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A study on low-frequency variability in current and sea level in the Lombok Strait and adjacent region

Arief, Dharma, Ph.D.
The Louisiana State University and Agricultural and Mechanical Col., 1992
A STUDY ON LOW FREQUENCY VARIABILITY
IN CURRENT AND SEA-LEVEL
IN THE LOMBOK STRAIT AND ADJACENT REGION

A Dissertation

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Doctor of Philosophy

In

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ABSTRACT

Current meter data in the mid-Lombok Strait region in 1985 show that 42% of the variance in the currents is within the intraseasonal frequency band (10-100 day\(^{-1}\)). It is about two times the variance of the tidal currents. This intraseasonal variability appeared as episodic northward flows of 20 to 80 cm/s recurring every 20 to 60 days. Our data indicate these northward flows are driven by episodes of sea-level rise south of the Lombok Strait which are controlled by large scale westerly wind events in the Southeastern Indian Ocean.

These northward flow events occur almost simultaneously with either a cyclone or a typhoon generated in south or north monsoon trough, respectively. The occurrence of these northward flows and cyclone/typhoon events follow 5 - 20 days after the appearance of an atmospheric pressure trough system over the Eastern Indian Ocean - Western Pacific region.

Sea-level data from 1984-1990 in the Eastern Indian Ocean and Western Pacific region show that the intraseasonal variability is a persistent phenomena along the southern coast of the Indonesian Archipelago. The energy of this variability is concentrated at a 40-60 day period band. The magnitude of this variability suggests that the intraseasonal variability can serve as a significant modulator of the Indonesian throughflow.

The surface currents in the Lombok Strait reach 150 cm/s in the mid Strait region and 300 cm/s in the sill region. The monthly mean flow varies from 20 cm/s northward into the Flores Sea to 60 cm/s southward into the Indian Ocean. Within the Lombok Strait, the geostrophic approximation computed from cross-strait sea-level differences explains most the sub-inertial along-strait flow less than 60 cm/s. The sea-level difference outside the entrances of the Strait suggests the flow through the Strait is a frictional balance with a Rayleigh friction
coefficient estimated $4.6 \times 10^{-5}$ sec$^{-1}$. 
I. INTRODUCTION

The Lombok Strait (Figure 1) is one of the important straits for the Indonesian throughflow phenomenon that transports Pacific Ocean water into the Indian Ocean. Throughflow introduces a warm and low salinity water mass into the Indian Ocean. The Lombok Strait is the second largest opening to the Indian Ocean after the Timor-Savu passages. While its physical scale is small compared to the scale of the Timor-Savu passages, the Strait contributes significantly in carrying Western Pacific water into the Indian Ocean.

Transport through the Lombok Strait was measured during 1985 to vary between 1.7 to 4 Sv (1 Sv = $10^6$ m$^3$/s) (Murray and Arief, 1988). Results from the Naval Ocean and Atmospheric Research Laboratory (NOARL) global ocean circulation model predicted that transport through the Lombok Strait is about 20 - 50 % of the total Indonesian throughflow depending on the season (Murray et.al., 1989). Similar results were also obtained using a two-layer reduced gravity model of the Pacific and Indian Ocean basins driven by monthly winds (Inoue and Welsh, 1991). Water mass analyses indicated the main component in the transport through the Lombok Strait was associated with the salinity maximum water mass from the West Pacific Ocean (Arief, 1988).

In the Indonesian Seas, the zone where most of this water transport occurs is within the upper 200 m layer (Wyrtki, 1961). A similar thickness for this dominant transport zone was also observed in the Lombok Strait (Murray and Arief, 1988). This zone in the Lombok Strait consisted of the surface thermal mixed layer (50-100 m thickness) and the upper thermocline layer above approximately 200 m depth (Arief, 1988). In this study, we will refer to this zone as the net surface transport zone.
consisted of the surface thermal mixed layer (50-100 m thickness) and the upper thermocline layer above approximately 200m depth (Arief, 1988). In this study, we will refer to this zone as the net surface transport zone.

