

Figure 7 presents monthly mean sea level variability (only 2 stations are shown (a) Tha Chin: a representative of the GOT station and (b) Ko Taphoa Noi: a representative of Andaman station). From Figure 7, the rate of sea level at Tha Chin is clearly greater than at Ko Taphoa Noi (scale on $y-$axis is equal for both plots). Visually, linear trends are approximately 20 and 3 mm yr$^{-1}$ at Tha Chin and Ko Taphoa Noi, respectively. The rate at Tha Chin station is 7 times larger than the global rate of sea level rise. This number is a relative rate (without correction). If it had been corrected by GIA value, the rate would not change much compare with global mean since the global averaged rate of sea level rise due to GIA is approximately $-0.3$ mm yr$^{-1}$ [Peltier, 2001; Peltier and Lutheke, 2009]. The rate determined from HHT trend at both stations agrees with the linear trend, but the HHT also shows sea level acceleration. The actual values of rate of sea level rise at all 12 tide gauge stations are shown in Figure 8 and Table 3.
Figure 7. Monthly mean sea level (gray dot) at (a) Tha Chin and (b) Ko Taphoa Noi stations. Black solid line and black dot denote a linear trend and the last mode of HHT (trend) approach.

From Figure 8, rate of sea level rise in the GOT and Andaman Sea is larger than the global rate and in some stations significantly larger. Only one station (Si Chang) has a rate less than the global rate for both estimated from linear and HHT trend. However, rate of sea level rise estimated from HHT trend at Bang Pakong and Ko Lak is also less than the global rate. In some stations, rate between linear and HHT trend is quite difference (more than 50%), for example, Bang Pakong, Ban Lam, Ko Lak, etc. This might be explained by how HHT trend at each station behaves. Consider HHT trend at Ko Taphoa Noi in Figure 7. It is found that the HHT trend is gradually increased and reached the peak (approximately 15 cm greater than from
linear trend) at the end of time series. Therefore, it is expected to have a higher rate than the linear trend. Unlike at Tha Chin, both linear and HHT trend are almost parallel. The rate of sea level rise at both estimates is expected to be almost the same, with rate from HHT trend being a bit smaller. Even with GIA correction applied, rate of sea level rise found at Phrachulachomkloa, Tha Chin, Mae Klong, and Ban Lam are still higher than the global rate (more than 5 times larger). What is the cause?

**Figure 8.** Rate of sea level rise ±95% confidence level in mm yr$^{-1}$ based on the linear regression of monthly sea level data (dark gray) and slope of HHT trend’s slope (light gray). Dashed line denotes global mean rate of sea level rise which is 1.7±0.3 mm yr$^{-1}$ [Church and White, 2006].
One explanation is that the northern part of the UGOT is sinking due to groundwater withdrawal [Poland, 1984; Therakomen, 2001; Phien-wej et al., 2006; Aobpaet et al., 2009]. Groundwater withdrawal or mining is an anthropogenic change in sea level. Globally, it can change sea level approximately 0.1–0.3 mm yr$^{-1}$ [Gornitz, 2001]. Groundwater has been pumping out in Bangkok and the surrounded area and it has caused land subsidence for almost 40 years. A sinking rate larger than 120 mm yr$^{-1}$ is found in the 1980s at central Bangkok, but it reduced to ~10 mm yr$^{-1}$ in the 2000s (Figure 7. in Phien-wej et al. [2006]).

In the recent study by Aobpaet et al. [2009], interferometric synthetic aperture radar (InSAR), a SAR technique to detect the land movement, found that the eastern central Bangkok area still has a sinking rate of 15 mm yr$^{-1}$. Since Phrachulachomkloa, Tha Chin, and Mae Klong tide gauge stations are situated in the area where land subsidence is still a problem, faster rate of sea level rise is expected at those stations.

Even after exclusion of tide gauge stations in the area of land subsidence and the correction using precise GPS technique, the rate of sea level rise in the GOT is still faster compared to the global rate [Trisirisatayawong et al., 2011]. In the study of Trisirisatayawong et al. [2011], they mention pre– and post–2004 Indian Ocean Tsunami sea level records and vertical uplift or emergence caused by 2004 Sumatra–Andaman earthquake. Based on continuous GPS measurements, after 2004 Sumatra–Andaman earthquake many parts of Thailand are sinking at rates up to 10 mm yr$^{-1}$. The projection of land subsidence for the next two decades in Bangkok area should not exceed 5 mm yr$^{-1}$ [Satirapod et al., 2013]. Note that the areas that have sinking rate as mentioned above are approximately 650–1,500 km away from the epicenter of the earthquake causing the tsunami.
Table 3. Rate of relative sea level rise (RSLR) and 95% confidence level at each tide gauge station in the GOT and Andaman Sea. RSLR_{Lin} and RSLR_{HHT} are rate of sea level rise obtained by a linear fit and the slope of the last mode of HHT sea level, respectively. Asterisks denote data obtained from PSMSL. At each station, GIA correction values are used.

<table>
<thead>
<tr>
<th>Station</th>
<th>RSLR_{Lin}</th>
<th>RSLR_{HHT}</th>
<th>GIA</th>
<th>Years</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rayong</td>
<td>3.19 ± 2.15</td>
<td>1.93 ± 0.41</td>
<td>-0.38</td>
<td>1986–2012</td>
</tr>
<tr>
<td>Si Chang*</td>
<td>0.86 ± 0.55</td>
<td>0.84 ± 0.02</td>
<td>-0.38</td>
<td>1940–2002</td>
</tr>
<tr>
<td>Bang Pakong</td>
<td>5.78 ± 1.26</td>
<td>0.18 ± 0.69</td>
<td>-0.38</td>
<td>1981–2012</td>
</tr>
<tr>
<td>Phrachulachomkloa*</td>
<td>15.10 ± 0.45</td>
<td>13.35 ± 0.28</td>
<td>-0.39</td>
<td>1940–2012</td>
</tr>
<tr>
<td>Tha Chin</td>
<td>19.80 ± 1.42</td>
<td>18.32 ± 0.46</td>
<td>-0.39</td>
<td>1977–2012</td>
</tr>
<tr>
<td>Mae Klong</td>
<td>15.53 ± 1.59</td>
<td>17.21 ± 0.36</td>
<td>-0.39</td>
<td>1980–2012</td>
</tr>
<tr>
<td>Ban Lam</td>
<td>7.74 ± 5.01</td>
<td>14.56 ± 2.19</td>
<td>-0.39</td>
<td>1997–2012</td>
</tr>
<tr>
<td>Ko Lak</td>
<td>4.79 ± 2.10</td>
<td>4.84 ± 0.84</td>
<td>-0.36</td>
<td>1985–2013</td>
</tr>
<tr>
<td>Ko Lak*</td>
<td>0.54 ± 0.52</td>
<td>0.38 ± 0.04</td>
<td>-0.36</td>
<td>1940–2012</td>
</tr>
<tr>
<td>Ko Mattaphon*</td>
<td>6.00 ± 4.11</td>
<td>6.06 ± 0.60</td>
<td>-0.34</td>
<td>1992–2012</td>
</tr>
<tr>
<td>Geting</td>
<td>1.92 ± 3.82</td>
<td>2.17 ± 0.63</td>
<td>-0.33</td>
<td>1987–2006</td>
</tr>
<tr>
<td>Ko Taphoa Noi</td>
<td>4.45 ± 1.43</td>
<td>7.11 ± 0.77</td>
<td>-0.28</td>
<td>1985–2013</td>
</tr>
<tr>
<td>Ko Taphoa Noi*</td>
<td>1.24 ± 0.44</td>
<td>1.90 ± 0.17</td>
<td>-0.28</td>
<td>1940–2012</td>
</tr>
<tr>
<td>Langkawi</td>
<td>2.53 ± 1.42</td>
<td>4.67 ± 0.25</td>
<td>-0.36</td>
<td>1985–2013</td>
</tr>
</tbody>
</table>
The vertical uplifts shown before 2004 Sumatra-Andaman earthquake (between 1994 and 2004) at GPS stations near Si Chang and Ko Mattaphon tide gauge stations are $2.2 \pm 0.8$ and $3.8 \pm 1.3$ mm yr$^{-1}$, respectively. However after the earthquake (between 2005 and 2009), the land is submerging at a rate of $-12.7 \pm 4.2$ and $-3.9 \pm 2.1$ mm yr$^{-1}$ at GPS stations near Si Chang and Ko Mattaphon, respectively [Trisirisatayawong et al., 2011]. This means that not only the rate of sea level rise in Thailand has to be corrected by GIA corrections but also by the vertical movement of the land due to seismic activity. Due to the lack of GPS stations data near other tide gauge stations, the pre- and post-2004 Sumatra-Andaman earthquake rate of sea level rise in the GOT and Andaman Sea will be shown without any corrections and the results are presented in Figure 9 and 10. These figures show how the rate of sea level rise changed before and after the earthquake.

Figure 9 and 10 show that the relative sea level rise before the earthquake is gradually increasing in most stations, except at Ban Lam and Ko Lak stations (Figure 9e and 10a) where the relative sea level is falling. However, the relative rate of sea level rise at stations situated in the severe land subsidence zone remains high (i.e., Tha Chin and Mae Klong). Clearly at Tha Chin and Mae Klong stations (Figure 9c and 9d), land subsidence affects the rate of sea level rise. At both stations after the earthquake, the rates stay about the same as the pre-earthquake rates (in addition of land subsidence, land is actually sinking due to seismic activity [Trisirisatayawong et al., 2011; Satirapod et al., 2013]). More interesting, the rate of relative sea level rise at Bang Pakong (Figure 9b) is not so high, even though it has been affected by land subsidence due to groundwater withdrawal. In addition, sea level data between 2000 and 2004 seems to positively deviate from the group, but rate is still low. At Ban Lam tide gauge station (Figure 9e), when the pre- and post-earthquake are included, the analysis now has sea level falling at the rate of $-3.83 \pm 14.68$ mm yr$^{-1}$. This may or may not be the correct relative rate, because the time series data covers
only a short period (~7 years) and it may not be affected by land subsidence that appears at Tha Chin and Mae Klong stations (the distance from Mae Klong and Ban Lam is approximately 20 km). Note that the rate shown in Figure 9 and 10 (before and after the earthquake) are not corrected by the GIA, which is \(-0.3 \text{ mm yr}^{-1}\) (approximated from Table 3 for the whole region) and vertical land movement, which are 3 and \(-8 \text{ mm yr}^{-1}\) for the pre- and post-earthquake, respectively (the numbers are approximated from only 2 GPS stations shown in Figure 4 of Trisirisatayawong et al. [2011]).
Figure 9. Monthly sea level at each tide gauge station. Black and gray dots represent pre- and post-2004 Sumatra-Andaman earthquake, respectively with a linear trend shown on top of it. The number shows the rate of relative sea level rise ±95% confidence level in mm yr⁻¹. The stations situated in the UGOT where the land subsidence is still active are b) Bang Pakong c) Tha Chin and d) Mae Klong. Rayong station is on the eastern side of the GOT.
Figure 10. Monthly sea level at each tide gauge station. Black and gray dots represent pre- and post-2004 Sumatra-Andaman earthquake, respectively with a linear trend shown on top of it. The number shows the rate of relative sea level rise ±95% confidence level in mm yr⁻¹. Station (a) Ko Lak, (b) Ko Mattaphon, and (c) Geting are located in the GOT (western side), while (d) Ko Taphoa Noi and (e) Langkawi are on the Andaman Sea.
Farther south from the area of land subsidence (in the UGOT), the rate of relative sea level rise is small compared with those found in the UGOT (see Figure 10). This emphasizes the large land subsidence found in the UGOT, especially near the Bangkok area. One interesting feature shown in Figure 9 and 10 is that the relative rates of sea level rise after the earthquake at all stations are rising regardless of distance from 2004 Sumatra-Andaman earthquake (seismic activity). This is again confirms that the land is sinking due to seismic activity. Outside the zone of land subsidence due to groundwater withdrawal, after the 2004 Sumatra-Andaman earthquake, the relative rate of sea level rise is in the range of 19 to 30 mm yr\(^{-1}\). The highest rate is found at Taphoa Noi (~30 mm yr\(^{-1}\)), which is the station that is close to the epicenter. Similar rates are also found at Rayong, Bang Pakong, and Tha Chin stations.

Of the more than 20 tide gauge stations, fewer have reliable data long enough for calculating sea level rise. It has been shown here that in addition to global sea level rise, several local factors may impact the sea level signal in the GOT, including anthropogenic sea level change from groundwater withdrawal and seismic activity (vertical land movement). In the next section, the seasonal sea level cycle will be discussed.

### 3.3.2 Seasonal sea level cycle

In sea level time series, the seasonal cycle is accounted for as one of the largest components of sea level variability \cite{Tsimplis and Woodworth, 1994; Torres and Tsimplis, 2012}. Seasonal sea level cycle is mainly induced by oceanographic and meteorological forcings (i.e., changes in air pressure, winds and heating steric effects, etc.) \cite{Gill and Niiler, 1973}. Water runoff and geological contribution, can also add to changing the seasonal sea level cycle at stations near rivers \cite{Tsimplis and Woodworth, 1994}. The global seasonal sea level cycle analysis is investigated by using
the observed mean sea level data obtained from PSMSL data (see Woodworth and Player [2003] and Holgate et al. [2013]) [Pattullo et al., 1955; Tsimplis and Woodworth, 1994]. Mean sea level based on a general ocean circulation model is also used to study the seasonal sea level cycle [Vinogradov et al., 2008], and the combination of mean sea level obtained from tide gauge stations and altimetry data can also be used [Torres and Tsimplis, 2012, 2013].

