Study of Ocean Climate Variability (1959-2002) in the Eastern Indian Ocean, Java Sea and Sunda Strait Using the HAMburg Shelf Ocean Model

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Abstract

The Indonesian waters, located in the tropical area between the Pacific and Indian Oceans, comprise shelf and deep sea areas with their special characteristics. The Java Sea, as one of the shallow waters areas in Indonesia, has an average depth of about 40m, low salinity and high temperature. South of Java is the deep eastern Indian Ocean which is characterized by high salinity and low temperature. Both water masses are interacting and mixing through the Sunda Strait. Although the Sunda Strait is a small passage, its role in influencing the ocean climate variability is quite important and will be discussed as the main topic in this study.

The **HAM**burg Shelf Ocean Model (HAMSOM) is used to simulate the long period ocean-climate dynamics in the Java Sea and the eastern Indian Ocean. Numerous simulations and boundary treatments are performed to get results that best agree with the observations. Forcing data from 44 years (1959-2002) NCEP reanalysis are used in the simulation to analyze the variability of current circulation, water mass transport, upwelling south of Java, and the effect of ENSO (in the Pacific Ocean) and DME (in the Indian Ocean) to this area, as well as their interaction.

The seasonal variation in the Java Sea is mainly influenced by the monsoon. In January, representing the NW monsoon situation, water from the Southeast China Sea is transported to the Java Sea by (+)2.1Sv and flows out to the eastern part of the Java Sea and the Sunda Strait by (-)1.6Sv and (-)0.5Sv, respectively. In the contrary, in August (representing the SE monsoon situation), the total inflow to the Java Sea from the eastern part of the Java Sea is 1.0Sv and the water flows out to the Southeast China Sea and the Sunda Strait by (-)0.3Sv and (-)0.7Sv respectively. The SST is increased from $28^{\circ}C$ during the NW monsoon to $28.5 - 29^{\circ}C$ during the SE monsoon while the SSS is decreased from 32.6psu to 32.2psu. During the whole year, the water mass is transported from the Java Sea to the Indian Ocean through the Sunda Strait with variations of transport between 0.48Sv (minimum) in December and 0.72Sv (maximum) in August/September. The zonal wind during the SE monsoon is the important factor for triggering and propagating the upwelling south of Java. Normally, the upwelling causes the upwards of temperature lower than $24^{\circ}C$ (the peak in August) and salinity higher than 34.4psu (the peak in October).

The DME and ENSO have influence significantly the variability in the study area. Nevertheless, the strengths of their influence are controlled by the strength of the zonal wind in the eastern Indian Ocean, especially during the SE monsoon. The temperature decreases (increases) during the El Niño (La Niña) years when the zonal wind is stronger (weaker) and reversely the salinity increases (decreases) during the El Niño (La Niña). The strong influence of the DME and ENSO can be identified by the SST/SSS anomaly in the study area when they have the same phase as the strengthening zonal wind. Likewise, the strength of upwelling south of Java depends on the strength of SE wind along the south of Java and its interaction with DME and ENSO events. Under normal conditions, the upwelling causes negative SST and positive SSS anomalies $0.3-0.6^{\circ}C$ and 0.2-0.3psu, respectively. In the El Niño (La Niña) years, the anomaly of SST and SSS become lower than $-1.0^{\circ}C$ and higher than 0.3psu (+ $1.0^{\circ}C$ and -0.3psu), respectively.

During the normal years, the variation of SST and SSS anomalies in the Java Sea are about $\pm 0.5^{\circ}C$ and ± 0.2 psu, respectively. The SST decrease (SSS increase) during the DME and El Niño events because of the transport of water mass with cooler SST (higher SSS) from the eastern Java Sea, while during the La Niña events, the SST increase (SSS decrease) because of the transport of water mass with warmer SST (lower SSS) from the eastern Java Sea. The water mass is still transported through the Sunda Strait from the Java Sea, increased by 0.15-0.2Sv during the El Niño and DME years and decreased by 0.1Sv during the La Niña years.

Chapter 1 Introduction

The Indonesian archipelago is located at $6^{\circ}N$ -11°S and 94°-141°E. It lies between the Indian and Pacific Oceans and the continents of Asia and Australia. It encompasses the equator region in southeast Asia and consists of 17.508 islands (Indonesian Hydro-Oceanographic Office, 2003), with five main islands with an area bigger than 132.000 km^2 . These main islands are Kalimantan, Sumatra, Sulawesi, Irian Jaya, and Java. The total population is approximately 206 million (Population Census, 2000), most of whom concentrated on these five main islands.

Indonesia has many active volcanoes. Normally after the explosion of a volcano the land will be fertile and the agriculture will be blooming. There are two volcano lines crossing Indonesia. The first volcano line is along the western ridge of Sumatra and the southern ridge of Java, Bali, and Lombok, which meets with the second volcano line in Sulawesi and Irian Islands. One of the volcanos is the Krakatau, a volcano that lies in the Sunda Strait. The eruption and collapse of Krakatau's caldera in 1883 produced one of the largest natural explosions of the earth ever recorded. Its volcanic explosivity index (VEI) reached 6, whereby the VEI scale ranges from 0 to 8 (*Tilling et al.*, 1984), and the eruption destroyed most of Krakatau Island, leaving only a remnant (*Simkin* and *Silbert*, 1995). Since 1927, small eruptions have been frequent and have constructed a new island (*Decker* and *Hadikusumo*, 1960), Anak Krakatau (the Child of Krakatau). This episode



Figure 1.1: Position of Indonesia and area of the study

The Indonesian archipelago is located at $6^{\circ}N$ -11°S and $94^{\circ}E$ -141°E, between the Pacific Ocean in the east and Indian Ocean in the west, Asia in the northwest and Australia in the southeast. The red box indicate the area of the study

of Anak Krakatau's activity, which began in December of 1959, ended in 1963. Anak Krakatau had ever since at least nine episodes of volcanic activities, most of them lasted less than one year (*Simkin* and *Silbert*, 1995).

Based on the world's geological plate in the *Welt Atlas mit großem farbigen Länderlexikon*, the Indonesian archipelago spreads over on four plates. These are the Eurasian plate, where the islands of Sumatra, Kalimantan, Java, and Sulawesi and also part of the Maluku Islands are located, the Indian-Australian plate in the southwest of Sumatra, south of Java up to Irian Island, the Pacific plate in the north of Irian, and the Philippine plate in the north of the Maluku archipelago. They collide repeatedly. The Philippine plate moves to the west and together with the Pacific plate subducts the Eurasian plate to the south. The Indian-Australian plate moves to the north and subducts the Eurasian plate. The earthquake zones are lying along these subduction lines, and a lot of trenches are found along the western coast of Sumatra up to the southern coast of Java (see Figure 1.1).

Indonesia has many shelf seas with less than 200m depth and deep areas with more than 3000m depth (Figure 1.1) with a lot of natural wealth such as minerals, fuels, natural gas and also a large biodiversity (fish). The annual fish production in Indonesian waters is about 4.8 million tons/year and contributes to the country's income more than US\$ 2,05 billion ¹. The majority of the fish production is exported. Normally during the boreal summer the fish production is increased because of the upwelling, where the cool water rich of nutrients is transported up from the deep ocean to the surface areas near the coast. Most coastal areas in the Indonesian waters are upwelling zones. The strong currents in Indonesian waters could facilitate the dispersal of marine larva (*Barber et al.*, 2000) and sea surface current (SSC) patterns could be crucial for predicting ecological and genetic connections among threatened reef populations (*Roberts*, 1997). Therefore information on ocean dynamics and climate impact are also important to increasing the fishery product.

The specific geographical position of Indonesia influences the characteristics of climate and ocean dynamics. The inter-tropical convergence zone (ITCZ) shifts from the northern (July) to the southern (January) hemisphere crossing the equatorial line. Therefore, the difference of air pressure between the continents of Asia and Australia that shifts every 6 months causes the monsoonal winds over Indonesia and drives the sea surface current (SSC) patterns. The latter influences the sea surface temperature (SST) as well as the sea surface salinity (SSS). During the boreal summer, Indonesia exhibits a dry season and during the boreal winter a rainy season. The difference of the sea surface level (SSL) between the Pacific and the Indian Oceans causes the currents to flow from the Pacific to the Indian Ocean through Indonesian waters, which is usually called the Indonesian Throughflow or Arus Lintas Indonesia (Arlindo) in the Indonesian language. There are already some interesting researches on Arlindo (e.g. *Susanto et al.*, 2001; *Meyers et al.*, 1999; *Potemra*, 1999).

Several researchers have conducted their studies in the Indonesian waters. *Wyrtki* (1961) has analysed the properties of water mass and ocean dynamics of South East Asian waters base on survey data during 1959-1961. *Mihardja* (1991) and *Stawarz* (1994) using a numerical model have calculated the energy budget of the tides and its distribution in Indonesian waters. *Hendiarti* (1992) has identified the upwelling area using only one day data from the NOAA satellite of the southern coast of Java. *Rizal* (1994) has analysed the dynamics of the Malacca Strait and *Potemra* (1999) has performed a numerical model simulation in order to analyse the Indonesian Troughflow and the upwelling zones in the southwestern coast of Sumatra and Java and their variations, but he did not include the Sunda Strait area. *You* (1999) analysed the distribution of water mass in the Indian Ocean. *Hendiarti* (2000) has done a preliminary study of the water mass distribution around the Sunda Strait based on the data taken at the German-Indonesia "pre JIGSE 2000" cruise in October-November 2000 (these data will be taken also for validation of the numerical model results in our

¹source: http://www.bi.go.id/sipuk/lm/ind/ikan_laut/pemasaran.htm





The study area comprises two different characters of seawater. The Java Sea is a shallow waters and the eastern Indian Ocean is a deep sea. The Sunda Strait is a passage between them. The dots in the figure are locations of river runoff, which are used in this study (see also table 3.2).

study). The transport of upwelling during the southeast monsoon had been discussed by *Susanto et al.*(2001). The variability of meteorology-oceanography parameters and their relationship with ENSO and the Dipole Mode over Indonesia had been analysed by *Setiawan* (2003). *Wannasingha et al.* (2003) ² has simulated the Indonesian Throughflow with the Ocean Circulation and Climate Advanced Model (OCCAM) and calculated the annual transport of the north and south routes, where the Pacific water mass flows into the Indian Ocean.

One of the interesting areas in the Indonesian waters is the Sunda Strait, which is between the islands of Sumatra and Java that connects the Java Sea and the Indian Ocean (see red box in Figure 1.1 and also Figure 1.2). Unfortunately, studies in this area are still rare especially on the dynamics of water transport and exchange, mixing condition from/to the shelf water to/from the deep sea, and also on the variability between the shelf and deep seas. Our hypothesis is that the water transport through the Sunda Strait depends also on the atmospheric and the ocean dynamics of the Indian and the Pacific Ocean. Due to its narrowness only a small water exchange is possible between the Sunda Shelf (Java Sea) and the Indian Ocean through the Sunda Strait.

The analyses, which are given in the following sections, are mainly focused on the region of the model domain between $0.5^{\circ}S$ - $10.5^{\circ}S$ and 102.5° - $114.5^{\circ}E$ (see Figure 1.2). The northern boundary

²The poster presented at the European Geophysical Society (EGS) meeting in Nice, 2003

is the Karimata Strait, which is located between the eastern coast of Sumatra and the western coast of Kalimatan. The eastern boundary is located between the southern coast of Kalimantan and northern coast of East Java, and the other boundary is in the Indian Ocean at $10.5^{\circ}S$ latitude.

The main objective of this work is to study the ocean condition and its variability of the eastern Indian Ocean, the Java Sea and the Sunda Strait during the period 1959-2002 and to identify where the potential areas of upwelling are located and how the monsoon and its variability effects the upwelling south of Java. Furthermore, the analysis of this study can be used by the government and policy makers for planning and decision making.

Several simulations are done using the HAMburg Shelf Ocean Model (HAMSOM). In the first simulation (S1), HAMSOM is run in the barotropic mode only using the tides as a driving force. This simulation is performed to analyse the tidal conditions in the study area. The second simulation (S2) is done running HAMSOM in the baroclinic mode, in which the model is driven by tides, monthly climatological wind stress data from *Hellerman* (1983) and temperature and salinity data from the World Ocean Atlas (1998). In general, the purpose of S1 and S2 simulation is to check the performance of HAMSOM in the study area. In the third simulation (S3), HAMSOM is also set in baroclinic mode. The driving forces used are tides, 6-hourly atmospheric reanalysis fields from the Norld Ocean Atlas (1998). The model is run for only 3 years (1997-1999) and the results are validated by means of results from previous researches and observational data. The fourth simulation (S4) is the same as S3 but for a long period of 44 years (1959-2002). S4 is performed with 2 scenarios called Run IV.A and IV.B. Run IV.A uses zero mean sea level (MSL) along the whole open boundaries while the Run IV.B uses different constant MSL data at each open boundary taken from *Wyrtki* (1961).

The work is organized into 8 chapters. Chapter 2 describes the general ocean dynamics of Indonesian waters, characteristics of the shallow and deep waters in the tropical area, especially the Java Sea and the Indian Ocean, and the conditions of the Sunda Strait. Chapter 3 discusses the data used for the model input and validations. Chapter 4 describes the HAMSOM, its numerical treatments in the study area, and the model design. The detailled formulations and list of symbols, which are used in the HAMSOM description, are given in the Appendix A, B, and C. In Chapter 5, the model simulations and their results as well as the comparison with previous results and observed data are presented. The analysis of mean sea level differences along the boundaries will be explained as well. The analysis of the climatologycal ocean condition will be discussed in Chapter 6, whereas Chapter 7 explains the ocean variability during 1959 until 2002. Discussions on the latter two chapters are based on the model results. Finally, the conclusions are given in Chapter 8.

Chapter 2 General Dynamics of Indonesian Waters

Indonesia comprises more than ten thousand islands and lies at the equator. The climate is influenced dominantly by the monsoon, and also by the sea-air-land interactions. Indonesia is located at the action centers of the north-south (meridional) circulation, known as the Hadley Circulation, and the west-east (zonal) circulation, known as the Walker Circulation.

2.1 The Monsoon

Because of the tilt of the earth's axis on its orbits around the sun, the northern and southern hemispheres receive different solar radiation. Half of the year, the northern hemisphere receives more solar radiation than the southern hemisphere and reversely in the other half year. That causes surface air pressure differences between the northern and southern hemisphere, furthermore, changes of the prevailing wind directions occur which are called seasons. The term monsoon itself is traced back to an Arabic root meaning 'season' (*Hastenrath*, 1985). *Ramage*, 1971 (in *Hastenrath*, 1985) proposes four criteria to delineate the monsoon regions:

- 1. the prevailing wind directions shifts by at least 120 degrees in January and July
- 2. the average frequency of prevailing wind directions in January and July exceeds 40%
- 3. the mean resultant wind in at least one of the months exceeds 3m/s



Figure 2.1: Delineation of the World's monsoon region. (From *Ramage*, 1971 in *Hastenrath*, 1985) Hatched areas meet simultaneously the wind criteria (1) to (3), while heavy lines mark the Northern limit of the region within the Northern hemisphere where the cyclone/anticyclone criterion (4) is satisfied. Rectangle encloses the monsoon region. Indonesia is in the monsoon region with wind criteria (1) to (3)

4. fewer than one cyclone-anticyclone alternation occurs every two years in either month in a 5 degree latitude-longitude rectangle.

Under normal conditions, from December to March, the air pressure in Asia is higher than that of Australia. The northeasterly wind blows north of the equator, it is known as northeast monsoon and turns southeastward in the south of equator such as over the Java Sea. Therefore in Indonesia this monsoon is known as the northwest (NW) monsoon from December to March, or boreal winter. Normally during the NW monsoon Indonesia exhibits a rainy season, the highest peak occurring in January. The wind blows southeastward and eastward near the equator. In the southern hemisphere from $10^{\circ}S$ the wind blows to the north and then turns to the east. *Wyrtki* (1961) explained that the equatorial trough lies over the Indian Ocean around $10^{\circ}S$, in the southern part of which the southeast trades are found.

April to May is the transition period. In May, the system of the northeast wind over the South China Sea and the Philippines decreases, and the south monsoon predominates over the Indonesian archipelago. In the south of the equator, the southeast (SE) wind blows and enters to the Indian Ocean as the southeast trades. At the equator, south wind prevails, whereas in the north of the equator southwest winds dominate.

During the SE monsoon or the boreal summer, normally from June to September, the prevailing wind blows northwestward from Australia, which has a higher air pressure than Asia. In July-August, the SE monsoon reaches its peak (fully developed). Indonesia has a dry season. October and November is the transition period from the SE to the NW monsoon.

Based on the variance of the Indonesian rainfall, *Aldrian* and *Susanto* (2003) have divided the climate conditions of Indonesia into three climate regions. Region A covers south and central Indonesia from south Sumatra to Timor Island, parts of Kalimantan, parts of Sulawesi, and parts of Irian Jaya. Region B is located in northwestern Indonesia and covers the northern part of Sumatra and the northwestern part of Kalimantan. Region C covers Maluku and parts of Sulawesi (close to the western Pacific region). The mean annual rainfall cycle of each region is different. Region A has one peak and one trough and experiences strong influences of two monsoons, namely the wet northwest (NW) monsoon from November to March and the dry southeast (SE) monsoon from May to September. Region B has two peaks, in October/November and in March to May. Those two peaks are associated with the southward and northward movement of the inter-tropical convergence zone (ITCZ). Region C has one peak in June/July and one trough in November to February.

2.2 El Niño Southern Oscillation (ENSO)

At the sea surface, interactions between ocean and atmosphere occur. The air-sea interaction and the ocean dynamics in the Pacific and the Indian Ocean influence the conditions in the Indonesian waters. Therefore this section is dedicated to the descriptions of the ocean dynamics of the Pacific and the Indian Oceans.

Under normal conditions, in the Pacific Ocean the westerly trade winds blow. These winds pile up warm surface water in the West Pacific, so that the sea surface is about 0.5m higher close to Indonesia compared to Ecuador. The sea surface temperature is about $8^{\circ}C$ higher in the west, with cool temperatures off South America, due to an upwelling of cold water from deeper levels (Figure 2.2 b) ¹. This cold water is rich of nutrients and supports high levels of primary productivity, diverse marine ecosystems, and major fisheries around the western coast of South America.

¹http://www.pmel.noaa.gov/tao/elnino/el - nino - story.html



Figure 2.2: Sea Surface Temperature during (a) the La Niña Condition 1998, (b) Normal Condition 1993, and (c) the El Niño Condition 1997

The conditions were identified by Pacific Marine Environmental Laboratory (PMEL) - NOAA in the Eastern Indonesian waters in the Pacific Ocean

The reduction or intensification of air-sea interactions eventually causes the abnormal conditions. The abnormal conditions, known as El Niño and La Niña events, can change the atmospheric and ocean dynamics.

El-Niño is a phenomenon which is characterized by unusual warm ocean temperatures in the Equatorial Pacific Ocean. The trade winds relax in the central and western Pacific leading to a depression of the thermocline in the eastern Pacific, and an elevation of the thermocline in the west. The SST gets warm in this period, for example during the El-Niño in 1997 (see Figure 2.2 c). This condition reduces the efficiency of upwelling to cool the surface and cuts off the supply of nutrient rich water to the euphotic zone. The result is a rise of the SST and a drastic decline of the primary productivity, the latter adversely affected higher trophic levels of the food chain, including commercial fisheries in this region. Rainfall follows the warm water eastward, with associated flooding in Peru and drought in Indonesia and Australia. Over the Indonesian waters the surface temperature is lower than for normal conditions especially in the western part. The upwelling area in the south of Java during the SE monsoon spreads from east to west in the southwest of Sumatra. The eastward displacement of the atmospheric heat source overlaying the warmest water results in large changes in the global atmospheric circulation, which in turn forces changes in the weather over regions. Therefore, Indonesia is drier and the precipitation is lower than normal.

The opposite condition of El Niño is called La Niña, which is characterized by unusual cold ocean temperatures, called the 'cold tongue', in the Equatorial Pacific Ocean. The strong westward wind influences the SST in the equatorial Pacific. The upwelling gets stronger during the La Niña event and causes an extention of cold water and rich nutrient areas in the eastern Pacific (Figure 2.2 a).