The Indonesian throughflow is considered to be a fundamental component of the global ocean circulation and heat flux (Toole, 1987). Piola and Gordon (1984) show that the estimation of water mass exchange between all oceanic basins would be more reasonable if there was a significant transport through the Indonesian Seas. The actual magnitude of the transport, however, is still unknown due to the scarcity of data from the region. By using several approaches and hydrographic data from the Western Pacific and the Indian Ocean regions, the Indonesian throughflow transport is estimated at 1.7 Sv to 18 Sv (Wyrtki, 1961; Cox, 1975; Godfrey and Golding, 1981; Piola and Gordon, 1984; Godfrey and Ridgway, 1985; Fine, 1985; Fu, 1986; Gordon, 1986). Currently, the estimation has narrowed to about 8 - 12 Sv as shown by global numerical models (Murray et.al., 1989; Inoue and Welsh, 1991). Hydrographic data in the Indian Ocean provide support for such a large transport from the Western Pacific. A persistent and well defined westward intrusion of low salinity water between 5°S and 15°S, extends westward in the upper 500m from the Indonesian Seas as far as Madagascar (Wyrtki, 1961, 1971; Rochford, 1966; Sharma, 1972; Levitus, 1982).

Wyrtki (1961, 1987) proposed that the inherent sea level difference between the Western Pacific and the Indian Ocean controls the Indonesian throughflow. Dynamic height difference at the surface between the area south of Mindanao and the area south of Java relative to 1000 m depth was between 8 to 22 dynamic-cm (Wyrtki, 1987). This sea level gradient was influenced strongly by the monsoonal cycle. The NOARL global circulation model, for
example, indicated that variability in the throughflow involved complex coupling mechanisms between wind systems over the Indian and the Western Pacific Oceans (J. Kindle and H. Hurlburt, personal communications).

Up to the present, characteristics of the Indonesian throughflow have been restricted to results from numerical models driven by monthly average winds (e.g., NOARL and Inoue's models). In those models, the complexity of the Indonesian Archipelago was simplified by closing most of the shallow seas and straits. The size of the opening for small straits such as the Lombok Strait, was affected significantly by a relatively coarse grid size. Such a model domain may not generate any serious distortion to the large scale circulation, but we may expect significant alteration to circulation characteristics in a small scale region.

The Lombok Strait circulation in 1985 was characterized by strong northward flow episodes occurring every 20 to 60 days. The characteristics of variability in this frequency band in the Indonesian seas, however, remains unstudied. These low frequency variabilities are also likely to be critical to the Indonesian throughflow because of their large influence on the flow in the Lombok Strait. In contrast to the ocean, this frequency band has been an important topic in atmospheric research in the tropical region in recent years, wherein it is referred as the Intraseasonal period band (for examples Ghil and Mo, 1991; Mehta and Krishnamurti, 1988; Nakazawa, 1988; Quah, 1984; Wang and Rui, 1990). Hereafter, the term Intraseasonal is used in this study to refer to processes within the 10 to 100 day period band (0.01-0.1 cycle per day frequency band).
The objectives of this study are to determine the mechanism of intraseasonal processes in the Lombok Strait. To achieve these objectives, the following steps were taken:

1. to identify the intraseasonal characteristics of the Lombok Strait circulation and sea-level systems, and of sea-level, surface atmospheric pressure and winds in the adjacent regions.

2. to determine the relationship between sea-levels and the currents in the Lombok Strait in the intraseasonal band.

3. to determine the relationship between the Lombok Strait circulation and surface atmospheric pressure and surface winds in the Indian - Western Pacific region in intraseasonal period.

The study is mainly based on current meter data collected from the 1985 Lombok Strait Experiment, the Tropical Ocean Global Atmosphere / Joint Archive for Sea Level (TOGA/JASL) sea level data, daily weather charts from the Singapore Meteorological Agency, and the European Centre for Medium Range Weather Forecasting (ECMWF) level III wind-stress data set.