In Thailand, previous sea level research has very little information about seasonal sea level variability; therefore, it will be discussed in this section (later, modeling seasonal variations will be discussed as well). Note that ~3 tide gauge stations were included in the study of Tsimplis and Woodworth [1994], but in this study 14 tide gauge stations situated in the GOT and Andaman Sea will be analyzed (see Table 4). Seasonal sea level changes because of changing in gravitational potential. These changes are caused by two long-period tidal harmonic components which are annual (Sa) and semi-annual (Ssa) components. Sa accounts for the changing distance between the sun and the earth, while Ssa accounts for a changing solar declination [Torres and Tsimplis, 2012]. These two components can be estimated from harmonic analysis as explained in Section 3.2, but instead of more than 8 tidal components. For more convenience, the linear regression least-square fit methods will be used to fit mean monthly sea level anomaly of month $i$ ($\bar{M}_i$) equation shown in Tsimplis and Woodworth [1994]. It is written as follows

$$\bar{M}_i = A_{Sa} \cos \left[ \frac{\pi}{6} (t - \phi_{Sa}) \right] + A_{Ssa} \cos \left[ \frac{\pi}{3} (t - \phi_{Ssa}) \right].$$

(3)

where $A_{Sa}$ and $A_{Ssa}$ are amplitude of annual and semi-annual components, respectively, while $\phi_{Sa}$ and $\phi_{Ssa}$ are phase lags corresponding to the peak of two signals. They are in the range of $-6$ to $6$ and $-3$ to $3$ for $Sa$ and $Ssa$ components, respectively, and $t$ is time at the middle of each month $i$ ($t = i - 0.5$). $\bar{M}_i$ is estimated from
averaging monthly sea level anomaly, $M_{ik}$ (a deviation of month $i$ from annual mean of year $k$), over $N_{yr}$. Therefore mean monthly sea level anomaly can be expressed as

$$\bar{M}_i = \frac{1}{N_{yr}} \sum_{k=1}^{N_{yr}} M_{ik}.$$ 

As mentioned in Tsimlips and Woodworth [1994], a 5 year–long segments can provide stable amplitudes and phase lags because it minimizes the large variability of annual and semi–annual calculated from each year data. Therefore the harmonic analysis of mean monthly sea level anomaly will be based on 5 year averages of the time series.

Figure 11 compares sea level anomaly at Tha Chin (station in the GOT) and Ko Taphoa Noi (station in Andaman Sea). A black thick line is the mean monthly sea level anomaly representing all monthly sea level anomalies (gray thin lines). The sea level cycle at these stations are totally different with opposite phase, although they are influenced by similar meteorological conditions, monsoonal wind effects and air pressure, for instance. However, this difference means sea level cycle at both stations are affected by different forcings, a discussion will be mentioned later. This is true for the remaining stations (figure not shown).

In the GOT annual sea level is an important cycle, unlike in Andaman Sea where both annual and semi–annual sea level cycles are both important. The result of harmonic analysis of mean monthly sea level anomalies for all 14 tide gauge stations is presented in Table 4. $A_{Sa}$, $A_{Ssa}$, $\phi_{Sa}$, and $\phi_{Ssa}$ are shown together with ±95% confidence level. Because of shorten in time series, Sattahip and Hua Hin stations present the actual value obtained from harmonic analysis, not the mean values of each 5–year segment as shown at other stations (both Sattahip and Hua Hin stations have time series less than 10 years).
From Table 4, the annual amplitudes can be grouped into 3 groups which are upper and eastern GOT, central GOT (Ko Lak to Geting), and Andaman Sea. In the upper and eastern GOT, the amplitude of $S_a$ is in the range of 120 to 180 mm. In the central GOT area, annual amplitude is found to be greater than 200 mm but less than 240 mm. These 2 groups are the same as found by Tsimlpis and Woodworth [1994] which mentioned the amplitude of annual sea level in the GOT and Eastern Malaysian coast as being larger than 120 mm. The annual amplitudes found in the UGOT and central GOT decrease toward the north. In Andaman Sea, the annual amplitude is quite small compared with the one found in the GOT and Eastern Malaysian coast. The annual amplitude is approximately 100 mm at both Ko Taphoa Noi and Langkawi stations. Note that this value is approximately half that found at neighboring stations, Ko Mattaphon and Geting, but situated on the other side of Andaman Sea.
Figure 11. Sea level anomaly at (a) Tha Chin (in the GOT) and (b) Ko Taphoa Noi (Andaman Sea) tide gauge stations. A black thick line represents mean monthly sea level anomaly, while gray thin lines represent variation of sea level anomaly for each year before averaging to obtain the black line.
Table 4. Amplitudes ($A$) in mm and phase lags ($\phi$) in months of the maximum sea level from January ±95% confidence level of annual ($S_a$) and semi-annual ($S_{sa}$) sea level obtained from harmonic analysis at 14 tide gauge stations. Asterisks denote data obtained from PSMSL, while sharps denote stations that have only one 5-year segment. Note that the number shown is averaged from each 5-year segment. The last two stations are in the Andaman Sea.

<table>
<thead>
<tr>
<th>Station</th>
<th>$A_{S_a}$ (mm)</th>
<th>$A_{S_{sa}}$ (mm)</th>
<th>$\phi_{S_a}$ (mon)</th>
<th>$\phi_{S_{sa}}$ (mon)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rayong</td>
<td>170.06 ± 49.99</td>
<td>34.74 ± 18.09</td>
<td>-0.08 ± 0.39</td>
<td>-1.59 ± 0.76</td>
</tr>
<tr>
<td>Sattahip*</td>
<td>177.88</td>
<td>15.83</td>
<td>0.39</td>
<td>-1.90</td>
</tr>
<tr>
<td>Si Chang*</td>
<td>174.18 ± 12.09</td>
<td>28.28 ± 7.84</td>
<td>0.29 ± 0.11</td>
<td>-0.86 ± 1.35</td>
</tr>
<tr>
<td>Bang Pakong</td>
<td>122.80 ± 11.30</td>
<td>20.38 ± 7.60</td>
<td>-0.02 ± 0.17</td>
<td>-2.75 ± 0.41</td>
</tr>
<tr>
<td>Phrachulachomkloa*</td>
<td>145.86 ± 8.36</td>
<td>27.08 ± 5.17</td>
<td>0.23 ± 0.06</td>
<td>-1.83 ± 0.81</td>
</tr>
<tr>
<td>Tha Chin</td>
<td>164.42 ± 33.95</td>
<td>33.05 ± 12.57</td>
<td>0.18 ± 0.12</td>
<td>-1.82 ± 0.41</td>
</tr>
<tr>
<td>Mae Klong</td>
<td>166.39 ± 20.12</td>
<td>20.90 ± 14.67</td>
<td>0.38 ± 0.40</td>
<td>-1.11 ± 1.44</td>
</tr>
<tr>
<td>Ban Lam</td>
<td>167.26 ± 34.81</td>
<td>39.61 ± 37.77</td>
<td>0.17 ± 0.59</td>
<td>-1.38 ± 1.75</td>
</tr>
<tr>
<td>Hua Hin*</td>
<td>207.96</td>
<td>32.27</td>
<td>0.03</td>
<td>-1.96</td>
</tr>
<tr>
<td>Ko Lak*</td>
<td>215.61 ± 9.98</td>
<td>38.87 ± 5.66</td>
<td>0.17 ± 0.05</td>
<td>-1.73 ± 0.16</td>
</tr>
<tr>
<td>Ko Mattaphon*</td>
<td>239.77 ± 31.03</td>
<td>35.51 ± 12.10</td>
<td>0.12 ± 0.26</td>
<td>-1.35 ± 0.44</td>
</tr>
<tr>
<td>Geting</td>
<td>225.80 ± 13.00</td>
<td>51.51 ± 16.51</td>
<td>-0.03 ± 0.08</td>
<td>-1.00 ± 0.21</td>
</tr>
<tr>
<td>Ko Taphoa Noi*</td>
<td>97.68 ± 14.05</td>
<td>64.06 ± 7.69</td>
<td>-4.58 ± 0.23</td>
<td>-1.38 ± 0.17</td>
</tr>
<tr>
<td>Langkawi</td>
<td>97.08 ± 20.31</td>
<td>62.00 ± 9.27</td>
<td>-4.76 ± 0.13</td>
<td>-1.49 ± 0.18</td>
</tr>
</tbody>
</table>
Like the annual amplitude, the annual phase lag in the GOT is clearly different from the Andaman Sea. This result agrees with Tsimplis and Woodworth [1994]. In the GOT, the peak of the annual phase lag is approximately near the beginning of the year (mid-December to mid-January). The peak of the annual phase lag in Andaman Sea is approximately four and a half months earlier (~mid-August).

Tsimplis and Woodworth [1994] stated that generally the magnitude of annual amplitude is much larger than that of semi-annual. In this study, the semi-annual amplitude is much smaller than annual amplitude as expected, except for the two stations in Andaman Sea. In the GOT, the semi-annual amplitude is in the range of 20 to 40 mm, except at Geting where it is 51 mm. The values found in the GOT are approximately 4 times smaller than the annual and even smaller than the semi-annual amplitude found in Andaman Sea where the amplitude is about 60 mm. It is interesting to see the semi-annual amplitude in Andaman Sea is not quite different from the annual one.

The semi-annual phase in the GOT increases toward north. At Geting, its maximum peak occurs one month earlier. The peak is almost 3 months earlier at Bang Pakong station, where both annual and semi-annual amplitudes are small compared to other stations. In general, the semi-annual phase lag in the GOT is in the range of 1 to 3 months earlier. In Andaman Sea, the peak is one and a half months earlier at both Ko Taphoa Noi and Langkawi stations.

Figure 12 compares the monthly sea level anomaly with monthly sea level pressure anomaly at Tha Chin (GOT) and Ko Taphoa Noi (Andaman Sea). Sea level pressure data is retrieved from http://iridl.ldeo.columbia.edu/. It is based on Climate Data Assimilation System 1; NCEP-NCAR Reanalysis Project [Kalnay et al., 1996]. The monthly sea level pressure anomaly and monthly sea level anomaly are in millibar and decimeter, respectively (for visualization convenience). Clearly sea level anomaly and sea level pressure anomaly at Tha Chin are both in phase (Figure 12a; correlation
R = 0.69 at 95% confidence interval), while they are out of phase at Ko Taphoa Noi (Figure 12b; R = −0.45). Note that sea level pressure anomaly at both places is nearly the same. They are positively correlated with Multivariate ENSO Index (MEI) with R values approximately 0.50 (Figure not shown). The results suggest that seasonal sea level cycle in the GOT is mainly controlled by meteorological forcing, in particular the seasonal wind and sea level pressure associated with the monsoon. River discharge and thermal water expansion may have less impact in this region, since the maximum peak of seasonal sea level cycle in the GOT appears at the beginning of the year while the peak of sea surface temperature and water discharge are in April and in October (major) and June (minor), respectively [Singhrattina et al., 2005b]. The seasonal sea level pattern seen in Figure 11 is consistent with monsoon–driven influence. During the southwest monsoon, the sea level is piled up against the coast in the Andaman Sea (Ko Taphoa Noi and Langkawi) while water is pushed away from the coast in the lower GOT. The seasonal sea level pattern in the UGOT is similar to the GOT cycle, indicating that it is a large–scale pattern, not a local phenomena. The detailed seasonal variations in the UGOT will be discussed later using the numerical model results.
Figure 12. Comparison between mean sea level anomaly (gray dot line) and sea level pressure (black dot line) at (a) Tha Chin and (b) Ko Taphoa Noi tide gauge stations. Unit of mean sea level (MSL) and sea level pressure (SLP) are decimeter and millibar, respectively.
CHAPTER 4

NUMERICAL MODEL OF THE UPPER GULF OF THAILAND

As mentioned in Chapter 2, the UGOT is relatively shallow and the circulation is partially controlled by tides at the southern boundary where the UGOT and the whole GOT are connected. The seasonal circulation pattern is largely controlled by the monsoonal winds. The aim of this part of the study is to develop a three-dimensional numerical model of the UGOT, forced by astronomical tides at the southern boundary. Then other forces that affect the dynamics of the UGOT, such as winds, surface heat fluxes and river discharges, will be discussed in following chapter.

4.1 MODEL SETUP

A numerical model of the UGOT is developed using the Princeton Ocean Model (POM). The POM is a three-dimensional primitive equation ocean model that includes a completely implemented thermodynamics and the level 2.5 Mellor-Yamada turbulence closure scheme [Mellor and Yamada, 1982]. It is a free surface model with a time splitting integration scheme: an external mode with a short time step for barotropic (two-dimensional) waves and an internal mode with a longer time step for baroclinic (three-dimensional) flows [Mellor, 2004]. A curvilinear grid that is almost rectangular with a resolution of approximately 1 km and bottom topography based on a nautical chart (Figure 13) was used. The number of the horizontal grid points in $x$– and $y$–directions is $111 \times 96 = 10,656$. In the vertical, a terrain-following, sigma-coordinate grid is used with 21 layers. Sigma is defined as $\sigma = (z - \eta)/(z + H)$,
so $\sigma = 0$ at the free surface $z = \eta(x, y, t)$ and $\sigma = -1$ at the bottom $z = -H(x, y)$; see Mellor [2004] for details model equations in sigma coordinates. A minimum water depth in the model domain was set to be 3 m. The time steps used in this study were 6 s and 60 s for the external and internal modes, respectively. At the open boundary (blue solid line in Figure 13), a radiation boundary condition was applied. Eight tidal components, Q1, O1, P1, K1, N2, M2, S2, and K2, obtained from harmonic analysis of observed water elevations at Hua Hin and Sattahip tide gauge stations were linearly interpolated across the open boundary (see previous chapter for more information on the tide gauges and sea level in the GOT). Note that tidal amplitudes and tidal phases of all eight tidal components at open boundary stations are shown in Table 5. Initial conditions based on observations [Buranapratheprat et al., 2008a]. The water salinity was assigned as a constant (32.15) for the entire model domain, while water temperature was nearly constant. It varied from 30.63°C at the surface to 30.02 °C at the bottom.
Figure 13. Model grid and bottom topography of the UGOT model. A blue solid line depicts open boundary forced by astronomical tides. Blue and green dots represent tide gauge and current velocity measurement stations, respectively. Note that a colorbar has a scale for bottom topography in m.
Table 5. The eight tidal components obtained from harmonic analysis of observed water levels at Hua Hin and Sattahip tide gauge stations.