Years	Jan Feb Mar	Apr May Jun	Jul Aug Sep	Oct Nov Dec
1958	W+	W	W-	W-
1959	W-	Ν	Ν	Ν
1960	Ν	Ν	Ν	Ν
1961	Ν	Ν	Ν	Ν
1962	Ν	Ν	Ν	Ν
1963	Ν	Ν	W-	W
1964	Ν	Ν	C-	С
1965	C-	Ν	W	\mathbf{W} +
1966	W	W-	W-	Ν
1967	Ν	Ν	Ν	Ν
1968	Ν	Ν	Ν	W-
1969	W	W-	W-	W-
1970	W-	Ν	Ν	С
1971	С	C-	C-	C-
1972	Ν	W-	W	\mathbf{W} +
1973	W	Ν	C-	C+
1974	C+	С	C-	C-
1975	C-	C-	С	C+
1976	С	Ν	Ν	W-
1977	Ν	Ν	Ν	W-
1978	W-	Ν	Ν	Ν
1979	Ν	Ν	Ν	Ν
1980	W-	Ν	Ν	Ν
1981	Ν	Ν	Ν	Ν
1982	Ν	W-	W	\mathbf{W} +
1983	W+	W	Ν	C-
1984	C-	C-	Ν	C-
1985	C-	C-	Ν	Ν
1986	Ν	Ν	W-	W
1987	W	W	W+	W
1988	W-	Ν	C-	C+
1989	C+	C-	Ν	Ν
1990	Ν	Ν	W-	W-
1991	W-	W-	W	W
1992	W+	W+	W-	W-
1993	W-	W	W	W-
1994	Ν	Ν	W	W
1995	W	Ν	Ν	C-
1996	C-	Ν	Ν	Ν
1997	Ν	W	W+	W+
1998	W+	W	C-	С
1999	C+	С	C-	С
2000	С	C-	Ν	C-
2001	C-	Ν	Ν	Ν
2002	Ν	W-	W	W

Table 2.1: Cold (C) and warm (W) episodes by seasonWeak periods are designated as C- or W-, moderate strength periods as C or W,
strong periods as W+ or C+, and neutral periods as N.

source :

http://www.cpc.ncep.noaa.gov/products/analysismonitoring/ensostuff/ensoyears.html The cold and warm episodes in this table are based on the intensity of El Niño and La Niña along the equator of the tropical Pacific Ocean in the region 150°W to the date line As a consequence, the fishery productions rise in this area.

The monsoon, ENSO, and the complex bathymetry interplay and influence the air-sea exchange and the inter-ocean throughflow within the Indonesian Seas (*Potemra*, 1999; *Webster et al.*, 1999). During the SE monsoon, the southeasterly wind from Australia generates upwelling along the Java-Sumatra south coasts. The upwelling is mainly forced both locally by alongshore winds associated with the SE monsoon and remotely by atmosphere-ocean circulation associated with ENSO (*Susanto et al.*, 2001).

El Niño and La Niña are the opposite phases of the El Niño-Southern Oscillation (ENSO) cycle, with La Niña sometimes referred to the cold phase and El Niño to the warm phase. The cold and warm episodes had been investigated by season by NOAA as shown in Table 2.1. The sign W+, W, W-, C+, C, C-, and N are based on the Pacific Ocean conditions. The ENSO years used in the analysis of simulation results will be referred to this table.

2.3 Dipole Mode Event (DME)

The DME is characterized by an atmosphere and ocean interaction in the Indian Ocean, when an abnormal decrease of the SST in the eastern tropical Indian Ocean around Sumatra ($90^{\circ} - 110^{\circ}E$ and $10^{\circ}S$ - equator) and an abnormal rise of SST in the western tropical Indian Ocean around Africa ($50^{\circ} - 70^{\circ}E$, $10^{\circ}N - 10^{\circ}S$) occur. The difference of the SST between the western and eastern Indian Ocean is known as Dipole Mode Index (DMI). The first analysis of DME is presented by *Saji et al.* (1999). In order to investigate the cause of the unusual hot summer of 1994 in Japan, ocean and atmospheric data, such as sea surface temperature, winds, and precipitation, for the past forty years in the Indian Ocean, have been examined. The result revealed that the easterly winds raise the oceanic temperature in the west, the drought in the eastern Indian Ocean around Indonesia, and the flood in the area from India to East Africa. Since this phenomenon greatly influences the climate of the coastal nations around the Indian Ocean, Australia and Far Eastern Asia including Japan, it will have a great impact upon the social economy of this densely populated area. The predictability of this phenomenon will be a great importance for the mitigation of floods, droughts, and intense heat waves.

The relationship between the Indian monsoon and the Pacific El Niño is until recently unknown. *Saji et al.* (1999) found that the correlation between DMI and NINO3 ² is less than 35% and the correlation between DMI and the equatorial wind variability over the eastern equatorial Indian Ocean is more than 60%. Because of the research development in the Indian Ocean the mechanism that causes the complementary impact of these two main characteristic of the Indian monsoon rainfall has now been clarified (*Yamagata et al.*, 2001). *Grodsky and Carton* (2001) analysed the key momentum balance between wind-induced momentum flux and the pressure gradient force as well as the important role of horizontal temperature advection in the mixed layer heat response and argued that the eastern equatorial Indian Ocean is intimately related to ENSO forcing at least during the decade of 1990s.

Li et al. (2003) explained that there are four fundamental differences in the sea-air interaction between the tropical Pacific and Indian Oceans based on observational analysis and physical reasoning. The first difference is represented by a strong contrast of a zonal cloud-SST phase relationship between the warm and cool oceans. The second difference arises from the reversal of the basic zonal wind and the titling of the thermocline, which leads to distinctive effects of ocean waves. The third difference lies in the existence of the Asian monsoon and its adjacent oceans. The fourth difference is that the southeastern Indian Ocean is a region where a positive atmospheric-ocean thermodynamic feedback exists in boreal summer. They concluded that the El Niño impacts

²NINO3 covered area between $5^{o}S - 5^{o}N$ and $150^{o}W - 90^{o}W$ in the eastern equatorial Pacific.

the DME in their model through three processes; firstly, through the strength of the South Asian monsoon, which further alters the cross-equatorial wind along the coast of Africa and the zonal wind in the central equatorial Indian Ocean; secondly, through the anomalous Walker cell over the Indian Ocean that changes the SST in the Maritime Continent; and thirdly, through the intensity of convection over the Maritime Continent, which then induces anomalous along-coastal winds off Sumatra (as a direct Rossby wave response to an anomalous heat source).

Saji et al. (1999) ³ identified that the DME occurred in 1961, 1967, 1972, 1982, 1994 and 1997. The cool anomaly of SST migrates eastward of the Indian Ocean along the Java Sea and west of Sumatra to the equator, while the SST in the western Indian Ocean warms up. *Prasad* and *McClean* (2004) indicated that large interannual SST variabilities in the western Indian Ocean are forced locally by changes of monsoon winds.

2.4 Ocean Characteristics of the Study Area

Variations of the atmospheric circulations cause corresponding variations of the oceanic circulations. As the monsoon changes its direction twice per year, the oceanic circulation also reversely changes its direction in most areas. Other atmospheric phenomena such as ENSO influence the ocean conditions as well. *Setiawan* (2003) indicated that the ENSO and DME have a close relationship and their impacts to the Maritime Continent of Indonesia depend on the Asia- Australia monsoon system which is the dominant factor of this region.

2.4.1 The Deep Waters in the Eastern Indian Ocean

The shelf south of Sumatra and Java has a big slope towards the deep Indian Ocean. There are two trenches; the Sumatra trench, with depth up to 3000m in the west of Sumatra and the Sunda trench with depth up to 2000m in the south of Java. In the southern part of the Indian Ocean, the depth reaches more than 7000m.

Characteristics of the water mass in the eastern Indian Ocean are higher salinity and lower temperature in comparison to the Java Sea. At the surface the water is warm and the annual variation is normally small (*Wyrtki*, 1961). The vertical distribution of the temperature is more variable than the horizontal one. The temperature in the surface layer is more than $26^{\circ}C$, whereas in the deepest layer, it is about $1^{\circ}C$. The salinity is about 34psu at the surface and 35psu at the bottom.

The map of the regional current of Southeast Asia (Figure 2.3) shows that the surface current is mainly driven by the surface wind. *Quadfasel* and *Cresswell* (1992) have analysed the cruises data and have shown that the South Java Current varied seasonally due to the changing monsoon winds and the variations of the freshwater flux from the Indonesian archipelago.

2.4.2 The Shallow Waters of the Java Sea

The Java Sea lies on the Sunda Shelf with an average depth of about 40m and is a semi-closed sea located between Sumatra in the west, Kalimantan in the north, and Java in the south. The exchange of water mass from/to the Java Sea is through the Karimata Strait, the eastern entrances part of the Java Sea and the Sunda Strait. In the northern part, the Karimata Strait is the passage between the Java and the South China Sea. In the eastern Java Sea there is a passage between the Java Sea and the Indonesian waters. In the southern part, the Sunda Strait is a passage between the Java Sea and the Indonesian waters.

³Online in internet http://www.jamstec.go.jp/frsgc/research/d1/contents/paper/sap1.html



Figure 2.3: Global Sea Surface Circulation in the Southeast Asian Waters (Wyrtki, 1961)

The Java Sea is situated within the monsoon regime and is thereby strongly influenced by the semi-annual reversing between the North West (NW) and South East (SE) monsoon (*Pohlmann*, 1987). During the NW monsoon the surface current flows to the east and during the SE monsoon to the west and a part of it flows into the Indian Ocean through the Sunda Strait (see Figure 2.3).

As typical for shallow waters, the vertical temperature and salinity in the Java Sea are well mixed. In some places, the freshwater discharging from several big rivers into the Java Sea influence the temperature and salinity, especially during the rainy season. In general the density of the Java Sea is always below $\sigma_t = 22.0$ (*Wyrtki*, 1961). The temperature is high as typically for tropical regions, the range is about $27^{\circ}C$ - $30^{\circ}C$ and the salinity is low due to the heavy precipitation and large river runoff. *Gordon et al.* (2003) proposed that a major factor in restricting a surface layer contribution to the Indonesian Troughflow is the meridional buoyancy gradient set up within the Makassar Strait that accompanies the redistribution of low salinity Java Sea water.

2.4.3 The Sunda Strait

The Sunda Strait (around $6^{\circ}S$ and $106^{\circ}E$) exhibits a large slope with approximately 18km width and 35m depth near the Java Sea and 50km width and 500m depth in the Indian Ocean. *Lelgemann et al.* (2000) have analysed in more detail the evolution of bathymetry in the Sunda Strait which is linked to the obliquely convergent geodynamic setting of the Sumatran trench. This is reflected by the existence of two sub-basins separated by a basement ridge, a shallower domain in the eastern and a deeper one in the western part of the Sunda Strait.

The water mass properties of the Java Sea influence the temperature and salinity in this area more dominantly than those from the Indian Ocean (*Wyrtki*, 1961). You (1999) identified four major water masses in the Indian Ocean. They are Indian Central Water (ICW), North Indian Central Water (NICW), Australasian Mediterranean Water (AAWM) and Red Sea Water (RSW) / Persian Gulf Water (PGW). Two of them, NICW and AAMW, are dominant around the southwest coast of Sumatra (around the Sunda Strait).

So far the information of the Sunda Strait is sparse because of its small size, i.e., in almost all regional and global ocean models this strait is usually ignored. Some discussions about the Sunda Strait can be found in the Naga Report (*Wyrtki*, 1961) and the model simulation result by *Wannasingha et al.* (2003). Using the OCCAM, *Wannasingha et al.* found that the total annual transport through the Sunda Strait is 0.3Sv using mean winds and 0.2Sv using 6-hourly winds and always directed to the Indian Ocean. Nevertheless, no further explanations are given concerning the variability in the Sunda Strait.

Chapter 3 The Data

In this study we use several data for model input and validation. The data are collected from several institutions and downloaded via the Internet.

3.1 The Input Data

3.1.1 Bathymetry

The bathymetry data were taken from the Digital Relief of Surface of the Earth (ETOPO 5)¹, Boulder Colorado : National Geophysics Data Center. ETOPO5 was generated from a digital data base of land and model topography on a 5-minutes latitude/longitude grid.

This study requires an area with a grid size of 10 minutes, so the original data have been averaged over 2 grid cells in x- and y-coordinates to get a 10 minute resolution. For correction, especially in the coastal area, the result is compared to the available bathymetric charts produced by the Indonesian Hydro-Oceanographic Office (Dinas Hidro-Oseanografi - Dishidros, Indonesia), as follows:

- Sea Chart Nr. 99 : Java Sea at a scale of 1:1.000.000
- Sea Chart Nr. 38 : Natuna Sea at a scale of 1:1.000.000
- Sea Chart Nr. 31 : West coast of Sumatra, Padang until Sunda Strait at a scale of 1:1.000.000

The final result of the bathymetry data is shown in Figure 1.2.

3.1.2 Tides

The tidal boundary forcing in the Indian Ocean is taken from the Global Tide Model (*Zahel*, 2000). At the boundaries along the Karimata Strait (0^{o} 30'S) and the eastern part of the Java Sea (114^{o} 30'E), the shelf sea model simulation results by *Stawarz* (1994) are used as boundary forcing. The model is forced by 5 main tidal constituents: M₂, S₂, N₂, O₁, K₁.

3.1.3 Temperature and Salinity

The climatology temperature and salinity (TS) data are taken from the World Ocean Atlas (WOA) (*Conkright et al.*, 1998). The horizontal resolution is $1^{\circ} \times 1^{\circ}$ in longitude and latitude with 33 vertical layers at standard level depths (0, 10, 20, 30, 50, 75, 100, 125, 150, 200, 250, 300, 400, 500, 600, 700, 800, 900, 1000, 1100, 1200, 1300, 1400, 1500, 1750, 2000, 2500, 3000, 3500, 4000, 4500, 5000, 5500m).

The WOA is documented by the Ocean Climate Laboratory (OCL) - National Oceanographic Data Center (NODC). The NODC/OCL is supported by the NOAA Climate and Global Change program, the NOAA Environmental Science Data and Information Management program (ESDIM),

¹Online at *http*://www.ngdc.noaa.gov/mgg/global/relief/ETOPO5/TOPO/ETOPO5/



Figure 3.1: Temperature and salinity for initial value from WOA (1998) The initial values of temperature and salinity on 1 January were linearly interpolated from the data in December and January. The figures above show the temperature and salinity in the first and second layer.

and the joint NOAA/NASA Global Change element, to produce scientifically quality controlled oceanographic databases and objectively analyzed global fields of oceanographic variables, as well as to perform diagnostic studies based on these databases. The temperature data (^oCelcius) in the WOA are collected from the bottle data, high and low resolution CTD data, mechanical-, digital-, and expendable-bathythermograph, surface data, and also buoy data. The salinity data (psu) are collected from bottle data, high and low resolution CTD, and the surface data (*Boyer et al.*, 1998).

The initial TS data for 1 January were obtained by linear interpolation from the TS data between December and January. For every grid cells, the salinity and temperature data are interpolated horizontally to 10 minute grid size (about 18km) and vertically for each level used in the model (see model design in 4.4) respectively.

3.1.4 Atmospheric Forcing

The atmosphere forces the ocean by fluxes of momentum, heat and fresh water. Evaporation and precipitation influence the sea surface temperature and salinity, and together with the sea surface pressure they change the density of the sea water. These conditions together with wind and tidal forces influence the current patterns. At the beginning, the model will be run with a constant wind

Variable	Level	Unit	Total Grid
Sea Level Pressure	Surface	Pascal	147 x 73
Air Temperature	2 m	^o K	192 x 94
Specific Humidity	2 m	kg/kg	192 x 94
U-Wind	10 m	m/s	192 x 94
V-Wind	10 m	m/s	192 x 94
Precipitation Rate	Surface	kgm ² /s	192 x 94
Total Cloud Cover	Surface	%	192 x 94

Table 3.1: Atmospheric data input from NCEP 1958-2002

forcing, and then continued with Hellerman wind stress data. At the last step, the simulation will be forced by atmospheric data taken from the National Center for Environmental Prediction (NCEP).

3.1.4.1 The Hellerman Wind Stress Data

This data set (*Hellerman*, 1983) contains monthly climatological data, which are averaged from observations from 1870 until 1976. The data are available on a 2° latitude x 2° longitude grid in units of *dynes/cm*². Over 35 million surface observations covering the world oceans from 1870-1976 have been processed for the purpose of calculating monthly normal and standard errors of the eastward and northward components of the wind stress at the lower 10 m of the atmosphere. The fields are intended to serve as boundary conditions for models of the ocean circulation.

The Hellerman wind stress data are interpolated horizontally to get a 10 minute latitude x 10 minute longitude grid size. In time a linear interpolation was used to get daily values. The wind forces the current constantly over 24 hours. The influence of the heat flux is not included in this case.

3.1.4.2 Long Term Atmospheric Data

NCEP has performed operational computer weather forecasts since the 1950s. From 1955 to 1973, the forecasts cover only the northern hemisphere; they have been global since 1973. Over the years, the quality of the models and methods especially including atmospheric observations has been improved continuously, resulting in major forecast improvements (*Kalnay*, 2003).

The NCEP/NCAR Reanalysis project is using a state-of-the-art analysis/forecast system to perform data assimilation using past data from 1948 to the present ². A large subset of this data is available from Climate Diagnostic Center (CDC) in its original 6-hourly data format (0Z, 6Z, 12Z, and 18Z) (*Kalnay et al.*, 1996).

The grid size of NCEP data is interpolated onto a 10x10 minutes grid horizontally using the available interpolation tool in the GrADS (The Grid Analysis and Display System) software. The atmospheric data used as input in this study are the sea level pressure (slp), air temperature, specific humidity, u- and v-wind components, and precipitation rates (see table 3.1). The slp data have a spatial coverage of 2.5° latitude x 2.5° longitude and temporal coverage of 6 hours whereas the other variables are lying on the T62 Gaussian spatial grid and have a temporal coverage of 6 hours.

The atmospheric fluxes are calculated from the above data and also from the sea surface temperature, which is calculated in the hydrodynamics model (see Appendix C).

²The NCEP data are online in *htt p* : //www.cdc.noaa.gov/

	River	Position	Mean	Minimum	Maximum
1.	Musi	03° 20' 10"S 104° 46' 30"E	616.60	391.00 (Aug)	1023.00 (Apr)
2.	Seputih	05° 04' 00"S 105° 50' 00"E	13.71	4.68 (Oct)	23.90 (Mar)
3.	Sekampung	05° 22' 00"S 104° 00' 00"E	12.70	7.40 (Aug)	16.90 (Feb)
4.	Kapuas	00° 50' 00"S 110° 26' 00"E	47.20	33.10 (Mar)	86.70 (Dec)
5.	Kahayan	02° 11' 00"S 113° 55' 00"E	1135.00	312.00 (Mar)	1954.00 (Aug)
6.	Citanduy	7° 23' 00"S 108° 33' 00"E	130.51	47.30 (Aug)	206.00 (Apr)

Table 3.2: River discharge data 1998 in m^3/s

3.1.5 River Runoff

Some rivers in Sumatra, Java, and Kalimantan discharge into the Java Sea and some rivers in Sumatra and Java into the Indian Ocean (see Figure 1.2). The Indonesian Research Institute for Water Resources, the Agency for Research and Development, Ministry of Settlement and Regional Infrastructures has measured the discharge rates of rivers in Indonesia periodically. These data are used for planning and implementation of the development and the use of water resources in the sector of public works such as agriculture, industry, power generation and tourism.

In this study, the daily discharge data of the year 1998 were selected for six major rivers: Kapuas and Kahayan River in Kalimantan, Citanduy River in the northern part of Java, as well as Musi, Sekampung and Seputih River in Sumatra. The Kapuas, Kahayan, Citanduy, Musi and Sekampung Rivers flow into the Java Sea, while the Seputih River flows into the Indian Ocean. The positions and discharges of the rivers are shown in Table 3.2.

3.2 Data for Validation

In order to get more realistic model results, a validation must be performed. For this purpose, some observational data that are available for the study area have been collected from Indonesian institutions and also via the internet.

The sea surface elevation (SSE) data are obtained from the Agency for the Survey Coordination and National Mapping (Badan Koordinasi Survei dan Pemetaan Nasional - BAKOSURTANAL) Indonesia. The data are available hourly during 1999 in Cilacap, Panjang-Lampung, and Tanjung Priok-Jakarta (see Table 3.4 and the positions are indicated as blue points 1,2,3 in Figure 3.2).

Time series of sea surface temperature and salinity are obtained from the Seawatch Indonesia Project, Agency for the Assessment and Application of Technology (Badan Pengkajian dan Penerapan Teknologi BPPT), Indonesia. The hourly data are available at 3 locations: Jepara (1997-1999), Seribu Islands (1999), and Karawang (1997-1999) (indicated as green points a,b,c in Figure 3.2). The monthly averages of those data are used to validate the model results. The other SST data (T16 and T17) are retrieved from Global Satellite Data of AVHRR - NOAA (1985-now) that are available online on the Internet. The data are monthly and have a resolution of 18km x 18km in latitude and longitude (indicated as green points d and e in Figure 3.2). Nevertheless, in tropical areas such as Indonesia the SST data are often not measured perfectly especially at the time of the rainy season because of the cloud cover. That means the monthly SST data are not averaged over all days of a month but only partially averaged over the days where data are available.

The vertical profiles of temperature and salinity in the southern part of Sunda Strait and Java Sea are obtained from the Pre JIGSE cruises from 25 October to 11 November 2000. The positions of the stations are shown as black points in Figure 3.2, whereas the observation time is given in Table 3.3.