The dissertation is outlined as follow. Chapter II provides the status of oceanographic and meteorological knowledge of the Indonesian region and its adjacent regions. Chapter III presents details on the data and data processing and analysis methods used in this study. Chapter IV describes properties of observational currents and sea-level time series data collected during the 1985 Lombok Strait Experiment. Chapter V presents a discussion on the relationship between currents and sea-levels in the Lombok Strait. Chapter VI presents discussion on the relationship between along-strait currents in the Lombok Strait and sea-levels in the adjacent region. Chapter VII presents discussion on the relationship between
currents in the Lombok Strait and local and large scale surface atmospheric pressure in the Eastern Indian Ocean - Western Pacific Ocean region. Chapter VIII presents discussion on the relationship between currents in the Lombok Strait and wind field. Summary and conclusions of this study are presented in Chapter IX. Suggestions for future research are also given in this Chapter.
Figure 2. Location of Lombok Strait and the vicinity. The topographic trough along the east boundary of Sunda Shelf directs flow from the Makassar Strait southward to the northern opening of Lombok Strait. One mooring (DM-mooring) is at the topographic trough east of the Sunda shelf.
Figure 3. The Lombok Strait and the position of moored current meters and pressure gauge sites.
II. BACKGROUND KNOWLEDGE

II.A. THE INDONESIAN SEAS

II.A.1. Topographic setting

The Indonesian Archipelago consists of an island arc stretching across some 5150 km along the equator (94°E-141°E and 6°N-11°S) between the continents of Asia and Australia, the Sunda and Sahul Shelves, and the Indian and Pacific Oceans (Figure 1). The Archipelago forms a leaky barrier between the Western Pacific Ocean and the Indian Ocean, and the only warm water link between any of the earth's ocean basins.

The Indonesian Seas are commonly considered to consist of two distinct oceanographic regions, i.e. the western and eastern regions. The western region is shallow (with depth of 50-75m) and filled with low salinity water (32.0-33.0 ppt). This region consists of the South China Sea and Java Sea, and part of the Sunda shelf. The eastern region is characterized by deep basins of 1000-4000 m depth, separated by deep sills more than 1000m deep. The basins are the Sulawesi, Flores, Banda, and Maluku basins. They are filled mainly by the Western Pacific water masses (Wyrtki, 1961). The easternmost region, however, is shallow with depth less than 200m. It is part of the Sahul Shelf that separates New Guinea from the continent of Australia.

Despite its critical position between two oceanic basins, knowledge of circulation in the region is very limited. The seminal data synthesis by Wyrtki (1961) still serves as the primary reference on oceanography of this region. Actually, hydrographic data collection in Indonesian Waters began during the Challenger Expedition in 1874. Detailed and multi-disciplinary observations were performed during the Snellius Expedition in 1929-32, focusing on the eastern region of Indonesian waters. Results from the Snellius
Expedition provided the core knowledge of Pacific and Indian Ocean water mass circulation in Indonesian waters. Since the mid-1950s, regular hydrographic cruises have been conducted by Indonesian agencies and through joint research programs with other countries. The recent major oceanographic project was the Snellius II Expedition in 1985-86 which had a wide focus from coastal processes to deep water mass ventilation in the Indonesian Basin. Most of the oceanographic studies in Indonesian waters have focused on large scale water mass circulation over the seasonal time scale.

II.A.2. Circulation

Surface circulation in the Indonesian Seas strongly follows the semi-annual monsoonal variation (Visser, 1938; Wyrtki, 1961). The monsoonal wind forcing causes surface currents in most Indonesian Seas to reverse direction once a year. In the West monsoon (December - February), the surface water flows from the South China Sea to the Banda Sea pushing low salinity water eastward as far as the Flores Sea. The flow reverses to a westward direction during the East monsoon period (June - September), and the 33.0 ppt isohaline is pushed back to the western Java Sea. The 33.0 ppt isohaline oscillates in the east-west direction by about 1000 km in one monsoonal cycle.

Sub-surface circulation in the Indonesian Seas is mostly deduced from hydrographic observations. The main path of the flow between the 100 and 500m depths follows a path from the Western Pacific into the Indian Ocean through the Sulawesi Sea - the Makassar Strait - the Flores Sea - the Banda Sea, and the Timor Sea. The flow is assumed to be occurring all year long. However, no long term current observation has been done in the Makassar Strait or any other possible inflow ports to confirm that assumption. A day long Ekman Current meter observation during the Snellius 1929-30
Expedition in the Makassar Strait showed a southward net flow on the order of 25-50 cm/sec in the upper 200 m layer (Lek and Fjeldstad, 1938). A recent drifter experiment in the Western Pacific (Lukas et.al., 1991) confirms the flow into the Makassar Strait from the Mindanao Current as described by Wyrtki (1961). The effect of the monsoonal cycle on the rate of flow of the Western Pacific water masses is not well known.