<table>
<thead>
<tr>
<th>Tidal Components</th>
<th>Hua Hin Station</th>
<th>Sattahip Station</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q1</td>
<td>8.085</td>
<td>7.531</td>
</tr>
<tr>
<td>O1</td>
<td>39.950</td>
<td>38.324</td>
</tr>
<tr>
<td>P1</td>
<td>18.219</td>
<td>17.402</td>
</tr>
<tr>
<td>K1</td>
<td>60.973</td>
<td>58.176</td>
</tr>
<tr>
<td>N2</td>
<td>5.944</td>
<td>4.730</td>
</tr>
<tr>
<td>M2</td>
<td>33.734</td>
<td>26.880</td>
</tr>
<tr>
<td>S2</td>
<td>15.578</td>
<td>12.554</td>
</tr>
<tr>
<td>K2</td>
<td>5.939</td>
<td>5.062</td>
</tr>
</tbody>
</table>

From the above configuration, a model was run for two months between February and March 2000 (experiment Tide 1), but only March 2000 was used for the analysis of tides (longer simulations for seasonal and interannual variability studies are described in later chapters). Root-mean square error (RMSE) and correlation coefficient squared ($R^2$) were used to evaluate the performance of the model (see equations (1) and (2)).
4.2 TIDAL MODEL CALIBRATION AND VERIFICATION

In this section the calibration process of the UGOT model is described. The predicted surface elevations of March 2000 at each tide gauge stations obtained from experiment Tide.1 were plotted along with the observed surface elevations (see Figure 14, only Phrachulachomkloa station is shown). From the plot of all stations, the observed maximum tidal range during spring and neap tides for all stations were approximately 2.5 and 1.5 m, respectively, while from experiment Tide.1 maximum tidal range were approximately 1.9 and 0.9 m during spring and neap tide, respectively. These results clearly show that the model under-estimates the tidal range (by ~60 cm or ~25% difference; statistics of water surface elevation error is shown in Table 6). For experiment Tide.1, Bangkok Bar had a larger error compared with other stations and its $R^2$ was also smaller than others. From Figure 14a, surface elevation obtained from experiment Tide.1 (blue solid line) at Phrachulachomkloa station captures the observed (red asterisks) surface elevation pattern well, except the high and low tides during spring tide. To better examine the model performance, a harmonic analysis of observed and predicted surface elevation from experiment Tide.1 was performed and the 8 tidal components obtained from this performance are shown in Figure 14b and 14c for tidal amplitude and tidal phase, respectively. Clearly, the under-estimated surface elevation in experiment Tide.1 was mainly caused by an under-estimated tidal amplitude of the semi-diurnal components (see Figure 14b). The model did a good job in reproducing the observed tidal phase for all tidal components (Figure 14c). The RMSE of all 8 tidal components is estimated (Table 7). Clearly the model did not do well job in capturing tidal component of semi-diurnal tides. This deficiency is attributed to the fact that only 2 tidal stations are available near the open boundaries, so further calibration of the open boundary forcing was needed.
Figure 14. Surface elevation comparison between observed (red asterisks) and predicted blue (experiment Tide_1, before calibration) and black (experiment Tide_2, after calibration) solid lines at Phrachulachomkloa tide gauge station (a). Tidal components compared between observed (black) and predicted of experiment Tide_1 and experiment Tide_2 for tidal amplitude (b) and for tidal phase (c), respectively.

Based on the results shown in Figure 14 and Table 6 and 7, an empirical adjustment of tidal components (especially for the semi diurnal components) at the open boundary were done based on trial and error method and the results of this experiment, experiment Tide_2, are also shown in Figure 14 and Table 6 and 7. After adjusting tidal components at the open boundary, the predicted surface elevation from all tide gauge stations gave a good agreement with the observed surface elevation (see Figure 14a red asterisks and black solid line).

From Figure 14a, the surface elevation obtained from experiment Tide_2 could reproduce high and low tides during spring tide as seen from the observed surface elevation with the improvement of $R^2$ and RMSE values. The RMSE of surface
elevation obtained from experiment Tide.2 was in the range of 10 to 20 cm instead of 25 to 35 cm as obtained from experiment Tide.1. The value of $R^2$ was greater than 0.90 for all stations (see Table 6).

The calibration (experiment Tide.2) greatly improved the simulations (see Figure 14b and 14c), and the semi-diurnal components from experiments Tide.2 now agree well with observed data for both, tidal amplitude and tidal phase. From Table 7, RMSE of tidal amplitude and tidal phase for experiment Tide.2 showed an improvement compared with experiment Tide.1. Approximately 50% improvement has been seen from both tidal amplitude and tidal phase, except tidal phase of N2 tide. Overall the UGOT model has performed a good job in capturing the surface elevation. Evaluation of the simulated tidal currents is done next.

**Table 6.** Root-mean squared error (RMSE) and $R^2$ of surface elevation at all tide gauge stations before (experiment Tide.1) and after adjusting tidal components (experiment Tide.2) at open boundary.

| Station                | Tide.1 | | Tide.2 | |
|------------------------|--------|--------|--------|
|                        | RMSE (cm) | $R^2$ | RMSE (cm) | $R^2$ |
| Mae Klong              | 29.20  | 0.81   | 15.65  | 0.95 |
| Tha Chin               | 30.45  | 0.80   | 16.91  | 0.94 |
| Phrachulachomkloa      | 31.90  | 0.79   | 20.22  | 0.92 |
| Bangkok Bar            | 34.85  | 0.75   | 22.14  | 0.90 |
| Bang Pakong            | 26.95  | 0.84   | 12.10  | 0.97 |
Table 7. Root-mean squared error of tidal amplitude (Amp) and tidal phase (Pha) for experiment Tide.1 and Tide.2 of all tide gauge stations.

<table>
<thead>
<tr>
<th></th>
<th>Q1</th>
<th>O1</th>
<th>K1</th>
<th>P1</th>
<th>N2</th>
<th>M2</th>
<th>S2</th>
<th>K2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amp Tide.1 (cm)</td>
<td>1.14</td>
<td>5.04</td>
<td>8.64</td>
<td>2.19</td>
<td>8.69</td>
<td>26.52</td>
<td>14.85</td>
<td>4.53</td>
</tr>
<tr>
<td>Amp Tide.2 (cm)</td>
<td>0.78</td>
<td>3.56</td>
<td>4.38</td>
<td>1.44</td>
<td>1.75</td>
<td>1.71</td>
<td>1.04</td>
<td>0.30</td>
</tr>
<tr>
<td>Pha Tide.1 (degree)</td>
<td>14.92</td>
<td>10.00</td>
<td>11.74</td>
<td>17.42</td>
<td>22.71</td>
<td>15.08</td>
<td>21.90</td>
<td>11.95</td>
</tr>
<tr>
<td>Pha Tide.2 (degree)</td>
<td>8.94</td>
<td>6.03</td>
<td>4.13</td>
<td>4.13</td>
<td>19.71</td>
<td>7.83</td>
<td>8.72</td>
<td>8.72</td>
</tr>
</tbody>
</table>

Since tidal currents in the UGOT are mainly in north–south direction, only $v$—velocity component of the observed and predicted currents are compared. Note that $u$—component for both stations is in the range of $-5$ to $5$ cm s$^{-1}$, while $v$—component are in the range of $-25$ to $25$ cm s$^{-1}$ and $-40$ to $40$ cm s$^{-1}$ for Si Chang and Petchburi buoy stations, respectively. Velocity $v$—components obtained from Si Chang and Petchburi buoy stations are shown in Figure 15a and 15b, respectively. From Figure 15, it was found that $v$—velocity obtained from experiment Tide.2 gave a good agreement compared with the observed data with $R^2 \sim 0.90$ for both stations and RMSE approximately $3.75$ and $5.32$ cm s$^{-1}$ at Si Chang and Petchburi buoy stations, respectively.

Like surface elevation, harmonic analysis of the $v$—velocity component at both stations has been performed. From the harmonic analysis, it was found that the two major tidal components of $v$—velocity were M2 and K1 for both stations (Figure not shown). Unlike surface elevation, the amplitude of M2 obtained from harmonic analysis of $v$—velocity is larger than that of K1. Tidal velocity amplitude comparison between observed and predicted data shows that the model has done a good job for
all tidal components, except M2 (both stations) and other semi-diurnal components at Petchburi station. For the tidal velocity phase comparison, Q1 has the largest error at both stations while all other diurnal components at Si Chang station show some discrepancies with observations. Simulations of the N2 and M2 components have considerable errors. Table 8 summarizes the results and shows the large tidal amplitude error for M2 and tidal phase error for Q1 and N2.

![Map showing buoy locations](image)

**Figure 15.** Velocity $v$–component comparisons (in cm s$^{-1}$) between observed (asterisks) and predicted (solid line; experiment Tide.2) at Si Chang (a) and Petchburi (b) buoy stations (circles on the map). On bottom right corner of each plot, $R^2$ and RMSE are presented.
Table 8. Root-mean squared error of \( v \)–velocity amplitude (Amp) and phase (Pha) for experiment Tide_2 of two buoy stations.

<table>
<thead>
<tr>
<th></th>
<th>Q1</th>
<th>O1</th>
<th>K1</th>
<th>P1</th>
<th>N2</th>
<th>M2</th>
<th>S2</th>
<th>K2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amp Tide_2 (cm s(^{-1}))</td>
<td>0.64</td>
<td>0.55</td>
<td>0.35</td>
<td>0.12</td>
<td>0.40</td>
<td>2.86</td>
<td>1.39</td>
<td>0.40</td>
</tr>
<tr>
<td>Pha Tide_2 (degree)</td>
<td>46.02</td>
<td>4.57</td>
<td>6.45</td>
<td>6.45</td>
<td>10.00</td>
<td>5.73</td>
<td>1.06</td>
<td>1.06</td>
</tr>
</tbody>
</table>

The above results show how the UGOT model has been calibrated and evaluated during 2000. To validate the model, the comparison with another period in 2001 (experiment Tide_3) when velocity observations were available was also performed. The model setup for experiment Tide_3 was similar to experiment Tide_2, except the time interval has been changed from year 2000 to year 2001. Two surface elevation data and one current velocity data were used to compare with the model results; the velocity measurement is only available only from 12 February 2001 14:00 to 1 March 2001 00:00. Figure 16 shows the comparison between observed (asterisks) and predicted (solid line) surface elevation at Mae Klong and Bang Pakong tide gauge stations and \( v \)–velocity at Si Chang buoy station obtained from experiment Tide_3. Visually, the model shows satisfactory agreement when compared with the observed data for both surface elevation and current velocity. For surface elevation, the RMSE (\( R^2 \)) are 20.72 cm (0.91) and 16.54 cm (0.94) for Mae Klong and Bang Pakong, respectively. The RMSE and \( R^2 \) of \( v \)–velocity at Si Chang station were 3.72 cm s\(^{-1}\) and 0.89, respectively. The results of experiment Tide_3 give us confidence that the UGOT model performed well and can be further used to study other forcing mechanisms (see following chapters).
Figure 16. Surface elevation ((a) and (b)) and $v$-velocity (c) comparisons between observed (asterisks) and predicted (solid line; experiment Tide_3) at Mae Klong (a) and Bang Pakong (b) tide gauge stations and Si Chang (c) buoy station. On the top right corner of each plot, $R^2$ and RMSE are presented.

Based on surface elevation data obtained from experiment Tide_2, harmonic analysis has been applied for all data points on the model domain and the plot of co-phase and co-range lines of all eight tidal components mentioned above (Figure 17). Since the barotropic Rossby radius of deformation of the UGOT is larger than the width of the UGOT (see Chapter 2 for detail), but it is not large enough to neglect the Coriolis parameter $f$. Kelvin waves can exist and their propagation along the coasts may be
seen in the tilted co phase lines (Figure 17): Kelvin waves would propagate northward/southward along the eastern/western coast. From Figure 17, it is found that in the UGOT, tidal amplitude of all the tidal components increase toward the head as expected (standing waves type) and tidal phase propagates in the counter clockwise direction, as expected in the northern hemisphere.

**Figure 17.** Co phase (blue solid line) and co range (black solid line) lines of all eight tidal components for the UGOT. Note that tidal amplitude and tidal phase are in cm and degree, respectively.
As mentioned above, tides in the UGOT move northward/southward during rising/falling of the tides. Therefore it will be interesting to see how current velocity in the UGOT varies during one month period (March 2000). Current velocities from sixteen selected locations (C01–C16) as seen on the map in Figure 18 obtained from experiment Tide_2 were analyzed and plotted in term of current variability and tidal ellipses. From Figure 18, it is found that northward/southward current appeared for the whole UGOT as expected. A strong current appears near the open boundary where deep water exists. In addition, the variability in x–direction has also been observed near the open boundary.

The current variability and tidal ellipses comparison between observed and predicted obtained experiment Tide_2 at Si Chang and Petchburi stations are shown in Figure 19. From Figure 19, it is found that predicted mean velocity was somewhat smaller than that observed. This might be because in the simulations shown so far the model was driven by tides only, neglecting wind-driven effects (discussed in Chapter 5). In general, the major axis and principal tidal current orientation obtained from model were comparable well with observed major axis and observed principal orientation.
Figure 18. Current variability from 16 locations (C01–C16) in the UGOT obtained from experiment Tide.2 during March 2000. Note that blue dots and red lines represent variability in $u$– and $v$–velocities (in cm s$^{-1}$) and tidal ellipses, respectively.
Figure 19. Current variability comparison between observed (top panel) and predicted flow obtained from experiment Tide 2 (bottom panel) at Si Chang ((a) and (c)) and Petchburi ((b) and (d)) stations during March 2000. Note that statistics are shown at the bottom left of each plot.
CHAPTER 5

FORCING MECHANISMS

5.1 SEASONAL WIND-DRIVEN CIRCULATION AND FLOW DYNAMICS

Wind is one of the most important forces that generate water movement in enclosed basins, semi-enclosed basins, and coastal regions. The linear, steady wind–induced circulation in such basins can be explained, for example, by a three-dimensional barotropic model in the elongated basin with arbitrary distributed depth and Coriolis parameter $f$ [Winant, 2004]. Sanay and Valle-Levinson [2005] studied wind–induced circulation in an elongated channel basin with variable depth using a numerical model. They found that in the non-rotating case, the alongshore wind induced transport in the direction of the wind in shallow areas, but flow is against the wind in deeper areas. This result agrees with results obtained by Csanady [1973], using an analytical model for Lake Ontario (a long and narrow lake with depth contour that runs parallel to the shore, neglecting the Coriolis effect). In a rotating system, the transport pattern is more complex and depends on the ratio of the maximum depth and the Ekman depth [Sanay and Valle-Levinson, 2005].