Figure 3.2: The locations of observed data for model validation and cross sections for analysis Sea Surface Elevation Data during 1999 (blue points) :

- 1. Cilacap at 108°59'E and 7°34'S
- 2. Tanjung Priok-Jakarta at 106°53'E and 6°6'S
- 3. Panjang-Lampung at 105°19'E and 2°27'S

SST Data during 1997 - 1999 (green points) :

- a. Jepara at 110°36'E and 6°36'S
- b. Karawang at $107^{o}4.26$ 'E and $5^{o}46.86$ 'S
- c. Seribu Islands at 106°37.5'E and 5°39'S
- d. T16 at 105°15'E and 6°32.5'S
- e. T17 at 104°32.5'E and 7°45'S

The PreJIGSE Data 2000 (black points); the station names and positions can be seen in Table 3.3

Station	Position	Date	Time
Sta 1.	106°15.00'E 5°50.13'S	25.10.2000	19.37
Sta 2.	105°47.37'E 6°10.97'S	26.10.2000	04.51
Sta 2a.	105°37.41'E 6°13.36'S	28.10.2000	18.50
Sta 3.	105°35.37'E 6°24.97'S	28.10.2000	00.33
Sta 4.	105°20.01'E 6°11.79'S	28.10.2000	07.05
Sta 5.	105°00.28'E 6°00.78'S	29.10.2000	17.27
Sta 6.	104°11.64'E 5°37.74'S	29.10.2000	07.27
Sta 6a.	104°13.79'E 5°36.95'S	30.10.2000	13.08
Sta 7.	104°09.07'E 5°46.43'S	29.10.2000	14.42
Sta 8.	103°54.89'E 6°03.47'S	29.10.2000	18.58
Sta 9.	103°41.33'E 6°30.64'S	30.10.2000	00.37
Sta 10.	103°20.55'E 7°00.74'S	30.10.2000	05.31
Sta 13.	103°49.74'E 7°30.60'S	30.10.2000	11.03
Sta 14.	104°07.29'E 7°05.38'S	30.10.2000	17.24
Sta 15.	104°21.30'E 6°44.17'S	30.10.2000	21.48
Sta 16.	104°38.44'E 6°19.01'S	28.10.2000	22.53
Sta 17.	104°52.61'E 6°31.89'S	31.10.2000	02.35
Sta 18.	108°52.08'E 8°24.00'S	03.11.2000	13.53
Sta 19.	108°59.77'E 7°49.20'S	08.11.2000	19.35
Sta 20.	108°00.00'E 8°00.00'S	09.11.2000	03.37
Sta 21.	108°00.00'E 8°19.80'S	09.11.2000	08.23
Sta 22.	109°01.07'E 8°30.12'S	09.11.2000	22.51
Sta 23.	109°00.00'E 8°00.00'S	10.11.2000	03.27

 Table 3.3:
 Location of observed data during the Pre JIGSE 2000

 Table 3.4: The available hourly sea surface elevation data

Month	Cilacap	Panjang	Tanjung Priok
Jan	-	Х	-
Feb	Х	Х	Х
Mar	-	х	Х
Apr	-	Х	Х
May	х	х	Х
Jun	Х	Х	Х
Jul	Х	Х	Х
Aug	Х	Х	Х
Sep	Х	Х	Х
Oct	Х	Х	-
Nov	Х	Х	-
Dec	Х	Х	-

note: x = available, - = not available

Chapter 4 The Numerical Model

4.1 Equations of the HAMburg Shelf Ocean Model (HAMSOM)

This study uses the <u>HAM</u>burg <u>Shelf Ocean Model(HAMSOM)</u>, which has been developed by *Backhaus* (1983). The HAMSOM is a primitive equation model with a free surface and utilises two time-levels. The horizontal and vertical grid spacing is defined in Z co-ordinates on the Arakawa C-grid (*Arakawa* and *Lamb*, 1977). The model uses a semi-implicit scheme for separating the internal and external mode, which largely exempt from the stability criteria usually required for explicit formulations.

The model has been applied in many shelf seas of the world, where a number of model experiments and treatments modifications have been performed. Numerous papers and research reports described this model in detail such as *Backhaus* (1983,1985), *Backhaus* and *Hainbucher* (1987), *Pohlmann* (1987,1991), *Huang* (1995). In this chapter the description of HAMSOM will be limited only to the main equations, boundary conditions applied to this area, and the model design. A complete and detail explanation of the model and its discretization are given in the Appendix B.

The main equations of HAMSOM in the xy-coordinate systems are as follows: (the detail explanation of the symbols are listed in Appendix A)

Momentum equations in x-coordinate:

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} + A_H\nabla^2 u + \frac{\partial}{\partial z}\left(A_v\frac{\partial u}{\partial z}\right) + F_x$$
(4.1)

and in y-coordinate:

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + fu = -\frac{1}{\rho}\frac{\partial p}{\partial y} + A_H\nabla^2 v + \frac{\partial}{\partial z}\left(A_v\frac{\partial v}{\partial z}\right) + F_y$$
(4.2)

The vertical motion in z-coordinate is approximated by the vertical hydrostatic equation:

$$\frac{\partial p}{\partial z} = -\rho g \tag{4.3}$$

Continuity equation:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$
(4.4)

Temperature conservation equation:

$$\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} = K_H \nabla^2 T + \frac{\partial}{\partial z} \left(K_v \frac{\partial T}{\partial z} \right) + S_T$$
(4.5)

$$\frac{\partial S}{\partial t} + u\frac{\partial S}{\partial x} + v\frac{\partial S}{\partial y} + w\frac{\partial S}{\partial z} = K_H \nabla^2 S + \frac{\partial}{\partial z} \left(K_v \frac{\partial S}{\partial z} \right) + S_S$$
(4.6)

State equation of seawater:

$$\rho = \rho(S, T, p) = \rho_o + \rho' \tag{4.7}$$

4.2 Boundary Conditions

At the surface $(z = \zeta)$ and the bottom (z = -H), the following boundaries are used: Kinematic boundary conditions:

$$w_{\zeta} = \frac{\partial \zeta}{\partial t} + u_{\zeta} \frac{\partial \zeta}{\partial x} + v_{\zeta} \frac{\partial \zeta}{\partial y}$$

$$w_{-H} = u_{-H} \frac{\partial H}{\partial x} + v_{-H} \frac{\partial H}{\partial y}$$
(4.8)

Dynamic boundary conditions:

at the surface :

$$A_{Iv}\frac{\partial u_{\zeta}}{\partial z} - A_{Ih} \cdot \left(\frac{\partial u_{\zeta}}{\partial x}\frac{\partial \zeta}{\partial x} + \frac{\partial u_{\zeta}}{\partial y}\frac{\partial \zeta}{\partial y}\right) = \tau_{s}^{(x)}$$

$$A_{Iv}\frac{\partial v_{\zeta}}{\partial z} - A_{Ih} \cdot \left(\frac{\partial v_{\zeta}}{\partial x}\frac{\partial \zeta}{\partial x} + \frac{\partial v_{\zeta}}{\partial y}\frac{\partial \zeta}{\partial y}\right) = \tau_{s}^{(y)}$$
at the bottom :

$$A_{Iv}\frac{\partial u_{-H}}{\partial z} - A_{Ih} \cdot \left(\frac{\partial u_{-H}}{\partial x}\frac{\partial H}{\partial x} + \frac{\partial u_{-H}}{\partial y}\frac{\partial H}{\partial y}\right) = \tau_{b}^{(x)}$$

$$A_{Iv}\frac{\partial v_{-H}}{\partial z} - A_{Ih} \cdot \left(\frac{\partial v_{-H}}{\partial x}\frac{\partial H}{\partial x} + \frac{\partial v_{-H}}{\partial y}\frac{\partial H}{\partial y}\right) = \tau_{b}^{(y)}$$
(4.9)

At the closed boundaries the normal velocity component and the elevation are zero. That means, there are no water mass transported through the closed boundaries.

$$un_{ob} = 0 \tag{4.10}$$

For the lateral velocity a semi-slip condition is applied, i.e., the gradients of the tangential velocity components at the closed boundaries are set to zero.

At the open boundaries, the sea surface elevation (ζ) are obtained from the data which are calculated from the tidal components (see chapter 3). The gradient of the normal velocity component is set to zero.

$$\frac{\partial un_{ob}}{\partial norm} = 0 \tag{4.11}$$

The sea surface temperature and salinity are changed due to the heat fluxes. The detailed formulations can be seen in Appendix C. At the bottom, it is assumed that no heat and salt fluxes across the seabed floor occurs.

$$\left(\frac{\partial T}{\partial z}\right)_{-H} = \left(\frac{\partial S}{\partial z}\right)_{-H} = 0 \tag{4.12}$$

Temperature and Salinity at the open lateral boundaries are adapted from observed values by Newtonian damping. The variable can also be damped to a reference value.

$$\frac{\partial T}{\partial t} = \frac{(T_{clim} - T)}{Td}$$
(4.13)

$$\frac{\partial S}{\partial t} = \frac{(S_{clim} - S)}{Td} \tag{4.14}$$

where T_{clim} and S_{clim} are the climatological temperature and salinity respectively, which are taken from WOA 1998 as references, Td is the damping timescale, i.e., the smaller the value of Td, the stronger the property is kept to the reference value (*Kantha* and *Clayson*, 2000).

4.3 The Treatment of Temperature and Salinity at the Boundaries

At the open lateral boundaries, the temperature and salinity were advected and according to the direction of the transport. If there is an inflow to the internal model area, the temperature and salinity references are dominant. In the contrary, if the transport causes an outflow, the temperature and salinity from interior influence the boundary data. This is called 'the active boundary condition'.

By using the idea from the Newtonian damping of equations 4.13 and 4.14, the boundary conditions of temperature and salinity have been changed. At western open boundaries, the temperature and salinity were calculated as follows:

If u < 0 then

$$T_{ob} = \gamma_1 T_{ob-2} + \gamma_2 T_{ob-1} + \gamma_3 T_{clim}$$
(4.15)

$$S_{ob} = \gamma_1 S_{ob-2} + \gamma_2 S_{ob-1} + \gamma_3 S_{clim}$$
(4.16)

and if u > 0 then

$$T_{ob} = \gamma_1 T_{clim} + \gamma_2 T_{ob} + \gamma_3 T_{ob-1}$$
(4.17)

$$S_{ob} = \gamma_1 S_{clim} + \gamma_2 S_{ob} + \gamma_3 S_{ob-1} \tag{4.18}$$

Layers	Depth (m)	Layer Thickness (m)
1	0 - 10	10
2	10 - 30	20
3	30 - 50	20
4	50 - 80	30
5	80 - 110	30
6	110 - 140	30
7	140 - 170	30
8	170 - 200	30
9	200 - 250	50
10	250 - 300	50
11	300 - 400	100
12	400 - 500	100
13	500 - 1000	500
14	1000 - 1500	500
15	1500 - 2500	1000
16	2500 - 3500	1000
17	3500 - 4500	1000
18	4500 - 5500	1000
19	> 5500	

Table 4.1: The vertical grid of model

where the T_{ob} and S_{ob} are the temperature and salinity at the open boundaries, T_{ob-1} , S_{ob-1} , T_{ob-2} , S_{ob-2} , T_{ob-3} , S_{ob-3} are respectively temperatures and salinities at the 1, 2, 3 grids inside of model area. The γ_1 , γ_2 , γ_3 are relaxation coefficients at the boundary and are defined as:

$$\gamma_1 = \delta t / T_{const}$$

$$\gamma_2 = \delta t / C_{const}$$

$$\gamma_3 = 1 - \gamma_1 - \gamma_2$$
(4.19)

where $T_{const} = 1*24*3600 \text{ sec } (1 \text{ day}) \text{ and } C_{const} = 3*24*3600 \text{ sec } (3 \text{ days})$

The analogues treatments with opposite conditions for inflow and outflow were applied at the eastern open boundaries. At the northern and southern open boundaries for temperature and salinity, the analogous boundary conditions were formulated for v > 0 and if v < 0.

On the other hand, the heat fluxes in the deep areas as the Indian Ocean were calculated not only at the surface (first layer), but also at lower depth. The depth of the penetration depends on the the decay length scale for the absorption of solar radiation, which in turn depends on the classification of the water types. Details about the formulation can be found in the Appendix C.

4.4 Model Design

The grids size in this study is 10 minutes in both horizontal directions. The model has 19 vertical layers (Table 4.1). The total number of cells are 72 in longitudinal (from $102.5^{\circ}E$ westward), 60 in latitudinal direction (from $0.5^{\circ}S$ southward) and 19 in the vertical.

The user can select the vector-upstream scheme or the Arakawa J7 scheme (*Messinger* and *Arakawa*, 1976) for the advection of momentum in the model. Additionally, the vertical eddy diffusity can be parameterized in three ways: Richardson-Hainbucher, Kochergin-Pohlmann, or

constant vertical eddy viscosity. In this study we used the Arakawa J7 scheme for the advection of momentum and Kochergin-Pohlmann for the vertical eddy viscosity (see Appendix B). Those schemes were chosen in order to improve the stability of the model for long term simulations.

The simulation time step is 300 seconds, whereas the total time of the simulations depends on the scenario we used.
Chapter 5 Model Simulations and Validation

The main goal of the model simulations is to analyse and explain the long term variability of ocean dynamics in the study area. This goal can be achieved by applying the HAMSOM model that has been validated by observational data. For those purposes, a number of simulations are performed in this study.

5.1 The First Simulation (S1) : Tides

The S1 simulation is done in the barotropic mode with five main tidal constituents as the generating forces (see Chapter 3) and run for one year in 1999 based on the available observational data. S1 is used as a preliminary test of HAMSOM in the study area.

The tidal results are compared with the previous studies by *Wyrtki* (1961), *Mihardja* (1991) and *Stawarz* (1994). The hourly SSE simulation results are compared to available observational data at three locations (see Figure 3.2).

5.1.1 Tidal Distribution

As shown in *Stawarz* (1994) and the HAMSOM result, in the southwest of Kalimantan there is a clockwise amphidromic point of K_1 (Figure 5.1). The amplitudes are lower in the central Java Sea (10cm) than at its eastern and northern boundaries (40cm and 50cm, respectively). The same result was also found in *Mihardja* (1991).

The HAMSOM simulation got an amphidromic point of M_2 in the Java Sea (Figure 5.2) while *Stawarz* (1994) got two amphidromic points in the Karimata Strait at the northern boundary, but *Wyrtki* (1961) and *Mihardja* (1991) got no amphidromic point in the Java Sea. The amplitude from HAMSOM reaches 20cm in the middle of Java Sea while *Mihardja* (1991) got 35cm and *Stawarz* (1994) got less than 10cm.

The amplitudes of M_2 from HAMSOM reach more than 50cm in the eastern part of the Indian Ocean, while *Stawarz* (1994) got 70cm. In general the HAMSOM result shows that in the Indian Ocean the M_2 -tide is dominant as already mentioned by *Wyrtki* (1961).

The amplitudes of O_1 are 25% lower than of the K_1 , as also shown in previous studies. The amplitudes of O_1 are higher in the Karimata Strait (northern boundary) and the eastern part of the Java Sea than in the central part of the Java Sea. *Mihardja* (1991) got two counterclockwise amplidromic points of S_2 in the Java Sea with amplitudes of 5 to 10cm, while in HAMSOM results only one clockwise amphidromic point was found in the Java Sea with the same range of amplitude. The other results of this scenario are not displayed because they are not as dominant as the K_1 and M_2 -tide.

5.1.2 The SSE Time Series

The validation of SSE time series was done in 1999 at Cilacap at the southern coast of Java, Panjang-Lampung in the Sunda Strait, and Tanjung Priok in Jakarta. In this section, at Cilacap



Figure 5.1: The K_1 -Distribution of HAMSOM (left) and *Stawarz* (1994) (right). Stawarz used a 2-dimension horizontal model with grids size of 10 minutes in latitude and longitude the same as in the HAMSOM but with an area larger than our study area. The right figure above is cropped to get the same area as in this study



Figure 5.2: The M_2 -Distribution of HAMSOM (left) and *Stawarz* (1994) (right). *Stawarz* used a 2 dimension horizontal model with grids size of 10 minutes in latitude and longitude the same as in the HAMSOM but with an area larger than our study area. The right figure above is cropped to get the same area as in this study



Figure 5.3: Verifications of SSE at Tanjung Priok-Jakarta. For the locations see Figure 3.2



Figure 5.4: Verifications of SSE at Cilacap. For the locations see Figure 3.2



Figure 5.5: Verifications of SSE at Panjang-Lampung. For the locations see Figure 3.2

only the SSE in February and July whereas at Tanjung Priok and Panjang in March and July will be shown.

In general, the HAMSOM results and observed data are in good agreement. At Tanjung Priok-Jakarta there are quite big differences of the SSE during the neap tide in March and July. The reason for those differences is because in the model we used only 5 major tidal constituents, whereas in the shallow waters, as Tanjung Priok-Jakarta, the shallow waters tidal constituents are also dominant.

At Cilacap and Panjang-Lampung (Figure 5.4 and 5.5), the S1 results are in good agreement with the observed data, despite the fact that there are still differences in 12 July at Cilacap because of bad observed data and also numeric instable of the model in 15 and 16 July 1999.

5.2 The Second Simulation (S2) : Climatological Wind Stress

The S2 simulation is performed in the baroclinic mode. The model uses TS data from WOA (1998) as initial values and is forced by five main tidal constituents and climatological wind stress from *Hellerman* (1983), but without atmospheric heat fluxes. The purpose of the S2 simulation is to check the model stability when the wind forcing is also taken into account.

The model is run only for 2 years (1998-1999) because after that the SST lower than values are found in the Indian Ocean, especially during August 1998 where the SST is below 21° C, (are reached indicated by white color in Figure 5.6). Based on previous researches (*Wyrtki*, 1961; *Saji*, 1999; *You*, 1999; *Susanto et al.*, 2003) and AVHRR satellite data, such low SST never occurs in the Indian Ocean. We assumed that is discrepancies have occurred due to no mixing heat fluxes in the ocean surface. This means that the model should be reformulated to obtain more realistic results. Therefore, we included the atmospheric heat fluxes in the next simulations.







The atmospheric force used in this simulation is only the Hellerman's wind stress. The upward movement of temperature is very big because the vertical advection and diffusion is bigger than the horizontal one and there is no sea-air exchange at the sea surface.

5.3 The Third Simulation (S3) : Forcing by Atmospheric Heat Fluxes - Short Term Simulation

The S3 simulation is also done in the baroclinic mode with several changes and additional forces. The 6-hourly atmospheric forcing from NCEP (see Chapter 3) is used. The sensible heat fluxes Q_{SH} , the latent heat fluxes Q_{LH} , the heat fluxes by long wave radiation Q_{LW} and the solar radiation Q_{SW} are calculated to obtain the total heat fluxes of the upper ocean layer, where Q_{SW} penetrates vertically with an attenuation function until 200m depth. The fluxes of fresh waters also provided, the formulations are described in Appendix C. Discharge data of the rivers are used from the year 1998 as input for this scenario with the assumption of no interannual variation. The MSL is set to zero and it is assumed that there are no differences of MSL along the whole boundaries.

The model run was performed for only 3 years (1997-1999). The daily results are averaged over one month and validated by observational data at three locations in the Java Sea from the Seawatch buoys and at two locations (T16 and T17) of satellite data from by the AVHRR-NOAA in the southwestern part of the Sunda Strait (green points in Figure 3.2).

The SST simulation results at Jepara are quite different compared to the observations. The possible cause for this difference is the local influences at the measurement position which can not be



Figure 5.7: Verification of SST from January 1997 to December 1999 : The results of S3. Locations are shown in Figure 3.2

covered by the simulation (see Figure 5.7 a). Most differences occur from October 1997 to August 1998 with a maximum difference of 2° C occurring in June 1998. During the northwest monsoon (October 1998 until April 1999) the SST is almost the same.

The SST simulation results at Karawang do well agree in trend with the observational data but with cooler value of 0.5° C in December 1997 and higher value of $0.3-0.4^{\circ}$ C during the SE monsoon. The same condition is also shown around the Seribu Islands. We suggest that those conditions are caused by the inflow of water masses with lower temperature from the Indian Ocean into the Java Sea (see Figure 5.7 b and c) which can not be covered by the model (see section 5.4.1 and 5.4.2).

In the southwestern part of the Sunda Strait (T16 and T17) the SST from the simulation have the same pattern as the satellite data (see Figure 5.7 d and e). At T16, the maximum difference of 1.0° C occurred in March 1998 while at T17, a SST difference of about 1.5° C occurred during the monsoonal transition (January-March in each year). The possible causes of those differences are: Firstly, the vertical mixing in the model is not simulated optimally so that the lower temperature from the lower layers will be diffused too fast to the sea surface. Secondly, the total cloud cover in



Figure 5.8: The climatological SSC and SSE in August (left) and January (right) : results from run IV.A In August, representative for the SE monsoon, the SSE in the Java Sea is lower than that in January or during the NW monsoon, but the magnitude of SSC in August is larger than in January. Coastal areas of the eastern Indian Ocean, the SSE in August is lower than in January. In the south of 8°S the SSE in January is lower than in August. The mean magnitude of SSC in August is larger than in January.

this area is extremely high, thus only one or two days of observation are considered as the monthly data (see Section 3.2).

5.4 The Fourth Simulation (S4) : Forcing by Atmospheric Heat Fluxes - Long Term Simulation

The S4 simulation is same as S3, but with a total simulation time of 45 years from 1958 until 2002. The results will be analysed only for the latter 44 years (1959-2002), whereas the first year of simulation (1958) is ignored to avoid the spin up.

The purpose of this simulation is to analyse the long term variability of water properties and dynamics in the study area. At first, the long term simulation was carried out with zero MSL along all open boundaries (run IV.A). However, the simulation results could not fully represent the dynamic condition in the study area in comparison to the observed data, especially in the Sunda Strait. Based on the theory and regional analysis from previous researches, the variations of the sea level have in a significant influence on the ocean dynamics in the study area. Because of that reason, we perform run a long term simulation with different constant MSL at each open boundary (run IV.B).