II.A.3. Water masses

Influx of Western Pacific water introduces two core water masses into the upper 500m layer of Indonesian waters. The Northern Subtropical Central Water (NSCW) flows at depth of 100 - 175 m with a salinity above 34.7 ppt forming a salinity maximum. The formation region of the NSCW is in the North Subtropical Pacific between 165°E and 195°E about 25°N (Wyrtki, 1961; Reid, 1965; Tsuchiya, 1968). The other core water mass, the North Pacific Intermediate Water (NPIW), flows at a depth of 300 - 400 m with a salinity value below 34.5 ppt forming a salinity minimum. This water mass is formed at the Arctic Polar Front by the sinking of cold surface water (Wyrtki, 1961; Tsuchiya, 1968; Reid 1973; Lukas et.al., 1991). The maxima and minima in the salinity profiles almost disappear in the Banda Sea and in the Timor-Savu passage (Wyrtki, 1961).
II.B. THE SOUTH COAST OF JAVA - LOMBOK REGION

II.B.1. Topographic setting

The coastal ocean along the south coast of Java and Lombok Islands and the arc of islands to the east is part of the Indian Ocean. In this region the 100 m depth contour is generally less than 20 km offshore. The 1000 m isobath is located commonly less than 50 km from the coast.

II.B.2. Circulation

In the Indian Ocean region, south of the Java - Lombok Islands, the circulation is usually characterized by monsoonal driven coastal currents. During the peak of the west monsoon in December - January, the eastward flowing South Equatorial Counter Current extends from the west coast of Sumatera to the south of the Java - Lombok coast. Along the Java coast this current is known as the Java Coastal Current (Wyrtki, 1961). Rainfall during the west monsoon period increases the cross-shore salinity gradient. The salinity difference between coastal waters and Indian Ocean water may reach more than 2 ppt which in turn may enhance the Java Coastal Current system (Soeriaatmadja, 1957).

Characteristics of the Java Coastal Current have also been observed from drifters trapped within this current system by Quadfasel and Cresswell, 1991. Speed of drifters in the Java Coastal Current region reached 1.5 m/s, and one drifter in the Java Coastal Current even went through the Lombok Strait into the Flores Sea. Quadfasel and Cresswell (1991) discussed the role of Kelvin waves from the equatorial Indian Ocean region on the dynamics of the coastal currents. Phase propagation of the internal Kelvin waves
that they interpreted from this data was estimated at 1.2 m/s. According to Quadfasel and Cresswell (1991), the excitation region for those Kelvin waves is the equatorial Indian Ocean. Propagation of waves from this region to the south coast of Java has been simulated in the NOARL global circulation model (J. Kindle, personal communication). The coastal currents cease during the East monsoon when southeasterly winds dominate the region. During this period the Indonesian throughflow reaches its maximum transport (Murray and Arief, 1988), and along the south Java coast region upwelling events are also observed (Wyrtki, 1962).

II.C. THE LOMBOK STRAIT

II.C.1. Topographic setting

The Lombok Strait is located at 115°37' E - 116°02' E and 8°20' S - 8°50' S between Bali and Lombok Islands (Figure 2, 3). It connects the Western Flores Sea (or the Bali Sea) and the Indian Ocean. The Strait spans a length of about 60 km with a north - south orientation. It is about 40 km wide and 1000 m deep at its north opening, but only 18 km wide and 300 m deep at the sill in the southern opening. Further southward into the Indian Ocean the depth increases rapidly, reaching the 1000 m isobath 20 km from the sill. The strait has a narrow beach with a steep slope, sometimes with a narrow coral reef shelf 1 to 5 km wide. At the southern part of the Strait, the Nusa Penida Island divides the strait into the Badung Strait (west side) and the Lombok Strait (east side). The depth of the Badung Strait is mostly less than 100 m with cross-sectional area roughly 1/4 to 1/3 of the Lombok Strait at its sill region.
The Lombok Strait is the second largest opening to the Indian Ocean after the Timor-Savu passage. Its cross-section above 200 m level is less than 1/15 of that of the Timor-Savu passage. The existence of a 600 m deep topographic trough along the edge of the Sunda Shelf, extending from south of the Makassar Strait to north of the Lombok Strait (Figure 2), apparently acts to increase the contribution of the Lombok Strait to the Indonesian throughflow. It effectively funnels the salinity maximum water mass from the Makassar Strait directly to the Lombok Strait. Furthermore, its position at the edge of the Sunda Shelf also provides the capability for the Strait to transport both Java Sea and Flores Sea surface waters to the Indian Ocean (Arief, 1988).