As mentioned in Chapter 2, the circulation in the UGOT is affected by monsoonal winds, in which the summer monsoon wind blows from the southwest direction while winter monsoon wind blows from the northeast direction. Based on two-dimensional and three-dimensional numerical model studies, the southwest/northeast monsoons cause the circulation in the UGOT to circulate in the clockwise/counter-clockwise direction [Buranapratheprat et al., 2006; Buranapratheprat, 2008; Buranapratheprat et al., 2009]. Note that these modeling studies used monthly mean wind fields while here much higher-frequency wind data, semi-daily winds, are used.
5.1.1 Model setup

In the study of the influence of wind, the 10-m wind vectors for the UGOT are retrieved from ERS-2 Scatterometer Mean-Wind-Fields (MWF) data and from the SeaWinds on QuikScat data (both data are available at http://poet.jpl.nasa.gov). The spatial resolution for ERS-2 and SeaWinds data are approximately 1.0 and 0.25 degree, respectively. The root mean square error of wind speed and wind direction obtained from SeaWinds data are 2 m s\(^{-1}\) and 20 degrees, respectively. In this section, only winds from SeaWinds data for the year 2000 are used. The spatially averaged wind vector covering the area of the UGOT is shown in Figure 20. The dominant winds are mainly prevailed from south or southwest direction, except during the northeast monsoon (typically, between November and February). During 2000, the northeast monsoon season lasts until January and February, though wind direction is variable during these months. March and April are mainly the southerly wind period, while November is mainly the north northeasterly wind period. These 3 months are chosen as examples to demonstrate the estuarine circulation caused by tides and winds (see later discussion in Section 5.1.3). The moderate winds (approximately 6.0 m s\(^{-1}\)) are mostly presented in the UGOT. Statistically, the wind speeds in year 2000 for the UGOT range from 0.36 to 18.55 m s\(^{-1}\).
Figure 20. The spatially averaged 10-m wind velocities over the UGOT in year 2000. In each plot, a semi-daily data for each month are plotted. The circle dashed lines represent the magnitude of wind velocity at 5 m s\(^{-1}\) interval.

Following the successful implementation of the tidal model in the UGOT (see Chapter 4), the semi-daily wind fields obtained from the SeaWinds data are then included as an additional forcing at the surface. Wind velocity components \(u\) and \(v\) are transformed into wind stresses (wind stress has negative sign for the eastward and northward wind velocities in the model). Winds applied on the surface are constant in space and variable in time. Thus, the influences of the wind caused by
the semi-daily variation on the circulation in the UGOT are expected. This model experiment was run for the whole year 2000. The model experiments in this section are shown in Table 9 (note that experiment Wnd.4 studies the sensitivity to Coriolis effect). For the purpose of analysis, the depth-integrated current velocities and surface elevation for the whole area are recorded in 1-hour intervals. The recorded data are then analyzed to find spatial mean of surface elevation and depth-integrated current velocities for each month. The results of this analysis are discussed next.

Table 9. Model experiments for the study of the impact of wind on the circulation feature in the UGOT.

<table>
<thead>
<tr>
<th>Exp.</th>
<th>Forcings</th>
<th>Model domain center</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wnd.1</td>
<td>Tides, Wind</td>
<td>~13°N</td>
</tr>
<tr>
<td>Wnd.2</td>
<td>Wind</td>
<td>~13°N</td>
</tr>
<tr>
<td>Wnd.3</td>
<td>Tides</td>
<td>~13°N</td>
</tr>
<tr>
<td>Wnd.4</td>
<td>Tides, Wind</td>
<td>~45°N</td>
</tr>
</tbody>
</table>

5.1.2 Model results

The monthly-averaged surface elevation and depth-averaged current velocities obtained from experiment Wnd.1 are shown in Figure 21 (January to June 2000) and Figure 22 (July to December 2000). During the northeast monsoon, a counterclockwise circulation has been observed, though the circulation features of January and February are not so clear compared with the circulation features in November and December. Wind fields in January and February are somewhat variable including
combination of northeasterly and southeasterly winds. The orientation of the spatially mean contour of monthly mean surface elevation (red solid line), clearly reflects the major wind direction; for instance in January 2000, a red solid line aligns in the northwest–southeast direction, when the winds mainly blow in northeast–southwest direction. A zero contour line of monthly mean surface elevation (white solid line) can explain the most dominant wind direction during that period. As seen in January 2000, a white solid line aligns in the northwest–southeast direction. The surface elevation is below mean on the right hand side which means that northerly or northeasterly winds are the most dominant. Hence the water has been pushed from the northern part the area toward the southwest corner. This is true for the duration of the northeast monsoon, except February when the negative mean surface elevation occurs near the bottom right corner and the winds mainly blow from the south. In March and April, even though southerly winds are clearly observed (see Figure 20), the circulation direction is not clear. Note that small features of clockwise circulation and counter-clockwise circulation have been observed near the top left corner and top right corner, respectively, for both months. In both months, mean surface elevation (mostly positive values) gradually increases toward the northern Gulf and red solid line lays nearly in the east–west direction. This means that during these periods winds mainly blow from the south or south southeast direction.

During the southwest monsoon, the prevailed winds cause the water in the UGOT to circulate in the clockwise direction (May to September). Southwesterly winds push water from the southwest boundary to accumulate at the northeast corner as seen in May to September. Southwest monsoon causes negative mean surface elevation to be located near the southwest corner. The clockwise circulation has been observed in October, despite winds mainly from west. That is why the red solid line aligns in the north–south direction and the water has been piled up parallel to the shoreline against the western boundary.
Figure 21. The plots of monthly mean depth integrated current velocities (cm s\(^{-1}\)) and monthly mean surface elevation (cm) from January to June 2000. Colorbar denotes mean surface elevation. The vectors on top right corner represent the 10 cm s\(^{-1}\) scale. White and red solid lines are zero and mean surface elevation contour lines, respectively. The circulation features are related to 10 m wind vectors as shown in Figure 20.
Figure 21 and 22 show some interesting features specifically at the northeastern part of the model domain between longitude 100.6°E to 101.0°E and latitude 13.1°N to 13.6°N. In this part of the model, the clockwise/counter-clockwise circulation is most apparent and its rotation direction changes seasonally. This is shown by the streamlines of relative vorticity (Figure 23). Vorticity, \( \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \), is a property that used to explain how the water spins. A negative/positive vorticity means water rotates clockwise/counter-clockwise. South of this area, a channel-like feature exists between Si Chang Island and mainland.

From Figure 23, the circulation features can be categorized into 4 groups. First, during northeast monsoon months (November-February 2000) besides the northward
flow along the eastern boundary that formed the counter-clockwise gyre for the whole model domain, a small clockwise gyre is generated north of Si Chang Island. Note that the shape of the clockwise gyre tends to elongate in north-south direction, except February in which the gyre is less pronounced. Second, during the southwest monsoon the eastward flow parallel to head of the model domain, which was part of the seasonal clockwise circulation, turned southwestward after reaching the eastern boundary. When it passes the headland (~13.3°N) and merges with the southward flow, it forms the counter-clockwise circulation between headland and Si Chang Island. The positive vorticity is clearly shown between May and September 2000. During this period, the counter-clockwise gyre tends to stick to the location between headland and Si Chang Island. Third, during the first inter-monsoon period (March-April 2000), the southerly winds dominate, and the effect is seen over half of the northeastern area in front of Bang Pakong River. The shape of the gyre extends from north of Si Chang Island farther north toward the top of the model domain. This gyre is larger than that of the clockwise circulation gyre during the northeast monsoon season. The fourth group, during the second inter-monsoon (October 2000), the major wind blows eastward, and except the large clockwise circulation in the whole model domain, the gyre in the northeastern area has disappeared.

From the findings above, one might question what happens if tides at the open boundary are neglected. Therefore, the model setup from experiment Wnd_1 was simulated but ignored tidal signals at the open boundary (experiment Wnd_2). The model results from the exclusion of tides at boundary show similar circulation of the UGOT as in experiment Wnd_1, except that it has a larger vorticity than in experiment Wnd_1 (Figure not shown).
Figure 23. Monthly depth averaged streamlines (white arrows) at northeastern part of the UGOT. Contour shows calculated vorticity in units of $\times 10^{-5}$ s$^{-1}$ based on depth averaged velocities and grid resolution. The colorbar denotes the magnitude of vorticity. Note that Si Chang Island is located at the southeast corner.

The model setup from Chapter 4 has been run for one year (experiment Wnd_3) in order to investigate the seasonal circulation features in the UGOT and to compare with experiment Wnd_1. Similar plots as shown in Figure 21 and 22. have been plotted for experiment Wnd_3 (Figures not shown). It is found that the tilting of
mean surface elevation and the clockwise/counter-clockwise circulations disappear. Hence, the tilting of surface and clockwise/counter-clockwise circulation features shown in Wnd_1 are purely caused by the wind forcing and not by residual tidal circulation.

Surface elevations at Mae Klong, Tha Chin, Phrachulachomkloa, Bangkok Bar, and Bang Pakong stations obtained from experiment Wnd_1 (March 2000) were compared with observations. Statistically, RMSE and R² show no difference from that found in Chapter 4. This is true for both harmonic analysis at five tide gauge stations and surface currents at two buoy stations. The conclusion is that inclusion of surface twice daily winds has little effect on instantaneous elevation or currents (which are mostly tidally driven). So, the wind mostly affects the seasonal circulation pattern as described above.

If Coriolis effect is not negligible even in this low latitudes, Ekman dynamics (i.e., wind-driven currents with balance between Coriolis and friction forces) need to be considered. To study the vertical distribution of velocity, the Ekman spirals were plotted along latitude 13.35°N for experiment Wnd_1 (Figure 24a–24f). The red arrow denotes the monthly average wind velocity in m s⁻¹ while blue arrow denotes the depth-averaged current velocity in cm s⁻¹. The locations of each plot are shown in Figure 24g (black plus sign). During November 2000 (wind mostly blows from the northeast direction) the surface current velocity is clearly deflected to right of ~5 m s⁻¹ mean surface wind (approximately 30–40 degrees) causing the net volume transport to be in the southwest direction. Hence the tilting up of sea level at the southwest corner can be seen as shown in Figure 22. In this case, the classical Ekman spiral is clearly present in the surface Ekman layer. In some locations and some periods, the surface currents are in the wind’s direction while the near-bottom flow is opposite to the wind direction (e.g., March 2000; Figure not shown). In the Ekman spiral (Figure 24), the current velocity decreases with depth as expected,
the exception is near the deep channel. On the right hand side of the deep channel looking northward the bottom flow was much larger than at the surface (Figure 24f) and outflow (southward flow) is situated a few meters below the surface. Beneath this depth, there is mostly inflow (northward flow).

Figure 24. Ekman spiral variations along latitude 13.35°N during northeast monsoon (November 2000). Thick and thin arrows represent wind and current velocities, respectively. (a)–(f) are corresponding to six locations of each Ekman spiral (plus signs in (g) from left to right). Note that current velocity has units of cm s\(^{-1}\), while wind velocity has units of m s\(^{-1}\).
With nearly the same mean wind speed as during the northeast monsoon, the Ekman spiral has also developed during southwest monsoon (Figure 25) and the current velocity at the surface is larger than that during northeast monsoon. The deflection of the surface current from the mean wind direction is also to the right with approximation 30 degrees (classic Ekman theory predicts 45 degrees angle). Most of the water across this section has been transported east to northeast as expected (a tilting of surface water at the northeast corner are shown in Figure 21), except at the deep channel where the northward flow appears a few meter near the surface while the southward flow occupies most of the water column (Figure 25e and 25f). It is interesting to note that when the wind is mainly from the west (October 2000), the Ekman spiral split into separate surface Ekman and bottom Ekman layers (Figure not shown).

A classical Ekman spiral in textbooks is based on several assumptions, such as constant vertical eddy viscosity \( K_M \), steady wind, no geostrophic flow, deep water, no stratification, etc [Pond and Pickard, 1983]. Assuming that the drag coefficient \( C_D \) is approximately \( 1.1 \times 10^{-3} \) (this number is calculated based on 5 m s\(^{-1}\) wind speed \( W \) at 10 m height above sea level), the Ekman depth \( D_E \) at known latitude \( \phi \) and wind speed \( W \) is \( D_E \approx 3.4 \frac{W}{\sqrt{\sin|\phi|}} \) [Pond and Pickard, 1983]. Given \( W = 5 \) m s\(^{-1}\) and \( \phi = 13^\circ\)N for the UGOT, the Ekman depth can be estimated as \( D_E = 3.4 \times 5.0/\sqrt{\sin|13|} = 35.8 \) m. Because the UGOT has an averaged depth approximately 15 m, the Ekman spiral is not likely to be fully developed (Figure 24 and 25). By definition the Ekman depth is the depth at which the current velocity flows in opposite direction as the current velocity at the surface and it represents the depth influenced by wind. The Ekman dynamics will further be discussed in the following section.
Figure 25. Similar to Figure 24, but for southwest monsoon (May 2000).

5.1.3 Discussions

5.1.3.1 Vertical distributions of velocity

Since the model domain is shallow and situated at low latitudes (~13°N), it is interesting to study how the Coriolis parameter, \( f \), would affect the Ekman spiral pattern (Figure 24 and 25). Therefore, another model experiment was performed (experiment Wnd.4). With the same model setup as experiment Wnd.1, except
that the model domain has been artificially shifted from low latitude to mid-latitude (~45°N). The Ekman spirals obtained from experiment Wnd.4 are quite similar to those in experiment Wnd.1, except two differences. First, at mid-latitude (experiment Wnd.4) the current velocity is weaker than that at low latitude (experiment Wnd.1). Second, the deflection angle of the current velocity to the right of the surface wind is larger at mid-latitudes, and the currents turn faster with depth than at low latitude. Figure 26 shows the Ekman spiral comparison at location (e) (Figure 24 and 25) between low latitude and mid-latitude during southwest and northeast monsoons. Figure 26 clearly shows the slower and more deflected of current velocities at mid latitude.