5.4.1 Run IV.A

In run IV.A, the MSL along all open boundaries is set to zero. This means that there are no differences of water column height as well as of the barotropic pressure along the open boundaries.

$$\zeta_{msl} = 0 \tag{5.1}$$



Figure 5.9: The climatological SST and SSS in January and August : results from run IV.A In January, during the peak of NW monsoon, the higher salinity from the Indian Ocean is transported into the Java Sea through the Sunda Strait. On the contrary, during August, representative for the SE monsoon, the lower salinity from the Jave Sea is transported to the Indian Ocean.

The SSE in the Java Sea is lower than in the Indian Ocean during the NW monsoon and on the contrary is higher during the SE monsoon. The SSC in the Java Sea is forced dominantly by the surface wind. During the SE monsoon, for example in August, the current flows westward and turns northward through the Karimata Strait (Figure 5.8). The water mass is transported out from the Java Sea to the Indian Ocean. Reversely during the NW monsoon in December, the water mass from Karimata Strait meets with the water mass from the Indian Oceans that was transported into the Java Sea through the Sunda Strait and moves eastward.

The water mass from the Indian Ocean to the Java Sea reverses in the Sunda Strait every monsoon (see Figure 5.9). Figure 5.9 showed that during the NW monsoon, the salinity in the Java Sea is increased due to the inflow of higher salinity from the Indian Ocean that is transported to the Java Sea through the Sunda Strait.

5.4.2 Run IV.B

In run IV.B, the MSL values along each open boundary are different. The MSL along the northern open boundary is set to 0.42cm (Prigi station), whereas the MSL along the eastern boundary of the Java Sea entrance is set to 0.34cm (Musang station) and in the Indian Ocean at the western and eastern boundaries the MSL are set to 0.13cm (Pelabuhan Ratu station) (Table 5.1).



Figure 5.10: The climatological SSC and SSE in August (left) and January (right) : results from run IV.B The difference between both figures above and figures 5.8 is that in the Sunda Strait, the SSE of the Java Sea is always higher than of the eastern Indian Ocean. Thus, the water is transported out from the Java Sea.

Station	Mean Height of	Nummber of Years
	Sea Level (m)	with Observations
Prigi, east Sumatra	0.42	3
Musang Kecil, south Kalimantan	0.34	2
Java north coast	0.34	6-7
Pelabuhan Ratu, South coast Java	0.13	4-7
Southwest coast Sumatra	0.13	2-7

Table 5.1: Adopted mean sea level heights (m) used in the run IV.B (Wyrtki, 1961)

These MSL data are taken from adopted MSL heights that are available from the Naga Report (*Wyrtki*, 1961). *Wyrtki* (1961) had calculated the mean sea level heights (MSLH) from the height of sea level relative to the mean which was already corrected according to the atmospheric pressure.

Figures 5.10 and 5.11 show that the SSE in the Java Sea is always higher than in the Indian Ocean below latitude 8°S, near the Sunda Strait. The water is always transported out of the Java Sea to the Indian Ocean. During the SE monsoon, the SSE difference of along the Sunda Strait is higher than during the NW monsoon, therefore the SSC through the Sunda Strait is faster during the SE monsoon. In the Java Sea, the wind induced current is balanced by the difference of sea surface height (SSH), as impact of the different MSL prescribed at the northern and eastern boundaries.

5.5 Analysis of the Open Boundaries

Near the surface (upper 30m) fluxes of momentum from wind stress and heat are strong contributors to ocean variations. The deformation of sea surface is affected by wind and the distribution of mass and therefore closely related to circulation. Variations occur on relatively short time scales on the order of hours to days in the surface mixed layer (*Jacobs et al.*, 2001). Aside from the mixed layer, the open ocean may be generally viewed as warm water mass overlying colder relatively





denser water mass. There is a strong correlation between thermocline depth and sea level, which produces a pressure anomaly throughout the water column. For example for barotropic conditions, a relatively high sea level indicates high pressure, and this pushes the thermocline downward.

In a simplified manner, the difference of density between two oceans causes the difference of the water column height. If there are two oceans, the first one (A) has density ρ_1 and the other one (B) has ρ_2 and the ρ_1 is larger than ρ_2 . Assuming the pressure (p) gradient in the bottom of both oceans is zero, and then the MSL in ocean B is higher than in ocean A. That means the MSL in every ocean is different and the difference changes not only in time but also in space and due to barotropic pressure gradients influences a small area between the two oceans. Therefore in the nature it is difficult to observe the fix MSL of the ocean.

The above description will be used to explain the difference of sea level between the Indian and Pacific Ocean. Figures 5.16 shows that the difference of sea level height (SLH) between the Indian and Pacific Ocean also affects the SLH in the Java Sea and the eastern Indian Ocean. This means that the difference of sea level in the large area (between the Indian and Pacific Ocean) induces also a difference of sea level in regional areas, such as our study area. At first, this phenomenon takes a place along the boundaries and then influences to the ocean dynamics inside the local area.

Although in the regional area the difference of SLH is normally small, it can contribute significantly to the ocean dynamics. In normal years, the difference SLH between the Java Sea and



Figure 5.12: Verification of SST in January 1997 - December 1999 : results from run IV.A (left) and run IV.B (right)



Figure 5.13: The SST distributions from the NOAA Satellite (top), run IV.A (centre) and run IV.B (bottom) in (a) January and (b) August 1997

eastern Indian Ocean is less than 5cm, but during extreme conditions such as El Niño, La Niña, and Dipole Mode years, the difference can reach more than 15cm (Figure 5.16). From the different results of run IV.A and IV.B, it can be concluded that the influence of SLH difference is significant in the study area. Unfortunately, these changes are not calculated in HAMSOM but have to be prescribed at the open boundaries.

The simulation results where different MSL are applied at each open boundary especially affect the shallow water area such as the Java Sea. In the deep sea area such as the Indian Ocean, the SSE and SSC pattern are not influenced significantly, but the horizontal distribution of water temperature and/or salinity are affected strongly.

The horizontal distribution of SST from the satellite data (top row Figure 5.13 b) shows that in August 1997 in the Indian Ocean and Java Sea, lower SST was transported to the west. In contrast in January 1997 (top row Figure 5.13 a), warm water was transported from the north to the Java Sea while in the Indian Ocean the transport was still to the west. It means that the influence of the monsoon is more dominant in shallow waters such as the Java Sea than in the deep Indian Ocean. The SST in the Java Sea reached more than 30° C in November 1996 and became cooler during the NW monsoon 1996/1997. In the eastern Indian Ocean the SST was cooler than $25^{\circ}C$ during the SE monsoon and warmer (28- $30^{\circ}C$) during the NW monsoon, especially in the western part of the study area. During the NW monsoon 1996/1997 (rainy season), the satellite data did not cover the whole time and area because of the strong cloud cover over Indonesia was high.

Center and bottom rows of Figure 5.13 show the SST results from run IV.A and IV.B, respectively.



Figure 5.14: The TS diagram with zero MSL along the open boundaries from run IV.A. The positions of the stations are shown in Figure 3.2

There are almost no differences in the SST results between run IV.A and IV.B (see also Figure 5.12), but for the SSS we can see that differences between run IV.A and IV.B are very substantial. Figure 5.9 and Figure 5.11 show that the MSL difference along each boundary will affect the change of SLH and also the density. The influence appeared more clearly in the horizontal distribution of salinity. Unfortunately, we did not have measured data of the SSS distribution in this area. Therefore to know the effect of different open boundary conditions (run IV.A and IV.B) on the simulation results, we analyze the TS-diagrams.

Figure 5.14 is from run IV.A and shows that the influence of the Java Sea water mass to the Indian Ocean is insignificant or, in other words, the influence of water mass from the Indian Ocean to the Java Sea is more dominant, especially during the NW monsoon (see also Figure 5.9). But the data (red color) show that the influence of higher temperature and lower salinity in the surface is dominant. Thus the water mass from the Java Sea is transported into the Indian Ocean as also described in *Wyrtki* (1961) and *Wannasingha et al.* (2003) who have used the OCCAM global ocean model and found that the water mass is always transported from the Java Sea to the Indian Ocean through the Sunda Strait.

The TS-profiles in Figure 5.15 are from run IV.B and show that the results are quite similar to



Figure 5.15: The TS diagram with different constant MSL at each open boundary from run IV.B. The positions of the stations are shown in Figure 3.2

the data. The low salinity and high temperature from the Java Sea are mixed well with the water mass from the Indian Ocean. Unfortunately, water masses with a salinity higher than 35psu and a temperature of 12-25°C can not be simulated in the model. We suppose that the higher salinity in reality is transported from southwest of Sumatra. In the model simulation, the salinity for the initial and boundary values are taken from WOA (1998) climatologically data and the values are lower than 35psu. Thus, it is still necessary to adjust the boundary conditions. The upper right of Figure 5.15 illustrates the TS profile in the Java Sea. We are not able to explain the difference between the simulation results and observed data, because the data is only from one station (Sta.1) at one certain moment, whereas the simulation results are daily averages during the entire Pre JIGSE observation time. The bottom left and right figures describe the TS profile in the south of West Java and south of Central Java, respectively (see Table 3.3), and show that the simulation results are in good agreement with the observed data.

The ocean current circulations in the Southeast Asian and Australian waters have been described and explained by Wyrtki (1961) and in general both run IV.A and IV.B are in good agreement with these previous results. In December the SSC in the coastal zone of Sumatra flows to southeast parallel to the coastline of West Sumatra and south of Java Island. The model results and *Wyrtki* (1961) showed that the surface current has velocity of 0.5m/s and smaller in the south of East Java.









Figure 5.16: Anomaly of Sea Surface Topography during (a) normal condition September 1993, (b) normal condition September 1996, (c) El Niño September 1997, and (d) La Niña September 1998 (Center for Space Research, University of Texas, available online at ftp.csr.utexas.edu/pub/sst/month/images). note: September 1993 in Table 2.1 is El Niño condition.

(c)

In the offshore of the eastern Indian Ocean, south of latitude 8° S, the SSC flows to the west during the whole year. In August the SSC in the south of Java was amplified to about 1m/s and flows to the west, and reduced the southeastward current from the coast of Sumatra. The SSC turns to the offshore and convergent in the area of 8° S (southwest of Sunda Strait). In this area, the current patterns of both runs (Figure 5.10 and Figure 5.8) agree with the description from *Wyrtki*, *1961* (see Figure 2.3).

In January, the surface current flows from the South China Sea into the Java Sea via the Karimata Strait and turns eastward in the Java Sea. The mean velocity of the SSC are about 0.5 to 0.75m/s with run IV.A (Figure 5.8), about 0.25 to 0.5m/s with run IV.B (Figure 5.10 right), and about 0.75 m/s in the Naga Report. On the other hand, in August the SSC flows westward 0.5m/s and turns northward in the Karimata Strait to the Southeast China Sea (Figure 2.3 top illustration by *Wyrtki*, 1961). The run IV.B shows that the westward current flows slower than 0.3m/s in the Java Sea (Figure 5.10 left) while run IV.A shows that the SSC flows more than 0.5m/s (Figure 5.8 left). The different SSE due to the dominant of SE-wind forces could explain why in run IV.A the water column in the east is higher than in the north. As a consequence, the SSC flows faster than run IV.B. In run IV.B, the MSL in the northern side is higher than in the east and north boundary.

In the Sunda Strait, *Wyrtki* (1961) only described that the current flows from the Java Sea to the Indian Ocean with velocity of 0.75m/s in August. He said that the total transport is not more than 0.5Sv from the Java Sea to the Indian Ocean, whereas *Wannasingha et al.* (2003) found that the transport is about 0.2-0.3Sv in the same direction. The same condition is also found in run IV.B while in run IV.A the water mass that is transported from and into the Java Sea depends on the monsoon.

From the analysis above, we can conclude here that the simulation results from run IV.B are better and more realistic. We thought that better and more realistic results can be obtained if the model used a combination of MSL as done in run IV.B and included the changes of SLH or the sea level anomaly from the altimetry satellite as a variable. The changes can be calculated every time so that it is possible to estimate the density gradient and consequently the baroclinic portion of the pressure gradient force (*Kennan* and *Niiler*, 2003).

Chapter 6 The Climatological Situation

The heat transport from the atmosphere to the ocean caused the ocean condition of the Indonesian waters to be varied. The heat fluxes change instrinsic every six months, depending on the sun soltice position.

The climatological conditions are analysed from the monthly averaged of model simulation for 44 years from 1959 to 2002.

6.1 Monsoonal Variation of SST and SSS

During the NW monsoon (rainy season), the precipitation in the study area is higher than the evaporation so that the SSS and SST become lower. On the other hand, during the SE monsoon (dry season), the evaporation is higher than the precipitation so that the SSS and SST increase. The solar radiation increases with the increasing clarity of the atmosphere and this condition also causes the SST to increase.

Based on the simulation result, the mean SST in the Java Sea is about $28-33^{\circ}C$ while in the eastern Indian Ocean it is about $24-28^{\circ}C$. In the beginning phase of the NW monsoon (November), the SST in the Java Sea starts to decrease rapid from $29^{\circ}C$, reaches the minimum in February at around $27 - 28^{\circ}C$, starts to increase again in April and reaches its maximum in May or June (higher than $30^{\circ}C$). In contrast, the SST in the eastern Indian Ocean starts to decrease in July, reaches the minimum in September (peak of SE monsoon), starts to increase during the NW monsoon and reaches the maximum in April. Nevertheless, in the southwest of Sumatra the SST is warmer by about $30^{\circ}C$ from April until June (Figure 6.1).

The mean SSS in the Java Sea is about 30-33.4psu. Along the east coast of Sumatra and south of Kalimantan, there are several rivers that discharge freshwater to the sea, which mixes well with the sea water in the shallow water area. The large freshwater discharge from the Musi River in Southeast Sumatera causes the SSS in some places around the east coast of Sumatra to be lower than 30psu (Figure 6.2). This low salinity is transported to the Indian Ocean through the Sunda Strait. During the SE monsoon, the water mass with 32.4psu is transported to the offshore of the Indian Ocean until more than 100km while during the NW monsoon the water mass with salinity lower than 32psu is transported only until 50km.

The mean SSS in the eastern Indian Ocean is higher than 34psu, except around the Sunda Strait. The water mass with lower salinity from the Java Sea is always transported to the Indian Ocean. During the NW monsoon the eastward transport can reach the south part of Central Java, while in the peak of SE monsoon (August) reaches only around the Sunda Strait (see Figure 6.2).

6.2 Monsoonal Circulation in the Java Sea

In the Java Sea, the SSC circulation is controlled by the monsoon. During the NW monsoon (see Figure 6.3) the SSC flows from the northern to the eastern part of the Karimata Strait. The SSE in the northern part of the Java Sea is higher than in the eastern part, which is due to the higher MSL (see Section 5.4.2). The maximum of total transport from the Karimata Strait to the Java Sea



Figure 6.1: The climatological variation of SST from the model output, averaged from 1959 to 2002



Figure 6.2: The climatological variation of SSS from the model output, averaged from 1959 to 2002





104°F

106°F

108°E

110°F

112°F

114°F



Figure 6.3: Variation of SSE and SSC during the NW monsoon (December-March) The maximum SSE in the Java Sea during the NW monsoon occurrs in March and the minimum in December. The SSC from the Java Sea to the Indian Ocean through the Sunda Strait reached it's minimum when the difference between both Seas is minimum.

is about 2.1Sv (occurred in January), the transport through the Sunda Strait to the Indian Ocean is about 0.5Sv while 1.6Sv flow to the eastern part of the Java Sea (Figure 6.9). The mean current velocity is about 0.5-0.75m/s from the north until the centre part of the Java Sea and decreases during flowing to the eastern part of the Java Sea.

Both temperature and salinity are vertical well mixed. In December, the salinity at the southern coast of Kalimantan (position G figure 3.2) is lower than at the northern coast of Java (position H in figure 3.2), about 32.4psu and 32.6psu, respectively, and the temperature is lower at H than at G, about 28.2 and 28.6°C respectively (Figure 6.4 left). This situation occurs because of several rivers, which is located around the western and southern coast of Kalimantan, such as the Kapuas and Kahayan Rivers, have large discharges during the rainy season (Figure 6.4 right). During the

4°S

6°S

8°S

10°S

-50 -40 -30 -20 -10



Figure 6.4: Vertical distribution of temperature (left) and salinity (right) during the NW monsoon (December-February) in the cross section GH (see Figure 3.2).

The water mass is well mixed in the Java Sea. There is almost no vertical variation of temperature and salinity with the depth. The maximum temperature occurs on December about $29^{\circ}C$ and the lowest salinity 32.5psu in January



Figure 6.5: Variation of SSE and SSC during the transition of NW to SE monsoon (April-May) The SSC flows reversely from the east to the west in the Java Sea and reaches the minimum in May. The westward flow of SSC in the eastern Indian Ocean was speeded up in May by the surface wind.





Figure 6.6: Variation of SSE and SSC during the SE monsoon (June-September) The maximum difference of SSE between the Java Sea and the Indian Ocean occurs in August/September, therefore the SSC through the Sunda Strait is maximum. The SSC in the eastern Indian Ocean also reaches its maximum due to the SE wind.

NW monsoon, the vertical distribution of temperature and salinity are almost homogen, $28^{\circ}C$ and 32.5psu, respectively. Nevertheless, in February the salinity at H is lower than at G. We suppose that there is a transport of water mass with higher salinity from the South China Sea along the coast of Kalimantan and water mass with lower salinity from the eastern coast of Sumatra to the eastern part of the Java Sea.

Figure 6.5 shows that during the monsoon transition, beginning in April, the SSC still flows eastward with lower velocity (compared to the SSC in the rainy season) and reaches the minimum in May. In the eastern part of the Java Sea, the SSC turned westward with slower velocity.

Reversely, during the SE monsoon the wind blows strongly to the west and turns northward in the north of the equator. Figure 6.6 shows that the SSC was flow westward and flow out to the north



Figure 6.7: Vertical distribution of temperature (left) and salinity (right) during the SE monsoon (July-September) in the cross section GH (see Figure 3.2)

The maximum temperature in July is more than $29^{\circ}C$ and the horizontal distribution of salinity in the north coast of Java is lower than in the south coast of Kalimantan (32.2 and 32.7psu, respectively). The salinity reaches 32.7psu in September in the middle of Java Sea, but the temperature become cooler to 27.6 and $28.7^{\circ}C$ at H and G respectively.

through the Karimata Strait with a mean velocity of 0.25m/s (Figure 6.6). The maximum of total transport occurs in August (about 1.0Sv), flows from the east into the Java Sea, and flows out to the north about 0.3Sv and the rest (0.7Sv) flows out through the Sunda Strait into the Indian Ocean. The total water mass transport during this monsoon is lower in comparison to the NW monsoon because the SSC in the Java Sea, which is dominantly forced by the southeasterly wind. Therefore, only in July and August (peak of the SE monsoon) the water mass can be transported out through the Karimata Strait. On the other months (June and September) during the SE monsoon, the water mass flows into the Java Sea from the eastern part and also from Karimata Strait, and most of it is transported out through the Sunda Strait (Figure 6.9).

During the SE monsoon the temperature increases and reaches its maximum in July. The temperature at G is warmer than at H (see Figure 6.7 left). In the Java Sea, the amount of evaporation is larger than the precipitation, so the salinity increases and reaches the maximum (more than 32.7psu) in September in the middle of the Java Sea. On the other hand, low salinity water from the east boundary reached the north coast of Java, and the salinity decreases down to 32.2psu (see Figure 6.7 right).

During the transition to the NW monsoon (October), the SSC still flows westward with a slower velocity and then reverse to the east in November and faster during the NW monsoon (Figure 6.8). The current in each layer in the Java Sea is almost homogen because of the shallow waters. The water is well mixed and the current velocity reduced only by 18% from the surface to 50m depth.



Figure 6.8: Variation of SSE and SSC during the transition from SE to NW monsoon (October-November) The SSC flows reversely from the west to the east in the Java Sea and reaches its minimum in October. In November the transport turns eastward while in the eastern Indian Ocean the SSC still flows westward but with decreased velocity.

6.3 Transport of Water Mass through the Sunda Strait

The change of the amount of monthly mean water mass transport during a year through the Sunda Strait is small, from about 0.48Sv in December to 0.72Sv in August and September. The water mass influence from the Java Sea is more dominant than from the Indian Ocean in the whole year (shown by negative transport in Figure 6.9). The transport during the NW monsoon is reduced by 33% compared to the transport during the SE monsoon and flows always to the Indian Ocean. *Wyrtki* (1961) estimated that the transport through the Sunda Strait does not exceed 0.5Sv and is always directed towards the Indian Ocean. Another simulation using the OCCAM Model done by *Wannasingha et al.* (2003) showed that the total transport through the Sunda Strait is only 0.3Sv using the mean wind and 0.2Sv using the 6-hourly wind and is always directed to the Indian Ocean.

From the surface to 150m depth, the water mass around the Sunda Strait has a low salinity and high temperature as the Java Sea water mass, but below 150m it is characterized by high salinity and low temperature as in the eastern Indian Ocean. In August the water mass with salinity lower than 34psu is transported from the Sunda Strait to 80 km offshore the Indian Ocean (Figure 6.10), but water mass with temperature higher than $29^{\circ}C$ from the Java Sea was kept due to the upwelling in the Indian Ocean. The upwelling caused the upward movement and mixing between two water masses. The outcome of this is a large value of temperature change and a low value of salinity change.