In the classical Ekman transport theory (e.g., Mellor [1996]), velocities $u$ and $v$ can be derived as

$$u = u_g + \frac{e^5}{\sqrt{K_M}f} \left[ \frac{\tau_{0x}}{\rho_0} \sin \left( \zeta + \frac{\pi}{4} \right) + \frac{\tau_{0y}}{\rho_0} \cos \left( \zeta + \frac{\pi}{4} \right) \right]$$  \hspace{1cm} (4)

$$v = v_g + \frac{e^5}{\sqrt{K_M}f} \left[ -\frac{\tau_{0x}}{\rho_0} \cos \left( \zeta + \frac{\pi}{4} \right) + \frac{\tau_{0y}}{\rho_0} \sin \left( \zeta + \frac{\pi}{4} \right) \right]$$ \hspace{1cm} (5)

where $\zeta \equiv z\sqrt{f/2K_M}$, $\rho_0$ is a reference density, $u_g$ and $v_g$ are geostrophic velocities in $x-$ and $y-$direction, respectively. $\tau_{0x}$ and $\tau_{0y}$ are wind stresses at the surface in $x-$ and $y-$direction, respectively. By assuming no wind stress in $y-$direction ($\tau_{0y} = 0$), $K_M$ is constant, and at $z = 0$, from equations (4) and (5) $u - u_g$ and $v - v_g$ are proportional to $A/\sqrt{f}$ and $B/\sqrt{f}$, respectively, where $A = \tau_{0x} \sin(\pi/4)/\sqrt{K_M}$ and $B = -\tau_{0x} \cos(\pi/4)/\sqrt{K_M}$. At low latitude, the value of $f$ is less than at mid-latitude. Therefore, the current velocity is stronger at low latitude than at mid-latitude (Figure 26). Again from equations (4) and (5) as the water depth $z$ decreases (more negative) the terms $\sin(\zeta + \pi/4)$ and $\cos(\zeta + \pi/4)$ at mid-latitude are going to change sign faster than that at low latitude. This leads to the Ekman spiral spins with depth faster at mid-latitude than at low latitude as shown in the above result.
Based on the equation shown in Section 5.1.2, the Ekman depth for experiment Wnd.4 can be estimated and it was $D_E = 3.4 \times 5.0 / \sqrt{\sin |45|} = 20$ m. This number was approximately the same as the maximum depth appears at the channel of cross-section as seen in Figure 24 and 25. Hence by definition the bottom velocity at the bottom of the Ekman depth was expected to be in the opposite direction to the surface, but this did not occur, probably due to bottom friction.

**Figure 26.** Ekman spiral comparison between experiment Wnd.1 (top panel) and Wnd.4 (lower panel) along latitude 44.35°N at 17.90 m depth during southwest monsoon ((a) and (c)) and northeast monsoon ((b) and (d)). Thick and thin arrows represent wind and current velocities, respectively. Note that current velocity has units of cm s$^{-1}$, while wind velocity has units of m s$^{-1}$. 
Because river discharge was excluded from the model simulation and the initial vertical salinity distribution was nearly constant, the homogeneous water column can be assumed. In a homogeneous non-rotating and frictionless coastal system the surface wind stress is in balance with the pressure gradient where the water depth is equal to the mean depth, hence the transport is zero. If the water depth is less than the mean depth, the surface wind stress overwhelms the pressure gradient and drives the water in the wind direction. On the other hand, if water depth is greater than the mean depth; the water flows against wind [Csanyi, 1973]. Therefore, coastal jets often develop along the shoreline (in the wind direction) and return flow develops at deeper channels (against the wind). Figure 27 shows the transport across latitude 13.35°N from experiment Wnd.1 for different wind directions.

The transport is influenced by the bottom topography. In the cases of wind blowing from the south or southwest direction (March and May 2000), the northward transport appeared in the shallow regions while the southward transport is found in the deeper regions. When the wind is from the northeast direction, the flow pattern reverses, so the southward and northward transports were located at the shallow and deep area, respectively. November 2000 has the largest transport because wind is strongest in the northeast direction. Even though bottom friction and Coriolis effects are included in experiment Wnd.1, the results agree quite well with the analytical results for the non-rotating frictionless coastal system shown in Csanyi [1973].

Coastal transport in the wind direction is seen at both sides of the cross-section for all months (Figure 27), except when the wind blows from the west (October 2000). October 2000 (blue solid line) show only the coastal transport on the western side, but it vanishes on the eastern side, otherwise. The October transport pattern is similar to that in March and May. The maximum transport is situated almost at the maximum depth for both northeast and southwest monsoons.
Figure 27. North-south transport (m$^3$ s$^{-1}$) across latitude 13.35°N looking to the north for different months in 2000, March (red), May (green), October (blue), and November (magenta). Colored arrows denote monthly averaged wind velocity for each month while the horizontal dotted line denotes the mean water depth level.

In the frictionless non rotating system, the transport is zero at locations with water depth close to the mean depth of the section (~11.5 m in Figure 27). However in this case, the nearly zero transport occurs at a depth of approximately 15 m. This is consistent with the case when the bottom friction was excluded in the model [Csanady, 1973].
The subtidal circulation in estuaries usually involves three-dimensional processes, though past studies are often interested in two-dimensional axial processes (i.e., vertical and longitudinal variability); transverse variability (i.e., across an estuary) is often neglected [Wong, 1994]. However, the dynamics of transverse distribution is also crucial [Kasai et al., 2000; Wong, 1994]. Therefore the velocity distributions across a north-south section along longitude 100.61°E were extracted for March, May, October, and November 2000 to investigate the effect of the wind direction on the velocity variability. The plots of the monthly averaged $u$— and $v$—velocities of these four months are shown on the left and the right panel of Figure 28, respectively. During southerly wind (March 2000) $u$—velocity is almost a unidirectional flow for the entire water column (Figure 28a), which is similar to $v$—velocity during westerly wind (Figure 28f). The velocity component perpendicular to the wind direction has this feature. Except for these two times, other $u$— and $v$—velocities are nearly two-layer.

Consider the $u$—velocity, farther north of latitude 13.3°N the flows for the entire water column are eastward and westward jets during southwesterly (Figure 28c) and westerly (Figure 28e) winds and northeasterly wind (Figure 28g), respectively. South of latitude 13.3°N, there exists a two-layer flow. There are eastward and westward flows at the top and bottom layers during southwesterly and westerly winds and vice versa during northeasterly wind. The $v$—velocity component shows strong two-layer flow, northward flow at the top layer and southward flow at the bottom layer for both southerly (Figure 28b) and southwesterly winds (Figure 28d). The flow reverses during the northeasterly wind (Figure 28h). Note that the surface layer flow of $v$—velocity occupies one-fourth of the water column, especially in the deeper area.

Consider the line of zero $v$—velocity on top of a small concave bottom topography near latitude 12.98°N, it has a raised dome and a bowl shapes during southerly or southwesterly wind and northeasterly wind, respectively. Because in the axial plane
the open boundary is at the south end, the dynamics might be different from the transverse plane with closed boundaries at both, the eastern and western sides.

The transverse distributions across an east–west section at latitude 13.35°N for those four months as described above are shown in Figure 29. The left panel shows the $u-$velocity while right panel shows the $v-$velocity. Like in the axial distribution $u-$ and $v-$velocities across latitude 13.35°N for March (Figure 29a) and October (Figure 29f), respectively, almost flew in one direction with depth (i.e., mostly a barotropic flow). Consider $u-$velocity, most of the flow is to the right hand side of the wind, except in the deeper part of the channel and over the right shoal (looking northward) where a reversely flow is found (Figure 29c, 29e, and 29f). For $v-$velocity, two-layer flow seemed to be developed in the area where the water depth is deeper than 10 m. Except March and October, a unidirectional flow has been observed in the deeper part of the channel (see Figure 29d and 29h).
Figure 28. The distributions of $u-$ (left panel) and $v-$ (right panel) velocities (monthly averaged) along longitude 100.61°E looking to the west for different months in 2000. White solid line separates positive and negative velocities from each other. Velocity units are cm s$^{-1}$. 
Figure 29. The distributions of $u-$ (left panel) and $v-$ (right panel) velocities (monthly averaged) along latitude 13.35°N looking to the north for different months in 2000. White solid line separates positive and negative velocities from each other. Velocity has units of cm s$^{-1}$. 
In the studies of transverse variability in estuaries, the Ekman number, $E_k$, determines what forces, viscosity or geostrophic, play the major role in controlling the flow structure [Kasai et al., 2000; Valle-Levinson et al., 2003]. The Ekman number is defined as $E_k = K_M/f H_0^2$, where $H_0$ is the maximum depth. In addition the Ekman depth can be written as $D_E = \sqrt{2K_M/f}$. The relationship between these two parameters is in the form $D_E = \sqrt{2E_k H_0}$. According to Kasai et al. [2000], for large Ekman number $E_k = 1$, $f$ can be neglected. This leads to $D_E = 1.4 H_0$, therefore the Ekman layer can occupy the whole water column and the pattern of the axial flow ($v$–velocity) will look similar to the one that shown in Wong [1994], in which an inflow is situated in the deeper part, while the outflows are on the shoals (both sides).

If $E_k = 0.1$, then $D_E = 0.45 H_0$. This means the upper part of the water column is controlled by $f$, while the lower half may be affected by a separate Ekman bottom layer. In the case of very small Ekman number $E_k = 0.01$ or $D_E = 0.14 H_0$, the vertical eddy viscosity is weak, the axial flow is nearly geostrophic and the Ekman layer is pressed near the bottom (i.e., a two–layer flow is expected).

The cross-sectional averaged vertical eddy viscosity (from the Mellor–Yamada turbulence closure scheme [Mellor and Yamada, 1982]) along latitude 13.35°N based on experiment Wnd.1 and corresponded Ekman number $E_k$ and Ekman depth $D_E$ for March, May, October, and November are shown in Table 10. Ekman number $E_k$ for March was approximately 0.3 and $D_E$ is less than the maximum depth ($H_0 = 20$ m). That is why two–layer flow is present (Figure 29b). The Ekman numbers for May and November gave the Ekman depths close to the maximum depth. Thus in the deeper channel the whole water column are occupied by the Ekman layer. Therefore in the deeper channel, the outflow and inflow (looking northward) appears for the whole water column during May and November, respectively (see Figure 29d and 29h).
Table 10. Mean vertical eddy viscosity cross-section ($K_M$) along latitude 13.35°N based on experiment Wnd.1 and calculated Ekman number ($E_k$) and Ekman depth ($D_E$) for March, May, October, and November 2000.

<table>
<thead>
<tr>
<th>Month</th>
<th>$K_M$ (m$^2$ s$^{-1}$)</th>
<th>$E_k$</th>
<th>$D_E$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>March</td>
<td>0.0033</td>
<td>0.3</td>
<td>14.2</td>
</tr>
<tr>
<td>May</td>
<td>0.0077</td>
<td>0.6</td>
<td>21.8</td>
</tr>
<tr>
<td>October</td>
<td>0.0164</td>
<td>1.2</td>
<td>31.7</td>
</tr>
<tr>
<td>November</td>
<td>0.0065</td>
<td>0.5</td>
<td>20.0</td>
</tr>
</tbody>
</table>

5.1.3.2 Comparisons between the numerical model and an analytical solution

Winant [2010] described an analytical solution for the wind-driven viscous flow in a closed shallow basin. It is assumed that the length, $L^*$ of the basin ($x-$direction) is larger than the width, $W^*$ ($y-$direction) and the width is much larger than the water depth, $H^*$. The momentum equations in $x-$ and $y-$directions can be written as

$$\frac{\partial^2 u}{\partial z^2} - \frac{\partial \eta}{\partial x} = 0$$

$$\frac{\partial^2 v}{\partial z^2} - \frac{\partial \eta}{\partial y} = 0,$$

respectively, where $u$, $v$, $\eta$, $x$, and $z$ are non-dimensional axial velocity, transverse velocity, surface elevation, axial coordinate, and vertical coordinate, respectively.
These four variables are non-dimensionalized as

\[ u = u^* \frac{\rho K_M}{\tau H^*}; \quad v = v^* \frac{\rho K_M}{\tau H^*}; \quad \eta = \eta^* \frac{\rho g H^*}{\tau L^*}; \quad x = x^* \; \frac{L^*}{x^*}; \quad y = \frac{y^*}{W^*}; \quad \text{and} \quad z = \frac{z^*}{H^*}, \]

where \( \rho \) is density of sea water; \( \tau \) is surface wind stress in positive \( x \)-direction. The variables with superscript \(*\) are dimensional variables. Equations (6) and (7) show the balance between the vertical stress divergence (first term on the left) and the pressure gradient (second term on the left). There are two boundary conditions for this problem which are

at \( z = -h \), \( u = v = 0 \) and at \( z = \eta \), \( \frac{\partial u}{\partial z} = 1 \) and \( \frac{\partial v}{\partial z} = 0 \),

where \( h \) is non-dimensional water depth. By integrating equations (6) and (7) and applying boundary conditions as shown above then the solution \( u \) and \( v \) can be expressed as

\begin{align*}
  u &= \frac{\partial \eta}{\partial x} \left[ \frac{z^2}{2} - h^2 - \eta(z + h) \right] + z + h \quad \text{(8)} \\
  v &= \frac{\partial \eta}{\partial y} \left[ \frac{z^2}{2} - h^2 - \eta(z + h) \right]. \quad \text{(9)}
\end{align*}

Note that the second term on the right hand side of equation (8) can be neglected if \( h \) is very small. By definition, the axial transport \([u]\) can be defined as

\[ [u] = \int_{-h}^{\eta} u \, dz \approx \int_{-h}^{0} u \, dz = \frac{h^2}{2} - \frac{\partial \eta}{\partial x} \frac{h^3}{3}. \quad \text{(10)} \]

By assuming no net flux in \( x \)-direction and \( v = 0 \), then equation (10) can be integrated across the basin and it can be shown that

\[ \int_{-1}^{1} [u] \, dy = \int_{-1}^{1} \frac{h^2}{2} \, dy - \frac{\partial \eta}{\partial x} \int_{1}^{3} \frac{h^3}{3} \, dy = 0. \quad \text{(11)} \]
From equation (11), it can be rewritten as

\[
\frac{\partial \eta}{\partial x} = \frac{3}{2} \int_{-1}^{1} h^2 \, dy = \frac{3 \langle h^2 \rangle}{2 \langle h^3 \rangle}.
\]  

Substituting equation (12) in equation (8) and applying the cross-section water depth along latitude 13.35°N the axial velocity can be estimated and it is shown in Figure 30b. A comparison between the result obtained from the three-dimensional numerical model (Figure 30a) and that obtained from the analytical model (Figure 30b), shows a very similar flow pattern. Flow against the wind appeared at the deeper channel and flow with the wind appeared on the shoals on both sides, as expected. Note that the numerical model also has additional small-scale variations near the surface, which are associated with three-dimensional eddies and gyres that cannot be captured by the analytical solution. The analytical model of wind-induced flow based on Winant [2004, 2010] has been also applied for cross-section at the entrance of the Nansemond River and their result showed the consistency with the observation data [Narváez and Valle-Levinson, 2008].
Figure 30. $v$–speed distribution comparison obtained from numerical model (a) and analytical model (b) along latitude 13.35°N during the northeast monsoon (November 2000). A white solid line represents a zero contour line. Note that velocity is in units of cm s$^{-1}$.

5.2 SURFACE HEAT FLUXES

Since surface wind stresses influence the circulation in the UGOT as shown above, they might also influence the distribution of surface temperature. For instance, the study of Buranapratheprat et al. [2008a] shows that temperature distributions in the UGOT are considered to be related to four major factors; surface heat fluxes, freshwater discharges, surface wind stresses, and tidal mixing. Therefore seasonal variability of temperature caused by seasonal changes in surface wind stresses (both magnitude and direction) and surface heat fluxes are crucial. In this section, simulation of the
UGOT model are considered with forcing by surface heat fluxes together with surface wind stresses and tidal mixing. The influence of seasonal surface wind stresses and surface heat fluxes on temperature distribution is presented in the following section.

5.2.1 Data acquisition and model setup

The model configuration of experiment Wnd.1, with constant (zero) surface heat fluxes and with only wind and tidal forcing is insufficient to simulate variations in temperature, thus surface heat fluxes are now added. The surface heat fluxes were implemented in the model based on the formulation of Ezer [2000] with additional considerations for the impact of cloud coverage on shortwave radiation. Surface heat fluxes, $Q$, in the model is calculated as

$$Q = Q_c + \left( \frac{\partial Q}{\partial T} \right) (T_o - T_m) + Q_{sw} \times F_s,$$

where $Q_c$ is the net heat flux ($Q_{net}$), excluding shortwave radiation term ($Q_{sw} \times F_s$). $T_o$ and $T_m$ are surface temperature obtained from observed and model data, respectively and $F_s$ is cloud factor. The second term on the right-hand side is the so-called feedback term with the magnitude of the coupling coefficient, $\partial Q/\partial T$, taken as approximately 50 W m$^{-2}$ K$^{-1}$ as used by Ezer [2000]. $F_s$ is calculated as $0.6 + 0.4 \times \%\text{Clear sky}/100$ based on empirical tests for this particular model domain; because of the large seasonal variations in cloud cover associated with the monsoon pattern, it is necessary to add this correction. $Q_{net}$, $Q_{sw}$, and $T_o$ are obtained from re-analyzed data from Yu et al. [2008]. All data used in this study was retrieved from http://rda.ucar.edu, except %Clear sky that was estimated here based on sea surface temperature retrieved from AVHRR Pathfinder version 5 (http://poet.jpl.nasa.gov); wind vectors, ERA-40, were retrieved from http://data-portal.ecmwf.int/data/d/era40.daily/. An ECMWF 40 Year Re-analysis data
(ERA-40) provides global wind vectors with a resolution of $2.5^\circ \times 2.5^\circ$ covering mid-1957 to mid-2002 (6-hour data). It has been used in this model experiment because it provides a continuous data coverage for all 4 years (1997 to 2000), while wind vectors used in experiment Wnd.1 are available only from mid-1999 to late-2009.

Figure 31 depicts variations in net heat flux, shortwave radiation, surface temperature, %Clear sky, and wind vector between 1997 and 2000. Note that unlike mid-latitudes, no clear seasonal variations are seen in either the net heat flux (Figure 31a) or shortwave radiation (Figure 31b). Most of the time net heat flux has a positive sign (positive means downward heat flux from air to ocean), while shortwave radiation always has downward flux from air to ocean. Unlike net heat flux and shortwave radiation, temperature data shows not only seasonal variation but also inter-annual variation (see Figure 31c). 1998 was a warmer (El Niño) year with mean surface temperature of approximately 29.47°C, while 2000 was colder (La Niña) with a mean surface temperature of approximately 28.42°C.

Cloudiness (Figure 31d) in terms of %Clear sky shows a smaller percentage of clear sky during southwest monsoon (wet season) and conversely for northeast monsoon (dry season). Consider the number of days in 1998 and 2000 in which %Clear sky is greater than mean %Clear sky of all 4 years. There were 138 and 87 days in 1998 and 2000, respectively. This might be one reason why 1998 is warmer than in 2000. As expected, a seasonal variability in wind direction corresponding to northeast and southwest monsoons is clearly shown (Figure 31e). Comparing to other years, southerly winds are frequently observed in both 1997 and 1998.
Figure 31. Daily net heat flux (a), shortwave radiation (b), surface temperature (c), %Clear sky, and wind vector (e) from year 1997 to 2000. Note that gray shading separates one year from the next and Julian day starts from January 1, 1997.
Sea surface temperature retrieved from satellite sensors illustrates some interesting features during the northeast and inter-monsoons (Figure 32). On January 31, 2000 (winter season), sea surface temperature in the UGOT showed an intrusion of warmer water from the south compared with ambient temperature (colder water, \(\sim 26^\circ C\)) along the east coast (Figure 32; left panel). This warmer water has been advected northward reaching latitude 13.20°N (approximated). This feature is opposite during the summer period (April 1, 2000, Figure 32; right panel). Colder water (\(\sim 29^\circ C\)) compared with ambient temperature (colder water, \(\sim 30^\circ C\)) has been advected northward along the west coast. Figure 32 demonstrates how important the monsoonal winds are in affecting sea surface temperature in the UGOT. Also revealed is the importance of Upper Gulf opening as a source of colder/warmer water in the UGOT. Note also that the small-scale variations in observed temperatures, similar features are seen in the model simulations as well, though there are not enough data for detailed comparisons. In addition, the model is forced by spatially constant (but time-dependent) surface forcings.
In the following experiment, a 4-year run was conducted for 1997 to 2000. It is found that about three months are needed to adjust to initial conditions spinning up period, so only the last three years (1998–2000) will be analyzed in the following section.

5.2.2 Model results and discussions

In this section the seasonal variation and inter-annual variability of surface temperature will be discussed.

5.2.2.1 Seasonal variation in surface temperature

The first year of the model results (1997) is neglected as part of the spin up. 1998 and 2000 are affected by El Niño and La Niña, respectively. Thus, seasonal variation
will be considered in this section, using 1999 as representing a typical year.

Monthly averaged sea surface temperature from 1999 is depicted in Figure 33. During the winter season (November to February) with northeast monsoon, it was found that a tongue of intermediate temperature water (~29°C) has been advected westward along the northern coast. An intrusion of colder water from the open boundary was observed along the southeastern coast reaching ~13.1°N. Northward and westward dispersion along the coast of surface temperature can be explained by a counter-clockwise circulation due to northeasterly wind (see detail in Section 5.1.2). During the first inter-monsoon (March to April), wind mainly blew northward causing flow to move southward against the wind direction. That is why the band of warmer water tends to propagate southward and the northward propagation of colder water along the southeastern coast tends to disappear. Similar features were illustrated for the second inter-monsoon (October). During the summer southwest monsoon (May to September), a plume of intermediate water has been pushed northward along the western coast causing warmer water to disperse southward along the eastern coast. This feature is explained by the clockwise circulation caused by the southwesterly wind. Tongue of intermediate water propagates eastward along the northern coast as seen in August, which is similar to the case of northeast monsoon season (between December and January). Warmer water dispersed southward reaching ~13.0°N. Note that warmer waters are found along the western coast throughout the year, even during southwest monsoon when water in the whole area circulated clockwise.

All features for different seasons as mentioned above have also been observed for other years (Figure not shown). The spatial variations in temperature is driven by the model dynamics despite the surface forcing which is constant in space. In addition to plumes coming from the south, the mechanism suggested here is that the shallower western and northern coasts are warmed faster than the deeper regions.
These warmer waters are dispersed by the wind-driven coastal currents.

Figure 33. Monthly averaged sea surface temperature obtained from model simulation that included surface heat fluxes and wind stresses and tides at open boundary in year 1999.

The results explained above and simulated sea surface temperature features seemed to agree well with observed sea surface temperature obtained from satellite images. Both warm and cold waters intrude to the UGOT through the open boundary along east and west coast during northeast and southwest monsoons, respectively. Comparing surface temperatures in 1999 (Figure 33) to those in 1998
and 2000 (Figure 34) show similar seasonal patterns, but with warmer/cooler temperatures in 1998/2000. Colder water intrusions northward along east coast and intermediate water along the west coast during northeast and southwest monsoons were shown in all three years.

Figure 34. Seasonal variations in sea surface temperature from 1998 to 2000 (top to bottom). Different seasons (from left to right) are NEXX XX (NDJF), InterXX-1st (MA), SWXX (MJJAS), and InterXX-2nd (O) are northeast monsoon, first inter monsoon, southwest monsoon, and second inter monsoon seasons, respectively. Note that months for each season are shown in parentheses and X denotes number (year).
5.2.2.2 Inter–annual variability in surface temperature

Sea surface temperature obtained from the model is spatially and daily averaged to remove tidal and other high–frequency variations. The model results are compared in Figure 35 with temperatures obtained from the re–analysis data (which is used as an input temperature $T_o$ in equation (31)) and with observed temperatures retrieved from satellite images for both day time and night time. Figure 35 indicates two important findings. First, observed sea surface temperature from both satellite images and re–analysis data are comparable, though few temperature data from satellite images are available during the southwest monsoon due to more clouds. With the inclusion of feedback terms, equation (13), a spatially and daily averaged surface temperature reproduces the seasonal pattern of temperature found in the UGOT, though 1–2°C observed–model differences are found in all three years (see also Figure 36; right panel). The absolute difference of sea surface temperature between observed and predicted (model) for each month between 1998 and 2000 decreases during the southwest monsoon and increase during the northeast monsoon. Note that monthly averaged modeled surface temperatures are over–estimated for the whole three years. Statistically, mean absolute error plus/minus one standard deviation of sea surface temperature for 1998, 1999, and 2000 are $0.80 \pm 0.29$, $1.08 \pm 0.32$, and $0.97 \pm 0.29^\circ$C, respectively.

Second, an inter–annual variability in surface temperature is presented. In 1998, surface temperature is generally warmer than that in 1999 and 2000. This result is consistent with multivariate ENSO Index or MEI [Wolter and Timlin, 1993, 1998] (see Figure 37). In 1998 which is an El Niño year, sea surface temperature in the ocean was expected to be warmer than La Niña years (1999 and 2000). The MEI data is retrieved from http://www.esrl.noaa.gov/psd//people/klaus.wolter/MEI/#Home. Monthly sea surface temperature differences between actual temperature and mean
temperature (Figure 36a, 36c, and 36e) also show large differences from mean temperature during summer of El Niño year (Figure 36a) and during winter of La Niña year (Figure 36c and 36e).

![Figure 35. Sea surface temperature comparison between observed and predicted for 1998 (a), 1999 (b), and 2000 (c). Red and green solid lines represent temperature $T_a$ in (13) and predicted temperature, respectively. Note that asterisks depict temperature obtained from satellite images during day (blue) and night (magenta) times.](image-url)
Figure 36. Monthly sea surface temperature anomaly, $T - T_{avg}$, (left panel) and absolute error, $|Obs$-$Pred|$, of monthly sea surface temperature and one standard deviation during 1998 to 2002 (right panel).
From Figure 34, sea surface temperature in 1998 is higher than the other two years, which is consistent with spatially averaged temperature shown in Figure 35. However, seasonal pattern of NE97 98 was somewhat different from other two northeast monsoon seasons (NE98 99 and NE99 00). Its distribution is similar to the first inter monsoon (Inter00 1st). One explanation might be due to inter annual changes in ocean circulation, so current velocity in 1998 and 2000 are compared in Figure 38. Since wind directions in January 1998 were mainly southerly, while in January 2000 were northeasterly and southwesterly a difference in current direction was expected especially along the northern boundary. Current velocities in July 1998 and in July 2000 were not much different since wind blew almost at the same direction (southwesterly wind).
Figure 38. Monthly depth-averaged current velocity for 1998 (top panel) and 2000 (bottom panel). Left panel is for January (northeast monsoon) while right panel is for July (southwest monsoon).
5.3 EFFECTS OF RIVER DISCHARGES

Coastal marine ecosystems are sensitive to the input of freshwater from the rivers which bring sediments, nutrients and polluted substances. Physically, plumes of freshwater generated by buoyancy inflow contributes a significant dynamical impact on the coastal circulation [Horner-Devine, 2009]. In coastal circulation models, the freshwater plume is deflected instantly to the right/left in the Northern/Southern hemisphere due to earth rotation (the Coriolis effect) and the coastal current will generally carry the water away from the river mouth [Horner-Devine, 2009]. The above behavior of a river plume is based on the assumptions that there is a water entering to the coastal from a medium to large-scale rivers and Coriolis is not negligible.

During the southwest monsoon season of year 2011, Thailand encountered a severe flooding that began at the end of July and ended around mid-January. By October, flood water reached the mouth of Chao Phraya River, bringing large amounts of material into the UGOT. A false color THEOS satellite image (Figure 39) obtained from the website of Geo-Informatics and Space Technology Development Agency (Public Organization), GISTDA, clearly shows the turbidity plume coming from upstream, especially the Chao Phraya River. Note that the image is taken on November 29, 2011 11:00 am (an hour after high tide). The turbidity plume from the Chao Phraya River is quite symmetric when it is compared to other plumes found at high latitude.

Uncles and Stephens [1997] found a linear relationship between turbidity and lower salinity (less than 30) in the Tweed Estuary, UK. Hence, the turbidity plume can be used to detect the freshwater plume.

In this section, the implementation of rivers in POM and the impacts of river discharge on the salinity and temperature distributions and the circulation near the freshwater plumes are demonstrated. The fact that the UGOT is located at low
latitude (small Coriolis parameter) is also evaluated.