Reversely during the NW monsoon (Figure 6.11), the water mass with salinity lower than 34psu is transported to 30-50km offshore the Indian Ocean and the water mass from the Java Sea with temperature higher than $27^{\circ}C$ is dispersed offshore the Indian Ocean until 70km.

In the cross section EF at 500-1000m water depth, as shown in Figure 6.11 and 6.10, there is a water mass with high salinity during the whole year. In July the water mass starts to be dominant and moves upward until 250m depth in August. It starts to reduce in the beginning of the NW monsoon (December) and can be identified only in 500m water depth.

You (1999) identified a mixing water mass between NICW and AAMW around the southwest of



Figure 6.9: The seasonal variation of water mass transport in the Sunda Strait (s1-s2), in the eastern (e1-e2) and northern (n1-n2) parts of the Java Sea (see purple lines in the figure 3.2). The positive (negative) signs mean that the water mass was transported into (out of) the Java Sea.

Sumatra and showed that the NICW is more dominant during the SE monsoon than during the NW monsoon, as shown by our simulation results. The contribution of AAMW and the NICW around the Sunda Strait (latitude $9-10^{\circ}S$ and longitude $100-105^{\circ}E$) are about 60-70% and 10-20% respectively, during the NW monsoon and about 50% and 30-40% respectively, during the SE monsoon.

We expect that the NICW water mass from the north is transported along the west coast of Sumatra and moves up due to upwelling in August. To identify this water mass, we made a new cross section C_1D_1 at $103^{\circ}E$ (western part of cross section CD). From the vertical profile of salinity in this cross section, (Figure 6.12), we can see that the water mass with higher salinity is more obvious and transported upwards until 150m depth in August and until 200m depth in December.

The water mass with higher salinity was not identified in the southern coast of Central Java (cross section AB). This condition indicates that the water mass with higher salinity, which is transported from the western coast of Sumatra, is not transported further to the east, but turns to the southwest (normal to the coastline), to the offshore of Sunda Strait as indicated by the current pattern (Figure



Figure 6.10: Vertical profile of temperature and salinity in August along the cross section EF (see Figure 3.2)

6.6).

6.4 Analysis of Upwelling in the Eastern Indian Ocean

In the offshore of the Indian Ocean, at more than 111km from the coast, the current flows continuously westward every month. During the SE monsoon, the SSC increases with the increasing of SE wind and reduces by 50% during the NW monsoon (Figure 6.3 and 6.6). The magnitude variation of SSC is about 0.5m/s during the NW monsoon (maximum in January) and 0.75-1.0m/s during the SE monsoon (maximum in August).

Below 50m depth, the horizontal current pattern is almost stable for the whole year. In the east of 109° E the current flows westward between 500m until 1000m depth and eastward below 3000m depth and independent to the monsoon (Figure 6.15). These currents are influenced by the difference of the density and the sea level pressure between the west and the east area. Above 50m depth, the current flows westward and is reduced by the westerly wind forcing.

In the near shore, less than 111km from the coast, the SSC flows from the western coast of Sumatra to the southeast parallel to the Sunda Strait. During the NW monsoon, the current in the depth less than 50m flows southeastward parallel to the western coast of Sumatra and meets with the current from the Sunda Strait, with a velocity about 0.20-0.25m/s (Figure 6.13 left), and flows further to



Figure 6.11: Vertical profile of temperature and salinity in January along the cross section EF (see Figure 3.2)

the east and known as the South Java Current. These current patterns still can be detected until 20km from the southern coast of Central Java. Figure 6.15 left shows that at $109^{\circ}E$ the SSC flows westward with the velocity of 0.20m/s.

On the other hand, during the SE monsoon, the SSC flows westward with a velocity of 0.8-1.0m/s in the southern coast of Central Java (Figure 6.15 right : east-west current component). In the cross section CD, the SSC flows westward only in July, whereas in other months, it flows eastward with a velocity lower than 0.15m/s (Figure 6.13 right).

This near shore current causes the occurrence of upwelling, which is identified by a positive sign in the vertical current velocity component. The signal is more obvious in the eastern than in the western part. At $109^{\circ}E$, 111km from the coast, above 200m the vertical current is about (+)0.004cm/s during the NW monsoon (Figure 6.16 left) and reaches its maximum in the SE monsoon, about (+)0.008cm/s (Figure 6.16 right). At $105^{\circ}E$ (Figure 6.14 left), during the NW monsoon the vertical current is about (-)0.001cm/s and moves downward until 125m depth, while from 200m to 125m depth the current moves upward with velocity of (+)0.002cm/s and caused of mixing of water mass in 125m depth. The downwelling above 125m reaches its minimum in July. The downward moving occurs because the Java Sea water mass, with low salinity and high temperature, always flows to the Indian Ocean. On the other hand, during August-September, the vertical current above 125m moves upward but with velocity of (+)0.0025cm/s (Figure 6.14 right).



Figure 6.12: Vertical profile of salinity in December (left) and August (right) in the cross section $C_1 D_1$ (see Figure 3.2)

Water mass with high salinity is found in the western part of CD.



Figure 6.13: The east-west current component during the NW monsoon (left) and SE monsoon (right), 111km from the coast of southwest Sumatra at 105° E (see Figure 3.2). Positive sign means that the current flows eastward.



Figure 6.14: The vertical current component during the NW monsoon (left) and SE monsoon (right), 111km from the coast of southwest Sumatra at 105°E (see Figure 3.2). Positive sign means that the current flows upward.



Figure 6.15: The east-west current component in the cross section AB (see Figure 3.2) at 109°E and 111km from the coast during the NW monsoon (left) and SE monsoon (right). Positive sign means that the current flows eastward

Due to the ocean dynamics, the temperature and salinity near coast are also transported upward. The climatological temperature and salinity shows that the upwelling occurs from July to the end of September in the eastern part of south Java, whereas in the western part of south Java from the middle of July to September. At $105^{\circ}E$, water mass with a temperature of $20^{\circ}C$ is found in 90m depth during the NW monsoon and transported upwards to 75m depth from June to the end of October (Figure 6.17 left). At the surface, the temperature becomes cooler until $27^{\circ}C$ during the SE monsoon due to the upwelling while during the NW monsoon it warms up back until $30^{\circ}C$. The upward movements of water mass with a salinity more than 34.4psu reaches its maximum in October (Figure 6.18 left) and the isotherm of $20^{\circ}C$ moved upward from 80m (normally in NW monsoon) to 75m depth during the SE monsoon and reached its maximum on August, whereas the upward movement of water mass with salinity higher than 34.4psu reached its maximum on



Figure 6.16: The vertical current component in the cross section AB (see Figure 3.2) at 109°E and 111km from the coast during the NW monsoon (left) and SE monsoon (right). Positive sign means that the current flows upward



Figure 6.17: Vertical profile of climatological temperature (left) and salinity (right) at C' in the cross section CD, about 111km from the coast (see Figure 3.2)



Figure 6.18: Vertical profile of climatological temperature (left) and salinity (right) about 111km from the coast in the cross section AB (see Figure 3.2)

October (Figure 6.18 right). During the SE monsoon, the upwelling caused the SST to be cooler than at $25^{\circ}C$.

Below 1250m (as example Figure 6.10 and 6.11), the water mass is stable, which is characterized by no fluctuation of temperature and salinity. The water temperature lower than 5° C and almost homogen is found in the depth of 1200-1900m while the water temperature lower than 3° C is found in the depth below 2000m. The salinity is almost homogen, about 34.4psu, in the depth below 200m. The fluctuation of temperature and salinity occurs in the upper above 200m layer only due to the air-sea exchange.

Chapter 7 Ocean Variability During 1959-2002

During 1959-2002, the variability in the Java Sea and eastern Indian Ocean were influenced by the ENSO and DME. In this study, the ENSO and DME years are based on the Table 2.1 from NOAA and *Saji et al.* (1999), respectively.

7.1 Upwelling Variability South of Java

There are 3 classifications of upwelling *Wyrtki* (1961): stationary, periodic and alternating. The stationary upwelling occurs in the whole year while the periodic upwelling occurs only during one season, and the alternating upwelling occurs alternately. As already mentioned in Chapter 6, the upwelling in the south of Java is classified as periodic upwelling because normally it occurs during the SE monsoon. The surface wind blows northwest-ward parallel to the south coast of Java and caused the offshore transport in the surface while from the deeper layers, cooler and more saline water mass moves upwards. The upwelling makes the waters rich of nutrients. The stronger the upwelling, the more nutrient is transported upward and more fishes can be found in that area. Therefore, the upwelling depend on the atmospheric condition especially the strength/weakness of SE-wind and the influence of El Niño, La Niña and DME.

Normally, the upwelling area in the south of Java (Figures 6.1 and 6.2) is found until 111km offshore (1^o latitude) and transported westward by the SE-wind. The strength of upwelling and its transport along south of Java and south west of Sumatra is different and depend on the zonal component of SE wind.

During the normal condition in 1959, the upwelling occurred with an anomaly of SST and SSS around $-1.5^{\circ}C$ and +0.3psu (Figures 7.1 and 7.2), respectively, and covered the area from south coast of Java to the latitude of $9^{\circ}S$. The upwelling got stronger during the SE monsoon 1960. The decreasing SST in the eastern Indian Ocean was transported to the west. The anomaly of SST in the eastern part is about $-1.5^{\circ}C$ from end of May until July 1960 while in the south coast of Central Java was about $-1.8^{\circ}C$ from mid of June until mid of July. On the other hand, in the western part the SST decreased only by $-0.3^{\circ}C$ (see Figure 7.4). The anomaly of SSS in the eastern part is only +0.3psu while in the western part (Sunda Strait) the SSS decreased. These conditions showed that the water mass from the Java Sea strongly influenced the water mass in the Sunda Strait.

Although 1959, 1960, 1979, and 1980 are identified as normal years, the strength of upwelling are different. In 1959 and 1960, the SE-wind blows almost parallel to the coast of south Java and the anomaly of zonal component is negative at $109^{\circ}E$ while at $105^{\circ}E$ the anomaly almost zero, therefore the upwelling in 1959 and 1960 are stronger in the eastern part and almost normal in the western part (see also climatological SST condition as shown in Figure 6.1). On the other hand, the weakness of zonal wind (positive anomaly) during the SE monsoon (Figure 7.3) in 1979 and 1980 caused the weakness of upwelling.

When the monsoon during the normal condition is weak, the propagating of upwelling to the west becomes slower. Unfortunately, the local monsoon index for Indonesia is not available yet so that we do not know exactly the strength of the monsoon. In this study we identified that the strength of the monsoon is depend on the strength/weakness of the wind and the strength of the upwelling is influenced by the zonal wind in the south of Java.


Figure 7.1: Anomaly of SST (1959-2002) in cross section AB (see Figure 3.2)



Figure 7.2: Anomaly of SSS (1959-2002) in cross section AB (see Figure 3.2)



Figure 7.3: Anomaly of zonal wind (m/s) during the SE monsoon (May-October) 1959-2002. The positive (negative) sign means that the wind blows stronger eastward (westward).

7.1.1 During ENSO

Figures 7.4, 7.5, 7.6 show that during the strong El Niño 1982/1983, the upwelling was stronger than during other El Niño events. Starting from May 1983, cooler water mass moved upward in the south coast of Java island and transported westward. In the south coast of East Java, the SST and SSS anomalies were more than $-1.0^{\circ}C$ and +0.2psu, respectively, and can be identified until the end of June, while in the south coast of Central Java the anomalies were $-1.5^{\circ}C$ and more than +0.1psu, respectively, and can be identified until August. In the Sunda Strait, the SST anomaly was $-0.6^{\circ}C$ and occurred between July and August or during the peak of SE monsoon.

Although 1997 has been the strongest El Niño year, its effect in the study area is not so obvious (the explanation is given in Section 7.1.2).

In the contrary, the SST in the southern coast of East and Central Java has been warmer during the strong La Niña 1998 with anomaly bigger than $+2.5^{\circ}C$ (Figure 7.1). Starting from the middle of April 1998, the SST anomaly increased by $+0.6^{\circ}C$ and became warmer to $+1.5^{\circ}C$ in May and reached its maximum in August with anomaly $+2.5^{\circ}C$. The SSS anomaly decreased -0.1psu in the end of May and the peak of minimum -0.2psu occurred in August in the eastern part, while in the western it occurred in September about -0.3psu. Warm SST existed in the western part during the strong La Niña 1998 from the middle of June until October 1998 with anomaly of $+1.5^{\circ}C$ (Figure 7.4, 7.5, 7.6). These conditions can be explained also by the reducing of zonal wind in the Indian Ocean. The warm conditions in the eastern Indian Ocean is caused not only by the transport of warm water mass from the Pacific due to the strong La Niña event but also by the decreasing of zonal wind in the Indian Ocean that caused downwelling Kelvin wave in the Indonesian waters (*Webster, et al.*, 1999).

7.1.2 During DME

In 1961, the DME coincided with the normal condition, when the El Niño or La Niña event did not occur, the SST decreased in the western part by more than $-1.5^{\circ}C$ in August and September 1961



(a)

0



Figure 7.4: Vertical profile of temperature (${}^{o}C$) and Salinity (psu) anomalies close to the southern coast of East Java at $112^{o}E$





Figure 7.5: Vertical profile of temperature (${}^{o}C$) and Salinity (psu) anomalies close to the southern coast of Central Java at 109 ${}^{o}E$, period 1959 - 1980 and 1981 - 2002 image a, c and b, d respectively



Figure 7.6: Vertical profile of temperature (${}^{o}C$) and Salinity (psu) anomalies near to the Sunda Strait, south of Sumatra at $105{}^{o}E$, period 1959 - 1980 and 1981 - 2002 image a, c and b, d respectively



Figure 7.7: Wavelet analysis of SST in the south of East Java (112°E, 8.5°S) with significant levels 95%

(Figure 7.4), but the increasing of SSS was only +0.03psu. *Saji* and *Yamagata* (2003) identified that the peak anomaly of positive DME 1961 began in April until the end of November as shown in Figure 7.4 with negative anomaly of temperature. Propagation of negative temperature anomaly was westward from May until November 1961. In July-Augustus the negative anomaly of temperature was stronger than $-1^{\circ}C$ at 109-110°*E* because of the upwelling due to SE monsoon, while at $105^{\circ}E$ in the western part the negative anomaly was stronger than $-1^{\circ}C$ in September. The SSS anomaly was only about +0.08psu.

The DME that occurred in the same years as the strong El Niño 1972 and 1997, have a different strength of influence in the eastern of Indian Ocean (Figure 7.4, 7.5, and 7.6). The influence of El Niño and DME 1972 occurred since March, when the SST in the eastern part decreased by $-0.9^{\circ}C$ and its reached its minimum $-1.8^{\circ}C$ in May, while the SSS increased by +0.15psu in March and reached maximum +2.9psu in November/December. In the south of Central Java, the maximum decreased of SST is about $-2.1^{\circ}C$ while the SSS increases by +0.25psu and was stronger than in the eastern part (Figure 7.5). Near the Sunda Strait or the western part of study area (Figure 7.6), the SST decreased by $-1.2^{\circ}C$ in September and October while the SSS decreased by -0.2psu.

On the other hand, during the El Niño and DME 1997, the decreasing/increasing of SST/SSS did not appear clearly in the eastern part (Figure 7.5). The upwelling, with SST decreased by $-1.2^{\circ}C$ and SSS increased by +0.2psu, occurred since April and appeared only in the south of Central Java (Figure 7.5). Near the Sunda Strait (Figure 7.6), from September to October, the SST anomaly was $-0.6^{\circ}C$ and the SSS anomaly was +0.05psu.

We expect that the effect of strong El Niño 1997 was not spread out to the Indian Ocean because of the decreasing of SE-wind. The SE wind become weaker than compare to the El Niño 1972 (Figure 7.3) in the southern coast of Java, so that the decreasing/increasing of SST/SSS in the eastern part of the Indian Ocean was reduced as shown in Figure 7.5. The increasing of SST/SSS in the western part was occurred mainly due to the strength of zonal wind that caused the DME which is reached the peak in October.

During weak La Niña 1967 (Saji et al., 1999), which was detected as a normal condition by NOAA



Figure 7.8: Wavelet analysis of SST in the Sunda Strait, western part of study area $(105^{\circ}E, 5.5^{\circ}S)$ with significant levels 95%

(see Table 2.1), the SST decreased from the beginning of April and reach its minimum in June in the south of East Java by more than $-2.4^{\circ}C$ (Figure 7.4). In the south of Central Java, the minimum SST anomaly occurred in May and the anomaly was still negative until the end of October (Figure 7.5). In the western part of the study area, negative anomaly of SST occurred since the end of May until the beginning of October (Figure 7.6). *Saji* and *Yamagata* (2003) identified also that the negative SST anomaly in the eastern box of Indian Ocean ¹ occurred from April and reached its peak in November.

7.1.3 Wavelet Analysis

The interdecadal fluctuations have the effect of modulating the amplitude and the frequency of occurrence of ENSO and DME events (*Torence* and *Compo*, 1998). Using the wavelet analysis, the signal of ENSO/DME events in this study area is identified.

The SE wind that parallel to the south Java coast triggered the upwelling and caused the cold water mass transported to the west. Therefore, the El Niño events normally will increase the strength of upwelling during the SE monsoon. With the increasing of westerly wind during the DME, the westward transport of upwelled water mass is also increased and affect to the negative anomaly in the west of Sumatra (eastern box of Indian Ocean as defined by *Saji, et al.*, 1999). Thus the interaction between the SE wind and ENSO/DME in this area is the most important factors that increased/decreased the upwelling.

In the eastern part of the study area (see Figure 7.7), the ENSO signals with period of 2-8 years are found during 1966-1983 (red contour). The signals with period of 1-2 years appeared only in 1973-1976, while the period of 1 year is significant in 1969. There are also ENSO signals in 1998 until 2002, but not significant in this area. The insignificant signals support our expectation in Section 7.1.2, that the strong ENSO are not always spread out to the Indian Ocean.

¹The eastern box of Indian Ocean is between 0^{o} - $10^{o}S$ and 90^{o} - $110^{o}E$



Figure 7.9: Wavelet analysis of zonal wind in the Sunda Strait, western part of study area (105°E, 5.5°S)

The ENSO signals can make a superposition with the DME signals in the western part near the Sunda Strait (see Figure 7.8). The significant power spectral with a period of 4-8 years was found during 1962-1983. The signals with a period of 1 year appear significantly in 1967, 1984, and 1996/1997.

From the wavelet analysis we found that the ENSO signals in the NINO 3.4 region 2 in the Pacific Ocean as showed in Table 2.1 are not always exist in the study area. As an example, El Niño 1982 that is identified as a strong El Niño, the signal did not appear in this study area. Reversely, the influence of La Niña 1998 was strong and seen almost in all study area (see also Figure 7.14). This condition became stronger by decreasing of zonal wind in 1998. Figure 7.9 and 7.10 show that during 1960-1964 and 1966-1978 the zonal wind is increased to the west (negative anomaly) than normal and stronger from eastern to western part of study area. Reversely, during 1993-2000 it is decreased (positive anomaly) than normal and more decreasing in the eastern part of study area.

7.2 Variability in the Java Sea

In the Java Sea the water are well mixed to depth and almost homogen (see Section 6.2). The ENSO influences the Java Sea through the Flores Sea in the eastern and Southeast China Sea in the northern part, meanwhile the influence of DME through the Sunda Strait is relatively small.

From 44 years simulations result, the range of SST and SSS anomalies along the cross section GH are normally only $\pm 0.5^{\circ}C$ and ± 0.2 psu respectively, as shown in Figures 7.11 and 7.12 (see also Figure 3.2). The maximum/minimum SSS anomalies occur always in the north coast of Java (H). The SSS at H is more vary than in the south coast of Kalimantan (G). We suggest that small variation of SSS at G is due to the river discharges.

During the El Niño 1982/83, the SST in the Java Sea is decreased by $-1^{\circ}C$ from the mean condition, while during the El Niño 1997 the decreasing of SST is quite small. This condition occurred

²NINO3.4 covers area between $5^{\circ}N - 5^{\circ}S$ and $120 - 170^{\circ}W$



Figure 7.10: Wavelet analysis of zonal wind in the south of Central Java (109°*E*, 8.5°*S*)

probably because the SE monsoon during 1997/98 is weaker than during 1982/83. The El Niño 1982, which is occurred at the end of NW monsoon (April), caused a big SSS variation in the north coast of Java (more than +1.0psu). It occurred also during the El Niño 1992/93, with a variation of SSS more than +0.8psu. Reversely the low SSS anomaly of -0.8psu occurred at the end of 1996.

During strong La Niña, warm water mass is transported from east to the Java Sea. The maximum SST anomaly was found in 1998 as big as $+2.5^{\circ}C$. In another La Niña events, such as 1988/1989 and 1999/2000, the SST anomaly is only between +1.0 to $+1.5^{\circ}C$. Figure 7.12 shows that strong variations of SST during the La Niña 1998 only affect to the SSS anomaly by +0.4psu.

Figure 7.14 shows that in the Java Sea (point H), the ENSO signals during 1965-1984 and 1996-2002 are quite significant, while during 1985-1995 the signals are weak. In 1965-1986, the ENSO signals with period of 4-8 years are strong while signals with period of 1-2 years are seen in 1965-1970, 1983-1987, and 1997-2002 and maximum in 1984.

7.3 Transport Variation through the Sunda Strait

Figures 7.13 and 7.15 show that the variation of SST and SSS in the Sunda Strait (point C') are more varied and rapid in month. Figure 7.6 shows the variability of temperature and salinity at C' above 110m depth. Comparing to the other Figures (7.4 and 7.5), the anomaly of salinity in the Sunda Strait is more varied because of the influence of water mass from the Java Sea and west coast of Sumatra to the Indian Ocean and also the upwelling processes in the eastern Indian Ocean during the SE monsoon.

During the La Niña years, the SST in Sunda Strait (C') is warmer than normal and reversely during the DME and El Niño years. These occur not only because of warm water mass transport from the Java Sea and western coast of Sumatra but also because of the decreasing of SE-wind, that it held the role.



Figure 7.11: SST Anomaly (1959-2002) in cross section GH (see Figure 3.2)



Figure 7.12: Anomaly of SSS (1959-2002) in cross section GH (see Figure 3.2)



Figure 7.13: Anomaly of SST (1959-2002) in cross section EF (see Figure 3.2)



Figure 7.14: Wavelet Analysis of SST in the Java Sea (109°E, 5.5°S) with significant levels 95%

During the DME years, strong SE-wind causes strong upwelling in the western coast of Sumatra around the Sunda Strait. The minimum anomaly occurs in 1961, 1967 and 1972 during the DME, DME coincide with weak La Ni ña and DME coincide with strong El Niño, respectively. Nevertheless, the SST anomaly can be identified obviously than the SSS anomaly (Figure 7.13 and 7.15), likewise the vertical profiles (Figure 7.6).

During pure strong El Niño 1976, 1982, 1991, 1994 and 1997, the SST anomalies less than -0.6, -0.6, -0.4, -0.4, and -0.2°C respectively are occurred from June until the end of September. In the meantime, the SSS anomaly is not significant due to the mixing of Java Sea water mass. Reversely during La Niña 1973 and 1998, the temperature anomaly in May and June is more than +0.6°C in 30m depth while at the surface reaches +1.4°C.

The influence of DME in the Sunda Strait can be identified clearly and reaches its maximum in October. Likewise, the influence of La Niña is also obvious in the whole year and its influence is stronger with the decreasing of zonal wind in the western part. On the other hand, the El Niño that appears in the beginning of SE monsoon is not always detectable. The strong El Niño can not be identified around the Sunda Strait if the monsoon is not strong enough to trigger the influences of El Niño to the west of Indonesian waters.

The other effect of DME and ENSO in the Sunda Strait is the increase/decrease of water mass transport. Normally, the water mass is transported from the Java Sea to the Indian Ocean (see Section 6.3) where the water mass with high temperature and low salinity from the Java Sea enters the eastern Indian Ocean. During the NW monsoon, the transport can reach the southern coast of East Java while during the SE monsoon reached only the southern coast of Central Java (see Chapter 6). The most important factor that affect the water mass transport is sea level difference between the Java Sea and the eastern Indian Ocean (see Section 5.5).

During the normal conditions April 1978 until December 1979 and April 1980 until December 1981, the minimum transport (about 0.3-0.4Sv) is occurred in December and January (NW monsoon) and the maximum transport (about 0.7-0.8Sv) is occurred in July and August (SE monsoon). The transport during the SE monsoon is strengthened by the strong SE-wind that blows to the west



Figure 7.15: Anomaly of SSS (1959-2002) in cross section EF (see Figure 3.2)



Figure 7.16: Variations of water mass transport through the Sunda Strait during normal conditions, ENSO, and DME. Negative sign means that the water mass is transported into the Indian Ocean

and impact to the Ekman transport perpendicular to the left of wind direction in southern hemisphere. However, the water mass is still transported into the eastern Indian Ocean during the NW monsoon (see Figure 7.16). This condition occurs because of SLH difference between the Java Sea and the eastern Indian Ocean where the role of Ekman transport is not strong enough to force the current entered the Java Sea.

In general, the effect of DME is strengthened the transport from the Java Sea to the eastern Indian Ocean by about 0.15-0.2Sv. The transport during the DME 1961, which coincided with the normal condition, was stronger than DME 1967 and DME 1972, which coincided with weak La Niña and strong El Niño, respectively (Figure 7.16).

During the La-Niña years, the surface wind in the Pacific Ocean is strengthened so that the sea level in Indonesian waters is higher. Therefore the difference of SLH between the Java Sea and eastern Indian Ocean is not as large as normal condition (for example Figure 5.16 b and d). As a result, the transport of water mass through the Sunda Strait is also reduced during these events.

During the El Niño years, the sea level in the eastern Indonesian waters is lower because of the weakness of equatorial wind in the Pacific Ocean. The sea level in the Java Sea is lower than in the eastern Indian Ocean (Figure 5.16 a). But unfortunately we can not generalize that condition for all El Niño events. In the El Niño 1997, which coincide with the DME, the difference of sea level is not so significant (Figure 5.16 c) and the zonal component of SE-wind takes a place as the most important factor here. In the eastern part (at $109^{\circ}E$), the zonal component of SE-wind is weaker than normal (indicated as positive anomaly) while at $105^{\circ}E$ is almost normal and become stronger in the Indian Ocean at $85^{\circ}E$. Nevertheless, the water mass in the Java Sea is still transported to

the western part of eastern Indian Ocean. We suggested that the local influences, such as wind, are more dominant in affecting the transport variability in the Sunda strait.

Chapter 8 Conclusions

The seasonal variation of the Java Sea was induced mainly by the monsoon. There are four conditions existing in a year: NW monsoon (December-March), SE monsoon (June-September), transition from NW to SE monsoon (April-May) and transition from SE to NW monsoon (October-November). These conditions result in the variation of sea surface current circulation and total transport in the Java Sea. During the NW monsoon, the maximum current velocity is 0.75m/s and points eastward while during the SE monsoon the maximum is 0.25m/s and points westward. In January (NW monsoon), the water mass is transported from the Karimata Strait to the Java Sea by (+)2.1Sv. This water mass is then transported out to the eastern Java Sea and through the Sunda Strait by (-)1.6Sv and (-)0.5Sv, respectively. On the contrary, in August (SE monsoon), the total transport in the Java Sea from the eastern part is (+)1.0Sv which is transported out through the Sunda Strait and the Karimata Strait by (-)0.7Sv and (-)0.3Sv, respectively. The SST and SSS in the Java Sea are $28^{\circ}C$ and 32.6psu respectively during the NW monsoon while during the SE monsoon, the SST increases to $28.5 - 29^{\circ}C$ and the SSS decreases to 32.2psu.

In the Sunda Strait, the water mass was transported from the Java Sea to the Indian Ocean with a range of variation between 0.48Sv (minimum, in December) and 0.72Sv (maximum, in August/September). This transport brings water mass with low salinity and high temperature to the Indian Ocean and is mixed with the water mass flowing from the western coast of Sumatra eastward along the south coast of Java. This is known as the south Java current with a mean velocity of 0.1m/s during the NW monsoon. During the SE monsoon, the westward south Java current merges the current from the western coast of Sumatra around the Sunda Strait (105-106°E and 6.5-7.5°S). The climatological condition shows that the influence of the Java Sea water mass, with low salinity and high temperature, in this strait is dominant and relative constant during the whole year. Normally, the upwelling occurs during the SE monsoon along the southern Java coast, and its propagation depends on the strength/weakness of the zonal wind. Stronger and more parallel the SE wind to the southern Java coast cause stronger upwelling. The deeper water mass with temperature lower than 26° C (occurring in August) and salinity higher than 34psu (occurring in October) is moved upward to the surface due to the upwelling process.

The ocean climate varies from 1959 to 2002. The strength of monsoon, DME, El Niño, and La Niña influence the ocean variability in the study area. In normal conditions the anomaly of SST and SSS are $\pm 0.5^{\circ}C$ and ± 0.2 psu, respectively. The anomaly of SST in the Java Sea decreases by (-)1°C due to the El Niño and increases by (+)1.5°C because of La Niña conditions with a maximum of (+)2.5°C in La Niña 1998. The influences of DME can be identified in the Java Sea, although there is no inflow from the Indian Ocean through the Sunda Strait due to the permanent north-south gradient in SLH. During DME, the increasing of zonal wind causes the sea level in the western part of the eastern Indian Ocean lower than in the Java Sea. Therefore, the transport of cooler and more saline water mass from the eastern part of the Java Sea increases. The anomaly of SST in the Java Sea decreases by (-)1.5°C.

The propagation and advection of upwelled water mass depend on the strength of the SE wind south of Java and its interaction with DME and ENSO events. Moreover, the DME and ENSO signals south of Java can not always be identified by the SST anomaly. During the DME 1961 that occurred in normal condition, the upwelling in the western part of the eastern Indian Ocean was stronger than normal. During the La Niña years, warm water from the eastern part of the eastern Indian Ocean is transported westward. The effect of La Niña 1998 in the study area is stronger

than La Niña 1975 and 1988, even though the La Niña 1988 was the strongest event in the Pacific Ocean. These conditions occurred because in the same time as La Niña 1988, the zonal wind during the SE monsoon is reduced and weaker than during La Niña 1975 and 1998. Normally, during El Niño years, the strength of upwelling in the study area is increased. Nevertheless it did not occur in the strongest El Niño 1997, where the zonal wind in the eastern part of the eastern Indian Ocean was decreased and caused the weakening of upwelling during the SE monsoon. On the other hand, the location of the upwelling moved to the west coast of Sumatra because of the strengthened zonal wind in the western part of the study area during the DME.

Generally, the HAMSOM can simulate the ocean variability and the influence of the interaction between the monsoon, ENSO and DME in the study area. From this study we found that the variation of SLH difference between the Indian and Pacific Oceans affects the ocean dynamics of the Indonesian waters. Although the SLH difference variation is only 5 to 15cm, the effect is very significant to the simulation result. Unfortunately, this variation can only be measured by satellite altimetry and did not apply yet as an input variable in HAMSOM. The best result we get now is by applying a constant MSL from *Wyrtki* (1961). Several ENSO and DME events could not be identified with the SST/SSS anomalies in this area. We expected that the use of climatological sea water temperature and salinity as boundary values could not represent the actual change of those parameters to time.

To get more realistic results, continuation of this study is needed by including the sea level as an input variable. Moreover, the air-sea couple model is needed to obtain more realistic variations of temperature and salinity to time at the open boundaries.

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Appendix A List of Symbols

$A_{Ih} = A_{Ih}, A_V$	horizontal and vertical eddy viscosity coefficients $(m^2 s^{-1})$
A_{Mhk}	in salinity transport
A_0	value of A_V in unstratified state $(m^2 s^{-1})$
a,b,c	coefficients in the vertical semi-implicit system B.41
alb	albedo (%)
ă,Ď,č	coefficients in the vertical semi-implicit system B.46
$\hat{a}, \hat{b}, \hat{c}$	coefficients in the horizontal semi-implicit system B.44
$\tilde{a}, \tilde{b}, \tilde{c}$	coefficients in the T-S semi-implicit system B.47
В	right hand side of elliptic system B.35
C_d	drag coefficient at the air-sea interface
C_N, \ldots, C_E, C_C	coefficients in the horizontal elliptic
C_H, C_L	the bulk coefficients
$C(p_{air})$	the specific heat of air $(J kg^{-1} \sigma K^{-1})$
cn	cloud cover (%)
D, E	auxilary parameter
d_k	depth in the k-layer (m)
d^x, d^y	right hand side of the U-V tridiagonal system B.41
\check{d}^T,\check{d}^y	right hand side of the system B.46
\hat{d}^T, \hat{d}^y	right hand side of the system B.47
d^T, d^S	right hand side of the T-S tridiagonal
F,F_C	friction function and drag coefficient at the seabed
F_T	tidal form number
F_x, F_y	external forces in the x and y directions (ms^{-1})
f	coriolis parameter (radian s^{-1})
g	gravitational acceleration (m s^{-2})
H	water depth below the mean sea level (m)
h	model vertical layer thickness (m)
h_L	mixing layer thickness (m)
h_B	the decay length scale for the absorption of solar radiation
$h_B 1$	first depth of downward irradiance (m)
$h_B 2$	second depth of downward irradiance (m)
Ι	internal (baroclinic) pressure component
i,j,k	index of (y,z,x) direction in the numerical discretization
Κ	the von Karman constant, $K \cong 0.4$ (Vihma, 1995)
K_H, K_v	horizontal and vertical eddy diffusivities $(m^2 s^{-1})$
K_0	value of K_v in unstratified state $(m^2 s^{-1})$
L	Obukhov-length (Obukhov, 1946 in Vihma, 1995)
n	time-level index
norm	index of normal component
O^S, O^T	further optimal terms of temperature and salinity
$p_a, p^{'}(\zeta)$	air pressure and its anomaly (Pa)
р	pressure at depth z (Pa)
Q_k	total heat flux in layer k (W m^{-2})
Q_{LH}	the latent heat flux (W m^{-2})

0 _{uw}	the longwave radiation (W m^{-2})
	the sensible heat flux (W m^{-2})
	the solar radiation (W m^{-2})
	air specific humidity (kg/kg)
	specific humidity closed to the surface of seawater (kg/kg)
Ri	bulk of Richardson number
S	salinity (nsu)
S. J.	climatological salinity (nsu)
Sciim Sc	solar constant (W m^{-2})
Sol	salinity on the boundary where they get from WOA data (psu)
ST Sc	source term of temperature $({}^{o}C)$ and salinity (psu)
$\tilde{S}_{M}\tilde{S}_{U}$	source term of temperature (${}^{\circ}C$) and salinity (psu)
<i>5м</i> , <i>5н</i> Т	sequater temperature $\binom{0}{1}$
T T	air temperature (${}^{o}K$ converted to ${}^{o}C$)
T_{0}	a reference air temperature (${}^{0}K {}^{0}C$)
T_1 T_2	$(2x^2)$ coriolis matrix in the equation
Td	the damping timescale
Tu Tulin	climatological temperature (^{o}C)
T _{ch}	temperature on the boundary where they got from WOA data (${}^{o}C$)
$t \cdot \Delta t$	time-axis and time-step (s)
(U,V)	depth mean transport (m^2s^{-1}) in layer i
(;;,,,)) II.V.W	velocities component (ms^{-1}) in the x-y-z direction
un _{ob}	normal velocity (ms^{-1}) on the bounderies
и	a wind velocity scaling parameter known as the friction velocity
W, W^x, W^y	wind scalar and wind speed (ms^{-1}) in the x and y direction
X.Y	further optional terms in momentum equation $(m^2 s^{-2})$
X,V,Z	east-, north-, upwards-coordinate (m)
$\Delta x, \Delta y, \Delta z$	space increments (m) in the (x,y,z) coordinate
20	arbitary reference level for vertical integration (m)
α, β	elements of rotation matrix T_1
η	solar elevation
γ	fraction of insolation
$\dot{\gamma}_{1}, \dot{\gamma}_{2}, \dot{\gamma}_{3}$	coefficient of the damping timescale, that they were applied in the HAMSOM
ρ_a	air density (kgm^{-3})
ρ, ρ_0, ρ'	actual, reference and anomaly density (kgm^{-3})
τ, τ_s, τ_b	shear stress, wind stress, and bottom stress $(m^2 s^{-2})$
ζ, δ	sea surface elevation and its increment (m)
Δ	vertical difference
∇	horizontal gradient operator
∇^2	horizontal Laplace operator
Φ	auxiliary variable used in the Eulerian-Lagrangian approximation
φ	any arbitrary variable of U,V,T, and S
Ψ	stability function in an integrated form (Vihma, 1995)
σ	the Stefan-Boltzmann constant $(Wm^{-2}K^{-4})$
θ	implicit factor
Θ	parameter used for bottom friction
8	emissivity of water

Appendix B HAMSOM - Detailed Description

The HAMSOM had developed since 1985 and applied not only for the shelf seas, but also for the deep sea. There are more modifications of the model due to the concentrations and the study area conditions. The basics of the model equations were adopted from Backhaus (1985), Huang (1995), Pohlmann (1987;1991;1996).

B.1 Layer-Averaged Equations

The equations of motion and continuity (see Chapter 4) are vertically integrated over a depth-range h, according to a computational model layer of the same thickness.

The horizontally transports are defined for every layer bellow:

$$U_{k} = \int_{d_{k}}^{z_{k-1}} u dz$$

$$V_{k} = \int_{d_{k}}^{z_{k-1}} v dz$$
(B.1)

With the Leibniz rules, the momentum and continuity equations can be convertioned as

$$\frac{\partial U_{k}}{\partial t} + \frac{\partial}{\partial x} \int_{d_{k}}^{d_{k-1}} uudz + \frac{\partial}{\partial y} \int_{d_{k}}^{d_{k-1}} uvdz - fV_{k} + \int_{d_{k}}^{d_{k-1}} \frac{1}{\rho} \frac{\partial p}{\partial x} dz =$$

$$= \frac{\partial}{\partial x} \int_{d_{k}}^{d_{k-1}} A_{Ih} \frac{\partial u}{\partial x} dz + \frac{\partial}{\partial y} \int_{d_{k}}^{d_{k-1}} A_{Ih} \frac{\partial u}{\partial y} dz + h_{k}F_{x} +$$

$$\left[A_{Iv} \frac{\partial u}{\partial z} - wu + u \frac{\partial d}{\partial t} + uu \frac{\partial d}{\partial y} + uv \frac{\partial d}{\partial y} + A_{Ih} \cdot \left(\frac{\partial u}{\partial x} \frac{\partial d}{\partial x} + \frac{\partial u}{\partial y} \frac{\partial d}{\partial y} \right) \right]_{d_{k}}^{d_{k-1}}$$
(B.2)

$$\frac{\partial V_{k}}{\partial t} + \frac{\partial}{\partial x} \int_{d_{k}}^{d_{k-1}} uvdz + \frac{\partial}{\partial y} \int_{d_{k}}^{d_{k-1}} vvdz + fU_{k} + \int_{d_{k}}^{d_{k-1}} \frac{1}{\rho} \frac{\partial p}{\partial x} dz =$$

$$= \frac{\partial}{\partial x} \int_{d_{k}}^{d_{k-1}} A_{Ih} \frac{\partial u}{\partial x} dz + \frac{\partial}{\partial y} \int_{d_{k}}^{d_{k-1}} A_{Ih} \frac{\partial v}{\partial y} dz + h_{k}F_{y} +$$

$$\left[A_{Iv} \frac{\partial v}{\partial z} - wv + v \frac{\partial d}{\partial t} + uv \frac{1}{u} v \frac{\partial d}{\partial x} + vv \frac{\partial d}{\partial y} + A_{Ih} \cdot \left(\frac{\partial v}{\partial x} \frac{\partial d}{\partial x} + \frac{\partial v}{\partial y} \frac{\partial d}{\partial y} \right) \right]_{d_{k}}^{d_{k-1}}$$
(B.3)

Term of u. $\left(\frac{\partial d}{\partial t} + u\frac{\partial d}{\partial x} + v\frac{\partial d}{\partial y}\right)$ and v. $\left(\frac{\partial d}{\partial t} + u\frac{\partial d}{\partial x} + v\frac{\partial d}{\partial y}\right)$ in equation B.2 and B.3 is zero only in the surface layer $(d = \zeta)$ and in the bottom (d = -H). Term of wu dan wv in equation B.2 and B.3

can be changed as a boundary condition in the equation 4.8. Thats means they can be changed as boundaries at the surface and bottom. So the vertical equation is :

$$w_{\zeta} = 0 \quad \text{and} \quad w_H = 0 \tag{B.4}$$

Likewise the terms A_{Ih} . $\left(\frac{\partial u}{\partial x}\frac{\partial d}{\partial x} + \frac{\partial u}{\partial y}\frac{\partial d}{\partial y}\right)$ and A_{Ih} . $\left(\frac{\partial v}{\partial x}\frac{\partial d}{\partial x} + \frac{\partial v}{\partial y}\frac{\partial d}{\partial y}\right)$ dalam (3.15) dan (3.16) $A_{Ih}\frac{\partial u}{\partial z}$ dan $A_{Ih}\frac{\partial v}{\partial z}$ are omitted with the equation of boundaries at the surface $(d = \zeta)$ and at the bottom (d = -H).