Figure 39. THEOS satellite image (a false color image) shows a turbidity plume (gray area in front of river mouths and along the coast) at the UGOT during the 2011 Thailand floods. White areas are cloud cover, while red areas are vegetation. Blue dots locate the four main rivers in the UGOT. This image is modified from the original image obtained from http://www.gistda.or.th/gistda_n/Gallery/img/Flood2011/.

5.3.1 Model setup

The implementation of rivers in the model is based on the work of Oey [1996]. In this study, there are four rivers situated along the northern boundary of the model (see Figure 13). From left to right, the rivers are Mae Klong, Tha Chin, Chao Phraya, and Bang Pakong rivers, respectively. The monthly discharge for each river is shown in Figure 40. The water discharge \( Q_{riv} \) from each river is assigned a
downward vertical velocity \(w = -\frac{Q_{riv}}{N_{riv}}dx dy\), where \(N_{riv}\) is the number grid cells in which the water discharge is to be distributed and \(dx\) and \(dy\) are grid cell sizes in \(x-\) and \(y-\)directions, respectively. Note that the zero salinity was defined at each river location.

In order to investigate the influences of river on the salinity, temperature distribution and coastal circulation, four different experiments are conducted (see Table 11). The first two experiments show how rivers, tides and the Coriolis parameter \(f\) impact the distribution of salinity, temperature and coastal circulation. So, these experiments have been forced by surface heat fluxes, tides at the open boundary and rivers at the river mouths. In experiment Riv.2, the original latitude of the UGOT from experiment Riv.1 is replaced by a latitude that is 32 degrees higher in order to see how river plumes behave in a mid-latitude model, but otherwise the same conditions. The last two experiments (Riv.3 and Riv.4), are similar to the first two experiments, except that the surface wind stresses are included. The surface wind stresses are spatially constant but vary in time. The model experiments are run for one year (year 2000). The initial condition for salinity and temperature at the surface are set to be spatially constant for the whole model domain but be varied with depth. The model sea surface salinity, surface temperature, and surface current velocity are saved every 3 hours.
Figure 40. Mean monthly mean water discharge (m³ s⁻¹) at Mae Klong (•-•), Tha Chin (○-○), Chao Phraya (□-□) and Bang Pakong (×-×) Rivers. Water discharge peaks between September and October with the Chao Phraya River discharging more water than the others.

Table 11. Model experiments for the study of the impact of river on the salinity and temperature distribution and coastal circulation near the river mouth.

<table>
<thead>
<tr>
<th>Exp.</th>
<th>Forcings</th>
<th>Model domain center</th>
</tr>
</thead>
<tbody>
<tr>
<td>Riv.1</td>
<td>Rivers, Surface heat fluxes, Tides</td>
<td>~13°N</td>
</tr>
<tr>
<td>Riv.2</td>
<td>Rivers, Surface heat fluxes, Tides</td>
<td>~45°N</td>
</tr>
<tr>
<td>Riv.3</td>
<td>Rivers, Surface heat fluxes, Tides, Wind</td>
<td>~13°N</td>
</tr>
<tr>
<td>Riv.4</td>
<td>Rivers, Surface heat fluxes, Tides, Wind</td>
<td>~45°N</td>
</tr>
</tbody>
</table>
5.3.2 Model results

5.3.2.1 Salinity and temperature distributions

Monthly averaged sea surface salinity and sea surface temperature are obtained from the model experiments for all months (not shown). However, September is selected as a good representative of the sea surface salinity as it is a time of freshwater discharge (Figure 40). June is selected for the temperature distribution as it is a warm month (Figure 35).

The spatial mean monthly surface salinity and surface temperature obtained from the analysis of experiments Riv_1 and Riv_2 are presented in Figure 41. From Figure 41, it is clearly seen that Chao Phraya River produces a large plume of freshwater across the coastal region; it is the largest river in the GOT. First, the case with a realistic low latitude Coriolis parameter, but without wind, is illustrated (upper panels of Figure 41). The 20 salinity contour of the plume symmetrically extends approximately 30 km from the river mouth (Figure 41a). The symmetric plume (similar to the observed plume in Figure 39) is also found at the other three rivers, but the distances of the plume from the river mouth are shorter than the Chao Phraya River. The amount of water discharge might be one of the key factors that controls how far from the river mouth the plume can penetrate. In addition, the tongue of freshwater (low salinity) is transported south along the western boundary.

Since the feedback term is included in the calculation of the surface heat fluxes (see Section 5.2 for more detail), the temperature distribution is expected to change as the time progresses, whereas the warmer river plume loses heat to the atmosphere. Figure 41b shows how warm and cool waters distribute in the model domain. Clearly, the rivers are the sources of warm waters (\(\sim29-30^\circ\text{C}\)), while cool waters (\(\sim25-26^\circ\text{C}\)) are transported from the lower GOT through the open boundary (southern
boundary). Similar to the plume of freshwater, the warm water plume is slightly transported to the right hand side along the coast. The extension of a warm water plume from the Chao Phraya River is approximately the same distance as the case of the plume of freshwater, but the shape of the plume is not quite symmetric. The interesting feature of temperature is that the warm and cool waters have been transported to the right along the coast and the warm water extended almost to the open boundary (around 12.80°N). In these results, tidal forcing is applied at the open boundary, but wind stress is ignored. So, the mechanisms that control the transport of both surface salinity and surface temperature is likely the tidal–driven circulation and tidal mixing.
Figure 41. Monthly mean surface salinity and mean surface temperature with river flows, but no winds. The vector fields represent the mean surface current velocities. (a) salinity distribution (September 2000) and (b) temperature distribution (June 2000) for experiment Riv_1. (c) salinity distribution and (d) temperature distribution for experiment Riv_2 (mid-latitude). Note that colorbar for salinity and temperature are different in scale.

As mentioned above, in the Northern hemisphere, the coastal models show the deflection to the right of the river plume in the case of medium to large river input. From Figure 41a, despite the Chao Phraya River is the largest river in Thailand,
the model result does not clearly show the plume deflection. Is this because the water discharge from Chao Phraya River is not large enough (in September, the water discharge from the Chao Phraya River is approximately 1,700 m$^3$ s$^{-1}$)? Or is it because the model domain is located at low latitude (~13°N)? Figure 41c and 41d show mean surface salinity and mean surface temperature in the case where the model domain has been shifted from low latitude to mid-latitude (~45°N, Riv_2). Experiment Riv_2 shows that the plume of the Chao Phraya River and other plumes tend to spread farther along the coastline, but extends less away from the coast, in a more asymmetric fashion than in experiment Riv_1.

Consider the 20 salinity contour of the Chao Phraya River plume in experiment Riv_2, it extends approximately 20 km from the river mouth offshore. In general, the deflection of salinity plume has been observed at mid-latitude and the fresher water is dispersed farther south along the western boundary in experiment Riv_2 than in experiment Riv_1 (see Figure 41a and 41c for comparison). The surface temperature show a similar pattern as the surface salinity, in which the warm water plume has been stuck near the coastline and it has also been transported to the south along the western boundary.

The results from experiments Riv_1 and Riv_2 demonstrate the distribution of surface salinity and surface temperature caused by tides, surface heat fluxes and the Coriolis effect, but these experiments neglect the impact of wind forcing at the surface; wind forcing has major impact on the circulation in the UGOT even without rivers (see Section 5.1). Hence, experiment Riv_3 and Riv_4 are performed in order to examine how the wind stresses might influence the salinity and temperature distribution in the UGOT when river plumes exist.

During the months of June and September, winds in the range of 5–10 m s$^{-1}$ are mainly prevailed from the southwest direction. Hence the piling up of water near the northeast corner is expected. Note that Bang Pakong River is situated
at the northeast corner of the model domain, so the river plume is affected by the winds. Figure 42 shows the salinity and temperature distributions when wind stresses are included (experiment Riv_3 and Riv_4). Unlike experiments Riv_1 and Riv_2, the distribution of surface salinity and surface temperature (Figure 42) are clearly influenced by winds more than by tides or by the Coriolis effect. Both freshwater and warm water plumes are pushed to the left by the winds in the direction that water has piled up (clockwise direction). The impact of the winds has forced the surface water, in general to circulate in the anti-cyclonic fashion. Therefore the transports in/out of the model domain of salinity and temperature are expected along the western/eastern boundary (Figure 42).

For experiments Riv_3 and Riv_4, the extension of the freshwater plume from the Chao Phraya River is less clear because the winds mix the surface plume with underlying waters, so the symmetric shape of the plume is destroyed. The winds drag the near surface water to the right hand side of the wind direction. This leads to the plume bends to the left hand side (Figure 42a and 42c). The effect of the Coriolis parameter when wind is in action is not significant. With smaller Coriolis, the plume stays a little closer to the coast. Consider the distance from the river mouth of the 20 ppt salinity contour line in experiment Riv_4, the plume extends farther along the east coast than in experiment Riv_3. In fact, for experiments Riv_1 and Riv_2 (without winds) the result is opposite. The distance of the 20 ppt salinity contour line from the river mouth of experiments Riv_2 and Riv_4 are similar because the plumes are pressed at approximately the same latitude (13.4°N; see Figure 41c and 42c). Another interesting result experiment Riv_3 and experiment Riv_4 is that the river plume from Mae Klong River, which is the second largest river in the model domain has completely disappeared. Similar result is also observed at Tha Chin River. Therefore, strong wind mixing may destroy river plumes when discharges are small.
Figure 42. Similar to Figure 41, but (a) salinity distribution (September 2000) and (b) temperature distribution (June 2000) for experiment Riv.3. (c) Salinity distribution and (d) temperature distribution for experiment Riv.4. Daily wind velocities for September 2000 and June 2000 are shown on the bottom right corner of (c) and (d), respectively. The winds from both months are mainly blown from southwest direction.
5.3.2.2 Circulation features

Not only the monthly mean surface salinity (September) and mean surface temperature (June) are revealed in Figure 41 and 42, but also the monthly mean surface current velocities for both months. In the case where winds are not included (Riv_1 and Riv_2), the strong influence of the river discharges on the surface currents impacts the entire model domain. Near the river mouth, strong surface current velocities are present, especially at the Chao Phraya River (20–40 cm s$^{-1}$ on average for both months). Moreover, radial flows are observed.

At low latitude (experiment Riv_1), the mean surface currents are mainly toward the southward direction in both months (June and September). The jet of the surface flow is observed near the western boundary which corresponds to both plumes of lower salinity and warm water that have been transported southward along the western boundary. The existence of an anti-cyclonic gyre is found near the river plume front (boundary between fresh and saline water or warmer and cooler water). Since the Coriolis force was not the most important force at low latitudes, the velocities from the plume itself are the main force that pushed the gyre farther south as seen from the Chao Phraya River plume in Figure 41a.

Consider the gyre near the Chao Phraya River plume front (Figure 41a). After the flow of the gyre circulates back north, some parts of the gyre moves westward and reaches the flow of Tha Chin River plume. It then merges with the plume from Mae Klong River and forms the southward jet that has been observed along the western boundary. The flow pattern from Figure 41b (June) shows the same features as in Figure 41a (September), but the magnitude of the flow is smaller. These might be because the river discharge in June is smaller than in September (nearly 2 times smaller). Note that the magnitude of the residual flow in experiment Wnd_3 (only tides are included) is small compare to the residual flow in experiment Riv_1.
At mid-latitude (experiment Riv.2), the mean surface currents are similar to experiment Riv.1, except two features that are different. First, the front of the Chao Phraya River plume is formed closer to the river mouth than experiment Riv.1 causing the center of the anti-cyclonic gyre focuses farther north (at almost 45.3°N) and be situated on the left hand side of the river mouth. Second, the saltier and colder water from the open boundary tends to disperse from the eastern side farther north toward the center of the model domain, a feature which has not been observed as much in experiment Riv.1. The westward surface currents moves toward the southward jet along the western boundary are found almost along half of the southern part of the model domain, specifically for the month of September.

The inclusion of wind stresses in the model (experiment Riv.3 and Riv.4) resulted in a completely different circulation pattern than in experiments Riv.1 and Riv.2. In general, the mean surface current magnitude with wind is smaller than in experiment Riv.1 and Riv.2 (without wind), possibly because the wind blows opposite to the direction of the river flow. In average, the magnitude of mean surface currents for experiment Riv.3 and Riv.4 are in the range of 10–20 cm s⁻¹, which is about half of those in experiment Riv.1 and Riv.2. The clockwise circulation is observed as expected (Figure 42). Most of the mean surface current near the open boundary were moved southward. From Figure 42, the anti-cyclonic gyres in front of the river plume are not clearly seen, except for the freshwater plume in experiment Riv.4. The location of the gyre in front of the Chao Phraya River plume is approximately 100.8°E and 45.25°N (Figure 42c). Comparing with experiment Riv.2, the center of the gyre of experiment Riv.4 is more shifted to the south. One explanation is that during the southwest monsoon, water in the UGOT circulates clockwise and near the northeast corner current flows southward causing the center of the gyre is shifted to the south with the dominant currents.
The results from Figure 42 shown above, demonstrate patterns during the southwesterly winds, with water circulating clockwise. From the same experiments (Riv.3 and Riv.4) but different prevailing wind (northeasterly wind; November 2000) the results are very different (Figure 43). As expected, the counter-clockwise circulation in the UGOT dominates and the surface currents are stronger than the southwest monsoon (June and September 2000), because the winds are now blowing toward south in the same direction as the rivers. The plumes of fresher/warmer water are transported from river mouths along the northern coast and western coasts and then out of the model domain at the southwest boundary. In low latitude case (experiment Riv.3), the anti-cyclonic gyre is not clearly seen. The anti-cyclonic gyre has been observed at approximately 100.75°E and 13.35°N for mid-latitude case (experiment Riv.4). Even though the circulation feature during northeast monsoon is counter-clockwise, the anti-cyclonic gyre near the Chao Phraya River plume still exists.
Figure 43. Monthly mean surface salinity (left panel), surface temperature (right panel), and surface velocity during northeast monsoon (November 2000). Experiment Riv.3 and Riv.4 are shown on top and bottom panels, respectively. Note that both experiments are with wind cases.
5.3.3 Discussions

The distribution of salinity and temperature shown in the previous section are based on the surface patterns, in this section cross-section of both parameters are discussed to look at the vertical distribution of the plume.