The pulse equation can not be yet solved in this form. Therefore the following approximations will be used : For the term of advection :

$$\int_{d_{k}}^{d_{k-1}} uudz \equiv \frac{U_{k}U_{k}}{h_{k}}$$

$$\int_{d_{k}}^{d_{k-1}} uvdz \equiv \frac{U_{k}V_{k}}{h_{k}}$$

$$\int_{d_{k}}^{d_{k-1}} vvdz \equiv \frac{V_{k}V_{k}}{h_{k}}$$
(B.5)

Horizontal diffusions term :

$$\int_{d_k}^{d_{k-1}} A_{Ih} \frac{\partial u}{\partial x} dz \equiv A_{Ih} \frac{\partial U_k}{\partial x} \quad \text{and} \quad \int_{d_k}^{d_{k-1}} A_{Ih} \frac{\partial u}{\partial y} dz \equiv A_{Ih} \frac{\partial U_k}{\partial y}$$
(B.6)

$$\int_{d_k}^{d_{k-1}} A_{Ih} \frac{\partial v}{\partial x} dz \equiv A_{Ih} \frac{\partial V_k}{\partial x} \quad \text{and} \quad \int_{d_k}^{d_{k-1}} A_{Ih} \frac{\partial v}{\partial y} dz \equiv A_{Ih} \frac{\partial V_k}{\partial y}$$
(B.7)

and the Boussinesq-Aproximation:

$$\int_{d_k}^{d_{k-1}} \frac{1}{\rho} \frac{\partial p}{\partial x} \equiv \frac{1}{\rho_k} \int_{d_k}^{d_{k-1}} \frac{\partial p}{\partial x} dz$$
(B.8)
$$\int_{d_k}^{d_{k-1}} \frac{1}{\rho} \frac{\partial p}{\partial y} \equiv \frac{1}{\rho_k} \int_{d_k}^{d_{k-1}} \frac{\partial p}{\partial y} dz$$

At the sea surface the windstress cause of the turbulent. At the bottom the turbulent dissipations are approximated by the quadratic Newton-Taylor.

$$\tau_{\zeta}^{(x)} \equiv C_d W^{(x)} \sqrt{(W^{(x)})^2 + (W^{(y)})^2}$$

$$\tau_{\zeta}^{(y)} \equiv C_d W^{(y)} \sqrt{(W^{(x)})^2 + (W^{(y)})^2}$$

$$\tau_{-H}^{(x)} \equiv r u_{-H} \sqrt{(u_{-H})^2 + (v_{-H})^2}$$

$$\tau_{-H}^{(y)} \equiv r v_{-H} \sqrt{(u_{-H})^2 + (v_{-H})^2}$$

(B.9)

With all of the approximation above, the momentum equation in x-coordinate for k-layer :

$$\frac{\partial U_k}{\partial t} + \frac{\partial}{\partial x} \left(\frac{U_k U_k}{h_k} \right) + \frac{\partial}{\partial y} \left(\frac{U_k V_k}{h_k} \right) - fV_k + \frac{1}{\rho_k} \int_{d_k}^{d_{k-1}} \frac{\partial p}{\partial x} dz =$$

$$= \frac{\partial}{\partial x} \left(A_{Ih} \frac{\partial U_k}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_{Ih} \frac{\partial U_k}{\partial y} \right) + h_k F_x + \left[A_{Iv} \frac{\partial u}{\partial z} - wu \right]_{d_k}^{d_{k-1}}$$
(B.10)

and in y-coordinate for k-layer :

$$\frac{\partial V_k}{\partial t} + \frac{\partial}{\partial x} \left(\frac{V_k U_k}{h_k} \right) + \frac{\partial}{\partial y} \left(\frac{V_k V_k}{h_k} \right) + f U_k + \frac{1}{\rho_k} \int_{d_k}^{d_{k-1}} \frac{\partial p}{\partial y} dz =$$

$$= \frac{\partial}{\partial x} \left(A_{Ih} \frac{\partial V_k}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_{Ih} \frac{\partial V_k}{\partial y} \right) + h_k F_y + \left[A_{Iv} \frac{\partial v}{\partial z} - wv \right]_{d_k}^{d_{k-1}}$$
(B.11)

By definition the velocities at the k-layers are :

$$u_{k} \equiv \frac{U_{k}}{h_{k}}$$

$$v_{k} \equiv \frac{V_{k}}{h_{k}}$$
(B.12)

so the terms of horizontal advection at equation B.11 and B.12 can be changed with the continuity term at equation 4.4 as follows :

in the x-direction for k-layers :

$$\frac{\partial U_k}{\partial t} + u_k \frac{\partial U_k}{\partial x} + v_k \frac{\partial U_k}{\partial y} - U_k \frac{\partial w}{\partial z} - fV_k + \frac{1}{\rho_k} \int_{d_k}^{d_{k-1}} \frac{\partial p}{\partial x} dz =$$

$$= \frac{\partial}{\partial x} \left(A_{Ih} \frac{\partial U_k}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_{Ih} \frac{\partial U_k}{\partial y} \right) + h_k F_x + \left[A_{Iv} \frac{\partial u}{\partial z} - wu \right]_{d_k}^{d_{k-1}}$$
(B.13)

in the y-direction for k-layers :

$$\frac{\partial V_k}{\partial t} + u_k \frac{\partial V_k}{\partial x} + v_k \frac{\partial V_k}{\partial y} - V_k \frac{\partial w}{\partial z} - fU_k + \frac{1}{\rho_k} \int_{d_k}^{d_{k-1}} \frac{\partial p}{\partial y} dz =$$

$$= \frac{\partial}{\partial x} \left(A_{Ih} \frac{\partial V_k}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_{Ih} \frac{\partial V_k}{\partial y} \right) + h_k F_y + \left[A_{Iv} \frac{\partial v}{\partial z} - wv \right]_{d_k}^{d_{k-1}}$$
(B.14)

Because of the term $\frac{\partial w}{\partial z}$ at the sea surface and at the bottom are small, the kinematic at the sea surface is changed to :

$$w_{\zeta} = \frac{\partial \zeta}{\partial t}$$

$$w_{-H} = 0$$
(B.15)

The dynamics boundary conditions are modificated by

$$A_{Iv} \frac{\partial u_{\zeta}}{\partial z} = C_d W^{(x)} \sqrt{(W^{(x)})^2 + (W^{(y)})^2}$$

$$A_{Iv} \frac{\partial v_{\zeta}}{\partial z} = C_d W^{(y)} \sqrt{(W^{(x)})^2 + (W^{(y)})^2}$$

$$A_{Iv} \frac{\partial u_{(-H)}}{\partial z} = r u_{-H} \sqrt{(u_{-H})^2 + (v_{-H})^2}$$

$$A_{Iv} \frac{\partial v_{(-H)}}{\partial z} = r v_{-H} \sqrt{(u_{-H})^2 + (v_{-H})^2}$$
(B.16)

and the continuity equation for k-layer is

$$\frac{\partial U_k}{\partial x} + \frac{\partial V_k}{\partial y} + \left[w - u \frac{\partial d}{\partial x} - v \frac{\partial d}{\partial y} \right]_{d_k}^{d_{k-1}} = 0$$
(B.17)

With the applications of boundary conditions at the surface and at the bottom, the continuity equation can be simplified

$$\frac{\partial U_k}{\partial x} + \frac{\partial V_k}{\partial y} + [w]_{d_k}^{d_{k-1}} = 0$$
(B.18)

The pressure, density, temperature and salinity are calculated at the every k-layers.

- for the pressure with the hydrostatic equation 4.3 :

$$\frac{\partial p_k}{\partial z} + g\rho_k = 0 \tag{B.19}$$

- for the density with the equation 4.7 :

$$\rho_k = \rho(S_k, T_k, p_k) \tag{B.20}$$

- for the temperature transport with the equation 4.5 :

$$\frac{\partial T_k}{\partial t} + u_k \frac{\partial T_k}{\partial x} + v_k \frac{\partial T_k}{\partial y} + w_k \frac{\partial T_k}{\partial z} =$$
$$= \frac{\partial}{\partial x} \left(A_{Mhk} \frac{\partial T_k}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_{Mhk} \frac{\partial T_k}{\partial y} \right) + \frac{\partial}{\partial z} \left(A_{Mhk} \frac{\partial T_k}{\partial z} \right) + R_{Tk}$$
(B.21)

- for the salinity transport with the equation 4.6 :

$$\frac{\partial S_k}{\partial t} + u_k \frac{\partial S_k}{\partial x} + v_k \frac{\partial S_k}{\partial y} + w_k \frac{\partial S_k}{\partial z} =$$
$$= \frac{\partial}{\partial x} \left(A_{Mhk} \frac{\partial S_k}{\partial x} \right) + \frac{\partial}{\partial y} \left(A_{Mhk} \frac{\partial S_k}{\partial y} \right) + \frac{\partial}{\partial z} \left(A_{Mhk} \frac{\partial S_k}{\partial z} \right) + R_{Sk}$$
(B.22)



Figure B.1: Horizontal Distributions of Grid-points in Arakawa C-grid (Backhaus, 1985) Dashed lines separate the grid-point and the grid increments along east and north axis Δx and Δy . The number 1 to 9 correspond to the indices of the coefficients c in equation B.36.

The Symbols of grid point :

+ Pressure-, Density, and $\boldsymbol{\zeta}$ point

x East (U) component of Transport

O North (V) component of Transport

B.2 The Numerical Scheme

The motion equations (in B.2 and B.3) are similar with the following equations, which used by Backhaus (1985) :

$$\frac{\partial}{\partial t} \begin{pmatrix} U \\ V \end{pmatrix} + \begin{pmatrix} 0 & -f \\ f & 0 \end{pmatrix} \begin{pmatrix} U \\ V \end{pmatrix} = -\frac{h}{\rho} \begin{pmatrix} \frac{\partial p}{\partial x} \\ \frac{\partial p}{\partial y} \end{pmatrix} + \begin{pmatrix} X \\ Y \end{pmatrix} + \begin{pmatrix} \Delta \tau^x \\ \Delta \tau^y \end{pmatrix}$$
(B.23)

where X and Y are the further optional terms, which are calculated with the explicit scheme.

The equation of temperature and salinity are written as follows :

$$\frac{\partial T}{\partial t} + w \frac{\partial T}{\partial z} = \frac{\partial}{\partial z} \left(K_v \frac{\partial T}{\partial z} \right) + O^T$$
(B.24)

$$\frac{\partial S}{\partial t} + w \frac{\partial S}{\partial z} = \frac{\partial}{\partial z} \left(K_v \frac{\partial S}{\partial z} \right) + O^S$$
(B.25)

where O^T and O^S are the further optional terms and sparated from B.24 and B.25 because they are calculated with the explicit scheme.

Determination of the vertical gradients $\frac{d\rho}{dz}$ und $\frac{du}{dz}$ in interface layer (for example in a x-z section) Grid-point symbols :

+ Pressure- und Density point

 \times east (U) transport component in *x*-coordinate

 \triangle calculation point of $\frac{d\rho}{dz}$

 \bigcirc calculation point of $\frac{du}{dz}$

 h_k The Tickness of k-th Layer

 hm_k Distance from the middle of (j-1)-th Layer to the middle of j-th Layer

 h_{Sp} vertical extending of interface layer

Ri Richardson Number

B.2.1 Numerical Coordinates

The HAMSOM model are discretized in the finite difference grid with the semi implicit and explicit numerical scheme. The model used the cartesian coordinates x, y, and z, which is on the east-west directions, on the north-south directions, and on the depth directions respectively. The layer-averaged equations above are solved on a three-dimensional grid, where x, y, and z variables are associated with the spatial index k,i, and j in the numerical scheme. The x-coordinate is positive in the east directions and it means that the k increases with increasing x. The positively of y-coordinate is in the north directions, therefore index i increases with decreasing y. And also the z-coordinate (layers) is positive in the depth directions. The increasing of index j means the decreasing of z. The variables are distributed over the grid according to the Arakawa C-grid, which they are visualised as a cube.

The time domain had a two time-level scheme, which are denoted with the n for the latter time index, which it is omitted for the sake of clarity, and n+1 for the future time-level, which is abbreviated to ⁽¹⁾. The prognostic variables as ζ , U, V, T, S are defined at staggerd grid time-levels. They enter to the semi implicit algoritma. Pressure is separated into its barotropic component $g\rho_1$ ζ and its baroclinic component I. A semi-implicit scheme is used in the barotropic pressure gradient terms and in the vertical shear stress terms (B.23) in the equations of motion, in the horizontal

divergence term in the equation of continuity, and in the vertical diffusion terms in the equations of temperature (B.24) and salinity (B.25).

B.2.2 Finite Difference Momentum Equations

According to the distribution of horizontal and vertical grid in the figure B.1 and B.2 the HAM-SOM equations can be discretized with the grouping of them.

$$\begin{pmatrix} U \\ V \end{pmatrix}_{j}^{(1)} = T_{1} \begin{pmatrix} U \\ V \end{pmatrix}_{j} - \Delta t \begin{pmatrix} h \\ \rho \end{pmatrix}_{j} \left[g\rho_{1}\theta_{1} \begin{pmatrix} \frac{\zeta_{x}}{\underline{\zeta_{y}}} \end{pmatrix}^{(1)} + g\rho_{1}(1-\theta) \begin{pmatrix} \frac{\zeta_{x}}{\underline{\zeta_{y}}} \end{pmatrix} + \begin{pmatrix} I_{x} \\ I_{y} \end{pmatrix}_{j} \right] + \Delta t \left[\begin{pmatrix} X \\ Y \end{pmatrix}_{j} + \begin{pmatrix} \Delta \tau^{x} \\ \Delta \tau^{y} \end{pmatrix}_{j}^{(1)} + (1-\theta) \begin{pmatrix} \Delta \tau^{x} \\ \Delta \tau^{y} \end{pmatrix}_{j} \right] = \begin{pmatrix} \hat{U} \\ \hat{V} \end{pmatrix}_{j} - \Delta t \theta g\rho_{1} \begin{pmatrix} h \\ \rho \end{pmatrix}_{j} \left(\frac{\underline{\delta x}}{\underline{\delta y}} \right) + \Delta t \theta \begin{pmatrix} \tau^{x}_{j} - \tau^{x}_{j+1} \\ \tau^{y}_{j} - \tau^{y}_{j+1} \end{pmatrix} B.26$$

with $T_1 = \begin{pmatrix} \alpha & \beta \\ -\beta & \alpha \end{pmatrix}$ and $\delta \equiv \zeta^1 - \zeta$.

The θ is the implicit weight factor and T_1 is the rotation matrix, where $\alpha = cos(f\Delta t)$ and $\beta = sin(f\Delta t)$. The underline terms with index of coordinates denotes the horizontal derivative with respect to the coordinate. The rotation matrix (T_2) applied to the pressure terms (Backhaus, 1985) is omitted, because the stability limitation in a shallow dissipative sea is less restrictive than the deep water shelf and shelfbreak where the T_2 should be included (Huang, 1995).

A quadratic stable semi-implicit friction form for the bottom layer is :

$$\begin{pmatrix} \tau^{x} \\ \tau^{y} \end{pmatrix}_{J+1}^{(1)} \equiv \begin{pmatrix} \tau^{x}_{b} \\ \tau^{y}_{b} \end{pmatrix}^{(1)} = F_{C}\Gamma \begin{pmatrix} U \\ V \end{pmatrix}_{J}^{(1)}$$
(B.27)

where for j = J, then $\Gamma = \sqrt{U_J^2 + V_J^2}/(H'h_J)$ and

$$\begin{pmatrix} U \\ V \end{pmatrix}_{J}^{(1)} = F\left[\begin{pmatrix} \hat{U} \\ \hat{V} \end{pmatrix}_{J} - \Delta t \theta g \rho_{1} \left(\frac{h}{\rho}\right)_{J} \left(\frac{\delta_{x}}{\underline{\delta_{y}}}\right) + \Delta t \theta \left(\frac{\tau_{J}^{x}}{\tau_{J}^{y}}\right)^{(1)}\right]$$
(B.28)

F is the friction function parameter and equal $1/(1 + \Delta t \theta F_C \Gamma)$. \hat{U} and \hat{V} are the sum of all explicit terms.

$$\begin{pmatrix} \hat{U} \\ \hat{V} \end{pmatrix}_{j} = T_{1} \begin{pmatrix} U \\ V \end{pmatrix}_{j} - \Delta t \begin{pmatrix} h \\ \rho \end{pmatrix}_{j} \left[g\rho_{1} \begin{pmatrix} \frac{\zeta_{x}}{\underline{\zeta_{y}}} \end{pmatrix} + \begin{pmatrix} \frac{I_{x}}{\underline{I_{y}}} \end{pmatrix}_{j} \right] + \Delta t \begin{pmatrix} X \\ Y \end{pmatrix}_{j} + \Delta t (1 - \theta) \begin{pmatrix} \tau_{j}^{x} - \tau_{j+1}^{x} \\ \tau_{j}^{y} - \tau_{j+1}^{y} \end{pmatrix}$$
(B.29)

A sum over the model layer J is approximation from the intergral of depth.

$$\begin{pmatrix} \bar{U} \\ \bar{V} \end{pmatrix} = \sum_{j=1}^{J} \begin{pmatrix} U \\ V \end{pmatrix}_{j}$$
(B.30)

And the equation of continuity in the equation 4.4 is approximated by

$$\zeta^{(1)} = \zeta \qquad \underbrace{-\Delta t \sigma(\underline{\bar{U}_x} + \underline{\bar{V}_y})^{(1)} - \Delta t(1 - \sigma)(\underline{\bar{U}_x} + \underline{\bar{V}_y})}_{\delta} \tag{B.31}$$

B.2.3 Horizontal Discretization

The equations above are discretized in horizontal directions (in x- and y-coordinates) with semiimplicit and explicit scheme.

B.2.3.1 Semi-Implicit System of U- and V-Transport

We separate the time levels n+1 and n and then sum the equation B.26. The equation B.30 can be written as follows:

$$\begin{pmatrix} \bar{U} \\ \bar{V} \end{pmatrix}^{(1)} = \begin{pmatrix} \tilde{U} \\ \tilde{V} \end{pmatrix} - \Delta t \theta g \rho_1 \left[\sum_{j=1}^{J-1} \left(\frac{h}{\rho} \right)_j + F \left(\frac{h}{\rho} \right)_J \right] \left(\frac{\underline{\delta}_x}{\underline{\delta}_y} \right)$$
(B.32)

where the first term in the right side equation B.32 is the sum of all the known terms

$$\begin{pmatrix} \tilde{U} \\ \tilde{V} \end{pmatrix} = \begin{bmatrix} J_{-1} \begin{pmatrix} \hat{U} \\ \hat{V} \end{pmatrix}_j + F \begin{pmatrix} \hat{U} \\ \hat{V} \end{pmatrix}_J \end{bmatrix} + \Delta t \theta \begin{pmatrix} \tau_s^x \\ \tau_s^y \end{pmatrix}^{(1)} - \Delta t \theta (1 - F) \begin{pmatrix} \tau_J^x \\ \tau_J^y \end{pmatrix}$$
(B.33)

and D is the auxiliary parameter in the second terms.

$$D \equiv \Delta t \theta g \rho_1 \left[\sum_{j=1}^{J-1} \left(\frac{h}{\rho} \right)_j + F \left(\frac{h}{\rho} \right)_J \right]$$
(B.34)

By inserting of equation B.32 and definition of D to the surface increment δ we obtain

$$\delta - \Delta t \theta \left(\underline{\bar{D}^{x}} \underline{\delta_{x}}_{x} + \underline{\bar{D}^{y}} \underline{\delta_{y}}_{y} \right) = -\Delta t \theta (\underline{\tilde{U}}_{x} + \underline{\tilde{V}}_{y}) - \Delta t (1 - \theta) (\underline{\bar{U}}_{x} + \underline{\bar{V}}_{y})$$
(B.35)

The coefficients can be defined as follows

$$C_{N} = \Delta t \theta \bar{D}_{i-1}^{y} / (\Delta y)^{2}$$

$$C_{S} = \Delta t \theta \bar{D}_{k}^{y} / (\Delta y)^{2}$$

$$C_{W} = \Delta t \theta \bar{D}_{k-1}^{x} / (\Delta x)^{2}$$

$$C_{E} = \Delta t \theta \bar{D}_{k}^{x} / (\Delta x)^{2}$$

$$C_{C} = C_{N} + C_{S} + C_{W} + C_{E}$$
(B.36)

where the overbar with a index-coodinate denote the spatial averages on the respecting coordinate and we obtain the elliptic system of δ

$$(1+C_C)\delta - C_N\delta_{i-1} - C_S\delta_{i+1} - C_W\delta_{k-1} - C_E\delta_{k+1} = B$$
(B.37)

and B is the right term in the equation B.35

The equation B.37 is solved by the successive over relaxation (SOR) incorporating with the appropriate boundary conditions (Huang, 1995), then the increment δ and surface elevation $\zeta^{(1)}$ are obtained.

B.2.3.2 Explicit System of Diffusion and Advection

The explicit system is used in the horizontal diffusion and advection terms of momentum, temperature and salinity. The explicit scheme is

$$\Delta t A_H \nabla^2 \phi = \Delta t A_H \left(\frac{\phi_{k+1} - 2\phi_k + \phi_k - 1}{\Delta x^2} + \frac{\phi_{i-1} - 2\phi_i + \phi_i + 1}{\Delta y^2} \right)$$
(B.38)

where ϕ represents variables of U, V, T, and S. The system is treated with the optional of vectorupstream scheme and the Eulerian-Lagrangian approximation (Huang, 1995)

B.2.4 Vertical Discretization

The vertical discretization used the semi-implicit system. The equations of heatflux on the sea surface are calculated in the Appendix B.

B.2.4.1 Semi-Implicit System of Vertical Component Velocity w

The vertical semi-implicit system for U and V in equation B.26 is the left term and the third term on the right side and they are equal with the first and second term of equation B.28, which are calculated in the increment of δ . We wrote

$$\begin{pmatrix} U \\ V \end{pmatrix}_{j}^{(1)} - \Delta t \theta \begin{pmatrix} \tau_{j}^{x} - \tau_{j+1}^{x} \\ \tau_{j}^{y} - \tau_{j+1}^{y} \end{pmatrix}^{(1)} = \begin{pmatrix} \hat{U} \\ \hat{V} \end{pmatrix}_{j} - \Delta t g \rho_{1} \left(\frac{h}{\rho} \right)_{j} \left(\frac{\underline{\delta}_{x}}{\underline{\delta}_{y}} \right)$$
(B.39)

And the aproximation of vertical shear stress (τ) is

$$\begin{pmatrix} \tau^{x} \\ \tau^{y} \end{pmatrix}_{j}^{(1)} = \frac{2Av_{j}}{h_{j-1} + h_{j}} \left(\begin{pmatrix} U/h \\ V/h \end{pmatrix}_{j-1} - \begin{pmatrix} U/h \\ V/h \end{pmatrix}_{j} \right)^{(1)}$$
(B.40)

From B.39 and B.40 the system of U and V are calculated

$$-a_{j} \begin{pmatrix} U \\ V \end{pmatrix}_{j-1}^{(1)} + b_{j} \begin{pmatrix} U \\ V \end{pmatrix}_{j}^{(1)} - c_{j} \begin{pmatrix} U \\ V \end{pmatrix}_{j+1}^{(1)} = \begin{pmatrix} d^{x} \\ d^{y} \end{pmatrix}_{j} \quad for \quad j = 1, 2, \dots, J \quad (B.41)$$

The equation B.41 is solved with the tridiagonal solving algorithm with the coefficients of the system are : $a_1 = c_J = 0$

$$\begin{aligned} a_j &= 2\Delta t \theta A v_j / [h_{j-1}(h_j + h_{j-1})] & \text{for } j = 2, 3, ..., J - 1 \\ a_J &= Fa_j = J \\ c_j &= 2\Delta t \theta A v_{j+1} / [h_{j+1}(h_j + h_{j+1})] & \text{for } j = 1, 2, ..., J - 1 \\ b_j &= 1 + (a_j h_{j-1} + c_j h_{j+1}) / h_j & \text{for } j = 1, 2, ..., J \\ (d^x, d^y)_j &= (T^x, Ty)_j & \text{for } j = 2, 3, ..., J \\ (d^x, d^y)_1 &= (T^x, Ty)_1 + \Delta t \theta (\tau^x_s, \tau^y_s)^{(1)} \\ (d^x, d^y)_J &= F(T^x, Ty)_{j=J} \end{aligned}$$

From the definition equation B.2 the horizontal velocity u and v are calculated as :

$$u_j^{(1)} = U_j^{(1)}/h_j,$$

 $v_j^{(1)} = V_j^{(1)}/h_j,$ for $j = 1, 2, ..., J$ (B.42)

and the vertical component of velocity w is calculated as following discretization :

$$w_{j} = w_{j+1} - \left(\frac{U_{k} - U_{k-1}}{\Delta x} + \frac{V_{i-1} - V_{i}}{\Delta y}\right)_{j}, \quad for \quad j = J, J - 1, \dots, 1$$
(B.43)
B.2.4.2 Semi-Implicit System of Diffusion and Advection

Similar with the equation of motion, in the equation of Temperature and Salinity the vertical diffusion is solved with the semi-implicit system. The semi-implicit system resulted more stabil than the explicit system. The equations of temperature and salinity are wrote as :

$$-\hat{a}_{j} \begin{pmatrix} T\\S \end{pmatrix}_{j-1}^{(1)} + \hat{b}_{j} \begin{pmatrix} T\\S \end{pmatrix}_{j}^{(1)} - \hat{c}_{j} \begin{pmatrix} T\\S \end{pmatrix}_{j+1}^{(1)} = \begin{pmatrix} \hat{d}^{T}\\\hat{d}^{S} \end{pmatrix}^{(1)}, \quad \text{for} \quad j = 1, 2, ..., J \quad (B.44)$$

where the coefficients are : $\hat{a}_1 = \hat{c}_j = 0$, $\hat{a}_j = 2\Delta t \theta K v_j / [h_j(h_j + h_{j-1})]$, for j = 2, 3, ..., J $\hat{c}_j = 2\Delta t \theta K v_{j+1} / [h_j(h_j + h_{j+1})]$, for j = 1, 2, ..., J - 1 $\hat{b}_j = 1 + \hat{a}_j + \hat{c}_j$, for j = 1, 2, ..., J $\hat{d}_j^T = T_j + \Delta t O_j^T + [\hat{a}_j T_{j-1} - (\hat{a}_j + \hat{c}_j) T_j + \hat{c}_j T_{j+1}] (1 - \theta) / \theta$, for j = 1, 2, ..., J $\hat{d}_j^S = S_j + \Delta t O_j^S + [\hat{a}_j S_{j-1} - (\hat{a}_j + \hat{c}_j) S_j + \hat{c}_j S_{j+1}] (1 - \theta) / \theta$, for j = 1, 2, ..., J

A semi-implicit system is applied to the vertical advection term in the temperature and salinity equations. They are solved by central difference. For the vertical layer j :

$$\frac{T_{j}^{(1)}}{\Delta t} = -\frac{1}{4}(w_{j} + w_{j+1}) \left[\left(\frac{\partial T}{\partial z} \right)_{j}^{(1)} + \left(\frac{\partial T}{\partial z} \right)_{j} \right]$$
$$= -\frac{w_{j} + w_{j+1}}{4h_{j}} \left[\frac{T_{j-1}^{(1)}h_{j} + T_{j}^{(1)}h_{j-1}}{h_{j-1} + h_{j}} - \frac{T_{j}^{(1)}h_{j+1} + T_{j+1}^{(1)}h_{j}}{h_{j} + h_{j+1}} \right]$$
$$+ \left[\frac{T_{j-1}h_{j} + T_{j}h_{j-1}}{h_{j-1} + h_{j}} - \frac{T_{j}h_{j+1} + T_{j+1}h_{j}}{h_{j} + h_{j+1}} \right]$$
(B.45)

and E is the auxiliary parameter $E_j = \frac{\Delta t(w_j + w_{j+1})}{4h_j}$. We obtain the finite difference equations for the vertical advection as follows :

$$-\check{a}_{j} \begin{pmatrix} T\\ S \end{pmatrix}_{j-1}^{(1)} + \check{b}_{j} \begin{pmatrix} T\\ S \end{pmatrix}_{j}^{(1)} - \check{c}_{j} \begin{pmatrix} T\\ S \end{pmatrix}_{j+1}^{(1)} = \begin{pmatrix} \check{d}^{T}\\ \check{d}^{S} \end{pmatrix}^{(1)}, \quad \text{for} \quad j = 1, 2, ..., J \quad (B.46)$$

$$\begin{split} \check{a}_1 &= \check{c}_j = 0, \\ \check{a}_j &= -E_j h_j / (h_{j-1} + h_j), \quad \text{for} \quad j = 2, 3, ..., J \\ \check{c}_j &= E_j h_j / (h_j + h_{j+1}), \quad \text{for} \quad j = 1, 2, ..., J - 1 \\ \check{b}_1 &= 1 + E_1 - E_1 h_2 / (h_1 + h_2), \\ \check{b}_j &= 1 - (\check{a}_j h_{j-1} + \check{c} h_{j+1}) / h_j, \quad \text{for} \quad j = 2, 3, ..., J - 1 \\ \check{b}_J &= 1 + E_J h_{J-1} / (h_{J-1} + h_J) - E_J, \\ \check{d}_j^T &= \check{a}_j T_{j-1} + (2 - \check{b}_j) T_j + \check{c}_j T_{j+1}, \quad \text{for} \quad j = 1, 2, ..., J - 1 \\ \check{d}_j^S &= \check{a}_j S_{j-1} + (2 - \check{b}_j) S_j + \check{c}_j S_{j+1}, \quad \text{for} \quad j = 1, 2, ..., J - 1 \end{split}$$

Finally for the temperature and salinity equations we obtain combination of equation B.44 and B.46 and they are calculated with the tridiagonal solving algorithm.

$$-\tilde{a}_{j} \begin{pmatrix} T \\ S \end{pmatrix}_{j-1}^{(1)} + \tilde{b}_{j} \begin{pmatrix} T \\ S \end{pmatrix}_{j}^{(1)} - \tilde{c}_{j} \begin{pmatrix} T \\ S \end{pmatrix}_{j+1}^{(1)} = \begin{pmatrix} d^{T} \\ d^{S} \end{pmatrix}_{j}, \quad \text{for} \quad j = 1, 2, \dots, J \quad (B.47)$$

with
$$\tilde{a}_1 = \tilde{c}_j = 0$$
,
 $\tilde{a}_j = \hat{a}_j + \check{a}_j$, for $j = 2, 3, ..., J$
 $\tilde{c}_j = \hat{c}_j + \check{c}_j$, for $j = 1, 2, ..., J - 1$
 $\tilde{b}_j = \hat{b}_j + \check{b}_j - 1$, for $j = 1, 2, ..., J$
 $d_j^T = \hat{d}_j^T + \check{d}_j)^T - T_j$, for $j = 1, 2, ..., J - 1$
 $d_j^S = \hat{d}_j^S + \check{d}_j)^S - S_j$, for $j = 1, 2, ..., J - 1$

B.3 Coefficient of Vertical Eddy Viscosity (A_v) and Vertical Eddy Diffusity (K_v)

The important things of turbulent mixing in the motion equations is the Richardson number, below:

$$R_{i} = \frac{g}{\rho} \frac{\partial \rho}{\partial z} \left[\left(\frac{\partial u}{\partial z} \right)^{2} + \left(\frac{\partial v}{\partial z} \right)^{2} \right]^{-1}$$
(B.48)

The coefficients of turbulent kinetic was calculated from transport equations of energy kinetic of turbulent :

$$\underbrace{\frac{\partial k}{\partial t}}_{A1} + \underbrace{u\frac{\partial k}{\partial x} + v\frac{\partial k}{\partial y} + w\frac{\partial k}{\partial z}}_{A2} = \underbrace{A_{Iv}\left[\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right]}_{A3} + \underbrace{\frac{\partial}{\partial z}A_{kv}\frac{\partial k}{\partial z}}_{A4} - \underbrace{\varepsilon}_{A5} + \underbrace{\frac{g}{\rho}A_{Mv}\frac{\partial \rho}{\partial z}}_{A6} \quad (B.49)$$

and transport equation for dissipationsrate of energy kinetic turbulent is

$$=\underbrace{c_{1\varepsilon}\frac{\varepsilon}{k}A_{Iv}\left[\left(\frac{\partial u}{\partial z}\right)^{2}+\left(\frac{\partial v}{\partial z}\right)^{2}\right]}_{B3}+\underbrace{\frac{\partial}{\partial z}A_{\varepsilon v}\frac{\partial \varepsilon}{\partial z}}_{B4}+\underbrace{c_{1\varepsilon}\frac{\varepsilon}{k}\left(1-c_{3\varepsilon}\right)\frac{g}{\rho}A_{Mv}\frac{\partial \rho}{\partial z}}_{B5}-\underbrace{c_{2\varepsilon}\frac{\varepsilon^{2}}{k}}_{B6}$$
(B.50)

where the terms in the equation B.49 and B.50 mean as follow :

A1 : Alteration rate of the kinetic energy

A2 : Advection of the kinetic energy by the middle zone of flow

A3 : Production of the kinetic energy by current shearing

A4 : Vertical diffusion of the kinetic energy

A5 : Stress dissipation of the kinetic energy

A6 : Change of the kinetic energy as a function of the stability of the layering (increase with instability, decrease with stability)

B1 : Alteration rate of the dissipation

B2 : Advection of the dissipation energy by the middle zone of flow

B3 : Production of the dissipation energy by current shearing

B4 : Vertical diffusion of the dissipation

B5 : Acceptance of the Dissipation (proportionally to the square of the dominant Dissipation and in reverse proportionally to the kinetic turbulent energy)

B6 : Change of the Dissipation as a function of the stability of the layering (increase with instability, acceptance with stability)

In addition there is the empiric equation, which it relate between the energy kinematic turbulent and its dissipation.

_

$$A_{I\nu} = c_{\nu}k^2/\epsilon, \tag{B.51}$$

where $c_v \equiv 0,08$

The results from Mellor and Yamada, 1974 (in Pohlmann, 1991) are

$$A_{Iv} \cdot \left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right] - \varepsilon + \frac{g}{\rho} A_{Mv} \frac{\partial \rho}{\partial z} = 0$$
(B.52)

and the empiric equation :

$$\varepsilon = c_{\varepsilon} k^{3/2} / L \tag{B.53}$$

We get the new vertical of changes impuls coefficien A_{Iv} as follows :

$$A_{Iv} = Lc'_v \sqrt{k} \tag{B.54}$$

and the Prandtl number, where is calculated from the equation B.48 :

$$S_M = Ri / \left[0,725. \left(Ri + 0,186 - \sqrt{Ri^2 - 0,316.Ri + 0,0346} \right) \right]$$
(B.55)

$$A_{M\nu} = A_{I\nu}/S_M \tag{B.56}$$

Insert equations B.53, B.54, and B.56 in to B.52, and we get :

$$c'_{\nu}L.\left[\left(\frac{\partial u}{\partial z}\right)^{2} + \left(\frac{\partial v}{\partial z}\right)^{2}\right] + \frac{1}{S_{M}}c'_{\nu}L\frac{g}{\rho}\frac{\partial\rho}{\partial z} = c_{\varepsilon}\frac{k}{L}$$
(B.57)

and

$$A_{Iv} = L^2 \sqrt{c_v^{'3}/c_{\varepsilon}} \cdot \sqrt{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2 + \frac{1}{S_M} \frac{g}{\rho} \frac{\partial \rho}{\partial z}}$$
(B.58)

where

$$L = c_L h_{ML} \tag{B.59}$$

And by definition :

$$c_{ML} \equiv c_L \cdot \sqrt[4]{c_v'^3/c_\varepsilon} \tag{B.60}$$

and from the other equations above, we can get :

$$A_{Iv} = (c_{ML}.h_{ML})^2.\sqrt{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2 + \frac{1}{S_M}\frac{g}{\rho}\frac{\partial \rho}{\partial z}}$$
(B.61)

where $h_M L$ the thickness of the mixing layer and Kv is determined by :

$$Kv = Av/S_M \tag{B.62}$$



Figure B.3: Vertical grid-points of exchanged coefficien (Pohlmann, 1991)

Determination of vertical exchanged coefficien A_{Iv} base on the Richardson Number Ri and definition of interface layer in model grid-points (for example in a x-z section)

Grid-points symbols :

+ Pressure- und Density point

 \times east (U) transport component in *x*-coordinate

 \equiv Shear stress point

Ako_{Iv} vertical exchanged coefficient, which are calculated with eq. (B.61)

Amin_{Iv} minimum vertical exchanged coefficient

A_{Ih} horizontal exchanged coefficien

Appendix C Air-Sea Heat Balance

The surface heat fluxes at the air-sea interface are central to the interaction an coupling between atmosphere and ocean. The processes that determine energy transfer between the surface and atmosphere include : net surface radiation flux, surface turbulent sensible heat and latent heat fluxes, heat transfer by precipitation, and storage and transport of energy below the ocean surface (*Curry, J.A.* and *P.J.* Webster, 1999).

In this study the equation, parameterisation and coefficient of turbulent flux were derivated by *Launiainen* and *Vihma* (1990) and *Vihma* (1995), which was actually applied in the Greenland and Weddell Seas. There are not yet the other study about the atmospheric-surface interaction over the Indonesia seawaters.

In the turbulent layer of the atmospheric, the flow is governed by a balance between the pressure gradient, coriolis and friction forces. The frictionally-induced momentum flux is almost constant with height in the lowest tens meters, which is called *the surface layer*. A horizontally homogeneous surface and in semi-stationary circumtances any dimensionless characteristic of the turbulence can only depend on a dimensionless stability parameter z/L, where z is the observation height and L is the Obukhov length (*Vihma*, 1995) as:

$$L = \frac{u_* T_0 \rho c_p}{g K H (1 + 0.61 T_0 c_p Q_{LH} / Q_{SH})}$$
(C.1)

Thus, the dimensionless profile gradients for the wind speed (w), the potential temperature of the air (θ) and the specific humidity (q) have a analogous form:

$$\frac{\partial V}{\partial z}Kz/u_* = \phi_M(z/L) \tag{C.2}$$

$$\frac{\partial \theta}{\partial z} K z / \theta_* = \phi_{Q_{LH}}(z/L) \tag{C.3}$$

$$\frac{\partial q}{\partial z}Kz/q_* = \phi_{Q_{SH}}(z/L) \tag{C.4}$$

Integrating the profile gradients, we get the bulk-aerodynamic formulation for the turbulent fluxes.

$$\tau_s = -\rho C_d W_z^2 = -\rho u_*^2 \tag{C.5}$$

$$Q_{SH} = \rho c_p C_H (\theta_s - \theta_z) W_z \tag{C.6}$$

$$Q_{LH} = \rho C_L (q_s - q_z) W_z \tag{C.7}$$

where the fluxes are positive upwards. The transfer coefficients are the drag coefficient C_d , the Stanton number C_L and the Dalton number C_H , where they depend on the reference height z and are relative to the roughness lengths of momentum (z_0) , heat (z_T) , and moisture (z_q) :

$$C_d = K^2 [lnz/z_0 - \Psi_M(z/L) + \Psi_M(z_0/L)]^{-2}$$
(C.8)

$$C_L = K^2 [lnz/z_0 - \Psi_M(z/L) + \Psi_M(z_0/L)]^{-1} [lnz/z_T - \Psi_{Q_{LH}}(z/L) + \Psi_{Q_{LH}}(z_T/L)]^{-1}$$
(C.9)

$$C_H = K^2 [lnz/z_0 - \Psi_M(z/L) + \Psi_M(z_0/L)]^{-1} [lnz/z_T - \Psi_{Q_{SH}}(z/L) + \Psi_{Q_{SH}}(z_q/L)]^{-1}$$
(C.10)

where Ψ_M , $\Psi_{Q_{LH}}$, and $\Psi_{Q_{SH}}$ are integrated forms of the ϕ -functions. Integrating the profile gradients between two level (z_1 and z_2) in the atmosphere, we get the gradient forms for the fluxes. Applying of equations C.5, C.7, and C.6 the turbulent flux at the sea surface can be calculated. The Sensible Heat :

$$Q_{SH} = \rho_a c_{p,air} C_H W (T_a - T) \tag{C.11}$$

The Latent Heat :

$$Q_{LH} = \rho_a L_w C_L W(q_a - q_s)) \tag{C.12}$$

The specifiq humidity of seawater are calculated with the following formulations :

$$q_s = \frac{0.622e}{10p_a - 0.378e} \tag{C.13}$$

and e = exp(-6763.6/(T + 273.15) - 4.9283 * log(T + 273.15) + 54.23)The heat flux by longwave radiation :

$$Q_{LW} = \varepsilon \sigma T^4 - \varepsilon \sigma T_a^4 \tag{C.14}$$

where $\varepsilon = 0.97$ is the emissivity of water and ε_a is the emissivity of atmosphere, which depend on cloud cover (cn).

$$\varepsilon_a = 0.7855 * (1 + (0.2232 * cn)^2.75)$$
(C.15)



Figure C.1: Regional distribution of optical water types (Jerlov, 1968)

Table C.1: Values of parameters determined by fitting the sum of two exponentials to observations of downward irradiance

Type of Oceanic Water	γ	h _{B1}	h _{B2}
		(m)	(m)
Ι	0.58	0.35	23.0
1(upper 50m)	0.68	1.20	28.0
IA	0.62	0.60	20.0
IB	0.67	1.00	17.0
II	0.77	1.50	14.0
III	0.78	1.40	7.9

The Solar radiation is calculated from the daily heat flux and inclination of sun.

$$cos\eta = sin\delta sin\phi + cos\delta cos\phi cost$$
$$Q_{SW} = S_o cos\eta * (a + b * cos\eta)(1 - alb)$$
(C.16)

The total heat flux into the ocean at the surface is given by

$$Q_1 = -(Q_{SH} + Q_{LH} + Q_{LW}) + Q_{SW}(1 - \gamma e^{(-h_1/h_B)})$$
(C.17)

where γ defines the faction of the insolation that is not immediately absorbed at the surface but penetrates into the depth of ocean, h_B is the decay length scale for the absorption of solar radiation (*Paulson* and *Simpson*, 1977). Due to insolation the deeper layers gain heat, are defined by :

$$Q_k = Q_{SW} \left(e^{\left(-\sum_{l=1}^{k-1} h_l / h_B \right)} - e^{\left(-\sum_{l=1}^{k} h_l / h_B \right)} \right) \gamma \tag{C.18}$$

Jerlov (1968) has proposed a scheme for classifying oceanic water accordings to its clarity and the optic property of the water. He defined five types: I, IA, IB, II, and III (see figure C.1. Based on

the classification, *Paul* and *Simpson* (1977) gave the constant coefficient of γ and h_B in the Table C.1.

The equations C.17 and C.18 are the source term of temperature in the equations 4.5 (Chapter 4) at the surface and the following layers until 200m depth. The source term of salinity in the equations 4.6 (Chapter 4) is calculated by the different between evaporation and precipitation. The evaporation is calculated from the turbulent flux of water vapour and the precipitation is got from the NCEP data (Chapter 3).