As mentioned in Section 5.3.1, salinity and temperature profiles along 100.61°E longitude through the Chao Phraya River mouth are collected at 3-hour interval and then averaged to obtain the monthly mean. The plot of mean salinity and mean temperature cross-section for experiments Riv_1 and Riv_2 are shown in Figure 44. From Figure 44a and 44b (low latitude case), it is clear that the plumes of both salinity and temperature are pushed offshore. The saltier and colder water are entering the model domain from the south through the open boundary beneath the surface water. The plume of the surface water is found far to the south, showing how far from the river mouth the impact is felt. For the mid-latitude case (experiment Riv_2), the near surface plumes of fresh and warm waters are trapped near the coast (Figure 44c and 44d). They are similar to what have been seen from Figure 41a and 41c that the 20 ppt salinity contour line of the Chao Phraya River plume is pressed approximately 20 km and 15 km away from the river mouth, respectively. The farther offshore location of the 20 ppt salinity contour across the river mouth in experiment Riv_2 compared with experiment Riv_1 is clearly seen. The alongshore transport might play important role in maintaining the width of the 20 ppt salinity contour line. Near the 20 ppt salinity frontal line, the anti-cyclonic gyre is also observed (Figure 41a and 41c).
Figure 44. Monthly mean salinity and mean temperature cross-section along longitude 100.61°E. (a) salinity distribution (September 2000) and (b) temperature distribution (June 2000) for experiment Riv_1. (c) salinity distribution and (d) temperature distribution for experiment Riv_2. Note that colorbar for salinity and temperature are shown on the left and the right hand side, respectively.

The dynamics of river plumes can be characterized by the Rossby number, \( Ro_i = \frac{U}{(fW)} \), where \( U \) and \( W \) are mean velocity and width of the river, respectively. Fong and Geyer [2002] used the Rossby number to explain the dynamics of the alongshore transport of the anti-cyclonic gyre near the river mouth. According to their model result, the anti-cyclonic gyre is pushed close to the shoreline and strong alongshore transport exists in the case of low \( Ro_i \).

In experiments Riv_1 and Riv_2, mean velocity, \( U \), is almost the same (~30 cm s\(^{-1}\)) and the river width, \( W \), is approximately 1,300 m for both experiments. The Coriolis parameter \( f \) for experiment Riv_1 and Riv_2 are \( 3.2718 \times 10^{-5} \) and \( 1.0284 \times 10^{-4} \) s\(^{-1}\), respectively. Based on the numbers shown above, the Rossby numbers for
experiments Riv.1 and Riv.2 are approximately 7.05 and 2.24, respectively. This explains why the anti-cyclonic gyre in front of the Chao Phraya River plume in experiment Riv.2 is pressed closer to the coast than in experiment Riv.1 and why the width of the 20 ppt salinity contour is wider in experiment Riv.2 than experiment Riv.1 (the plume is transported by along the coast by stronger alongshore current).

The baroclinic Rossby radius of deformation, $L_R$, is a length scale that can also be used to explain the dynamics of fronts and eddies. It is defined as, $L_R = \sqrt{g' H / f}$, where $g' = g \Delta \rho / \rho_0$ is the reduced gravity and $H$ is depth scale of the plume. The plume density can be estimated from the average density ($\bar{\rho}$) of the water that has salinity less than 26 ppt (the large density gradient is observed) [Horner-Devine, 2009]. From the model results, the average density for experiment Riv.1 and Riv.2 are 1,008 and 1,006 kg m$^{-3}$, respectively. Based on the averaged density, one can estimate the reduced gravity as 0.19 and 0.21 m s$^{-2}$ for experiment Riv.1 and Riv.2, respectively where 1,026 kg m$^{-3}$ is used as the reference density. If $H \sim 3$ m (Figure 44), $L_R$ for experiments Riv.1 and Riv.2 then can be estimated to be $\sim 23.0$ and $\sim 7.8$ km, respectively. The small Rossby radius of deformation for experiment Riv.2 (with larger Coriolis) is smaller than the length of the plume shown in Figure 41c; the Coriolis parameter $f$ can play a role in bending the plume to the right and keeping it close to the coast, as well as influencing the anti-cyclonic gyre that is observed near the coast. On the other hand, $L_R$ of experiment Riv.1 (at low latitude) is large compared to the length of the plume shown in Figure 41a. Therefore, the plume is affected by $f$ and thus be more symmetric.

Based on laboratory experiment [Horner-Devine et al., 2006] and field observation [Horner-Devine, 2009], there are other two length scales that can be used to identify the dynamics of river plumes, which are the inertial radius, $L_i = U / f$, and the Rossby radius of deformation of the anti-cyclonic gyre, $L_a = (2Q_{riv} g' / f^3)^{0.25}$. Avicola and Huq [2003] show that $L_a$ is actually the width of the anti-cyclonic gyre of the plume.
The ratio of these two scales, $L' = L_t/L_a$, is used to explain the degree of the anti-cyclonic gyre (or bulge) that is often seen near the coast in front of the river mouth. The gyre is focused near the shoreline relative to its radius and large fraction of water discharge be transported away from the river mouth by coastal current when $L' \ll 1$. If $L' \rightarrow 1$, the gyre is pushed away from the shoreline due to the strength of the river discharge and relatively small amount of freshwater be transported away from the river mouth by coastal current, hence the large portion of the water discharge remains in the gyre [Horner-Devine, 2009; Horner-Devine et al., 2006]. According to the values and formula shown above, $L_t$, can be calculated as 9.2 and 2.9 km for experiment Riv.1 and Riv.2, respectively. Like $L_t$, $L_a$ can also be estimated and they were 11.7 and 5.1 km for experiment Riv.1 and Riv.2, respectively. Then the ratio of both two scales is $L' = L_t/L_a = 0.79$ for experiment Riv.1 and 0.57 for experiment Riv.2. Regarding to the criteria shown above, $L'$ in experiment Riv.1 are close to 1, then the gyre is pushed away offshore. These calculations agree with the results seen in Figure 41a. In experiment Riv.2, the calculated $L'$ is also close to 1 but it still less than the value of $L'$ for experiment Riv.1, so some of water discharge from river mouth might be transported by coastal current causing the width of the plume is more wider, specifically the Chao Phraya River plume (see Figure 41a).

Figure 41c shows the existence of two gyres near the Chao Phraya River plume of the mid-latitude case (experiment Riv.2). Anti-cyclonic gyre is the first one situated approximately 100.7°E and 45.3°N. The second gyre is the cyclonic gyre which is located at approximately at 100.5°E and 45.23°N. The salinity, temperature and velocity cross-sections in the north-south and east-west directions cut through the center of the gyres of experiment Riv.2 is shown in Figure 45 (Figure shown only the anti-cyclonic gyre). Consider $u-$ and $v-$velocity mid-depth upward near the center of the gyre (grey dashed line in Figure 45a and 45b). Both components flow in opposite direction near the center of the gyre. Therefore, the clockwise circulation
around the center is developed. At the center of the gyre, the deepening of the fresh/warm water may indicate a downwelling process. In contrast to the cyclonic gyre, the saline/cold water from below is raised to the surface (upwelling) due to the current velocity spin counter-clockwise around the center. The deepening of fresh/warm water shown in experiment Riv.2 is deeper than in experiment Riv.1 (Figure of experiment Riv.1 was not show here). This might be because the Coriolis parameter $f$ in high latitude plays a significant role causing velocity to spin faster with depth, following classical Ekman spiral dynamics. The faster the spin, the more stretch the water column. Along the right channel, the northward/southward flows present on top/bottom layer of the water column causing the saline/cold water and fresh/warm water is transported to the north and to the south, respectively (Figure 45b, 45d and 45f). The existence of cyclonic and anti-cyclonic gyres in the UGOT have been observed using cylindrical drogue in which the center has been set at 5 m below the surface and salinity and temperature measurements between 2004 and 2005 [Michida et al., 2006]. Unlike the result found in this study, their results show cyclonic gyre with high chlorophyll.a concentration situated in between the Chao Phraya and Bang Pakong rivers (see Figure 13 for location of the rivers) for both monsoon seasons and they conclude that freshwater discharges from both rivers helped maintain the gyre to rotate counter-clockwise in both seasons. The model is different from the observation might due to the analysis of observation is based on the 5–6 hours period.
Figure 45. Mean current velocity, salinity and temperature cross section along longitude 100.70°E (left panel) and latitude 45.30°N (right panel) of experiment Riv.2 (September 2000). The velocity component on the left panel showed only $u-$component (a), while on the right panel shown only $v-$component (b). (c) and (d) are for salinity and (e) and (f) are for temperature. The vertical grey dashed lines in each plot represent the place where both two cross sections cut through each other.
CHAPTER 6

SUMMARY AND CONCLUSIONS

The study includes two components, analysis of data, in particular sea level data in the GOT and the Andaman Sea, and numerical modeling of the circulation in the UGOT, so both regional and local processes are considered.

In Thailand both in the GOT and Andaman Sea, more than 20 tide gauge stations have been operating. Sixteen of them and other two from Malaysia are used in this study. In general, tides in the Thailand are in the range of 2-3 m, therefore the area can be classified as meso-tide. Tides in Thailand are found to have all three types of tides. Diurnal tides are found on the eastern side of the GOT and the whole lower GOT. Mixed mainly diurnal tides are found at Geting, the southernmost of the GOT. In the UGOT at the boundary between Sattahip and Hua Hin, mixed mainly diurnal tides are observed, while mixed mainly semi-diurnal tides are found at stations along the northern boundary. The semi-diurnal tides are only found in Andaman Sea. This means that the semi-diurnal tidal components, M2 and S2, are dominant in Andaman Sea, while diurnal tidal components, K1 and O1, are dominant in the GOT. In addition, tidal amplitudes, both semi-diurnal and diurnal components tend to increase toward the north.

Tidal residuals are affected by monsoonal winds and river discharge, especially stations near the river mouth. However, the residual found in the GOT is not the same as the one found in Andaman Sea, as the atmospheric forcings (pressure and wind) affect these differently.

The averaged relative rate of sea level rise in Thailand’s coast based on linear trend and HHT trend are $5.98 \pm 3.43$ and $6.53 \pm 3.8$ mm yr$^{-1}$, respectively. The faster rates ($> 10$ mm yr$^{-1}$) found at Phrachulachomkloa, Tha Chin, Mae Klong,
and Ban Lam stations are likely caused by significant land subsidence, with important consequences for the population of Bangkok and surrounded area. The vertical movement of land by seismic activity is also causing changes in sea level rise rates [Trisirisatayawong et al., 2011]. Satirapod et al. [2013] mentioned that vertical movement of land pre- and post-2004 Sumatra-Andaman Earthquake are uplift (~3 mm yr⁻¹) and emergence (~8 mm yr⁻¹), respectively. Therefore estimating rate of sea level rise in Thailand must take vertical motion into consideration too.

A seasonal sea level cycle in Thailand is studied following Tsimpolis and Woodworth [1994]. Like the residual, seasonal sea level cycle in the GOT and Andaman Sea is totally different. The seasonal sea level cycle in the GOT is annually dominant, while in Andaman Sea is the combination between annual and semi-annual components. The amplitudes of annual and semi-annual components in the GOT are in the range of 120 to 240 cm and 20 to 50 cm, respectively. In Andaman Sea, approximated 100 and 60 cm, respectively, are found for annual and semi-annual tidal components. Peaks of maximum annual component are approximately at the beginning of the year and 4.5 months earlier for stations in the GOT and Andaman Sea, respectively. For the peaks of semi-annual component, both places are on average one and a half month earlier. The comparison between mean sea level anomaly and sea level pressure shows positive and negative correlations for station in the GOT and Andaman Sea, respectively.

A simulation of the UGOT tidal model demonstrates how well the model can capture the observed water level and current velocity. Root-mean square error, RMSE, of water level is found in the range of 12 to 20 cm, while R² is greater than 0.90 at all stations. For the current velocity, RMSE of v-velocity is approximately 4 cm s⁻¹ with R² about 0.90. In addition, tidal current in the UGOT mainly flow in north–south direction. Co-phase and co-range lines for 4 major tidal components, M2, S2, K1, and O1, and other 4 minor tidal components, K2, N2, P1, and Q1, show
almost the same pattern. Tidal amplitudes increase toward the head of the UGOT, while tidal phases propagate counter-clockwise. Because the size of the UGOT is small compared to the Rossby radius of deformation, there is no amphidromic point.

The inclusion of monsoonal winds in the UGOT model generally causes the residual flow to circulate clockwise/counter-clockwise during the southwest/northeast monsoon. A small feature of clockwise/counter-clockwise gyre is also found near the northeast corner during the northeast/southwest monsoon. The piling up of water is also observed near the northeast/southwest corner during the southwest/northeast monsoon. Even though the UGOT is situated at low latitudes, the classical Ekman spiral is clearly present in the surface Ekman layer. East-west cross-section of $v-$velocity component shows flows are with the direction of the wind in the shallow regions and against wind at the surface layer of the deeper channel (estuarine-like circulation). This result is comparable well with a simple wind-driven analytical estuarine model [Winant, 2010].

A 4-year long simulation during 1997–2000, when spatially constant and temporally variation surface heat fluxes were included has been performed and the results showed the intrusion of cold water from the open boundary along the east coast and the dispersion of warm water towards south along the west coast during the northeast monsoon and vice versa during the southwest monsoon. This result agrees with a satellite image and the circulation pattern caused by the winds. The mean sea surface temperature difference between observed and model is approximately 1°C.

When the river discharges were included in the model, it is found that the freshwater dispersed as a coastal jet along the western boundary. The dispersion of the plume farther south along the coast at mid-latitudes compared to a low latitude case is because of the impact of the Coriolis effect. The generation of anti-cyclonic gyres near river plumes has been observed both monsoon seasons in mid-latitude case. Without wind, the river plumes have symmetrical shapes, but they are destroyed by
the strength of the winds.

The main conclusion of the study is that different forcings affect different scales in the UGOT. High-frequency (hourly-monthly) variations are dominated by tides, seasonal variations are most affected by the monsoon pattern of winds, and inter-annual variations are affected by the El Niño/La Niña pattern of the Pacific Ocean. The numerical model developed here will help to better understand the forcing mechanisms and address future environmental problems in the region.
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Publications


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