II.C.2. Oceanographic measurements:

Prior to 1985 only limited hydrographic data existed from the Lombok Strait, such as the hydrographic and tidal current observations obtained in 1942-43 (Japan Hydrographic Office, 1968). In 1985, Coastal Studies Institute, Louisiana State University, conducted an experiment in the Lombok Strait to study water mass transport in connection with the Indonesian throughflow phenomenon (Murray and Arief, 1988). The experiment produced high resolution time-series records of currents, sea level, and water temperature for nine months. In addition, nearly 200 STD profiles were obtained. Preliminary analyses and transport estimation from the experiment have been presented and published elsewhere (Arief, 1988; Arief and Murray, 1988; Murray and Arief, 1985a,b, 1986a,b, 1988; Murray et.al., 1989, 1990).
II.C.3. Water masses

The presence of the NSCW and NPIW water masses from the Western Pacific were well observed in the Lombok Strait (Japan Hydrographic Office, 1968, Arief, 1988). The salinity maximum value, centered at 150 - 175 m depth, varies seasonally according to the strength of the mean currents. It was above 34.7 ppt in September 1985 and below 34.65 ppt in June 1985. The salinity minimum, located at 300 - 350 m depth, in contrast, is practically constant at 34.50 - 34.52 ppt (Arief, 1988).

Thermohaline variation of more local origin is also seen in the Strait. A warm (>29°C) and low salinity (<32 ppt) water plume was well defined along the northern coast of Bali and the western Lombok Strait, especially during the west monsoon period. In September 1985, rather cool surface water, about 27°C, as thick as 50 m dominated the Lombok Strait. The cool water is expected to have come from an upwelling region in the Banda Sea (Arief, 1988).

II.C.4. Circulation:

The region is well known for its strong current system, potentially hazardous to navigation. Instantaneous currents are exceptionally strong and speeds of 400 cm/s have been reported in the sill region. Tidal currents in the strait flow northward during flood and southward during ebb (U.S. Naval Hydrographic Office, 1962).

Water mass composition in the Lombok Strait shows the importance of its geographical setting (Arief, 1988). The strait is transporting surface layer water from both the Java Sea and the Flores Sea but the total transport is dominated by the salinity maximum
water mass (NSCW) (Arief, 1988). The water mass transport to the Indian Ocean mainly occurs in the upper 200 m layer. The maximum transport is between July and September, during the East monsoon period (Murray and Arief, 1988,1989).

II.D. REGIONAL METEOROLOGICAL CONDITIONS

II.D.1. Seasonal Period

It is of paramount importance that the Indonesian Archipelago is located in the Indo-Pacific monsoon region. Movement of the monsoon pressure trough follows the sun’s position and generates a semi-annual monsoonal cycle. The monsoon trough is a region of low pressure, high wind shear stress and strong convection. It stretches in a nearly east-west direction from Southeast Asia to the Western Pacific and across the Indian Ocean to Australia (Cheang, 1987). The monsoon trough is a component of the atmospheric system of the Intertropical Convergence Zone or ITCZ. From December to February the monsoon trough is located in the southern hemisphere. During this period northwesterly winds develop over Southeast Asia while the northern Indonesian region is under the influence of northern hemisphere trade winds. From south of Java to the Lesser Sunda Islands region, westerly winds prevail. This period is known as the West monsoon in Indonesia, and is also the cyclone season for the Indian Ocean region.

From June to September the monsoon trough is in the northern hemisphere. The wind field is almost the opposite of that during the West monsoon. The southern hemisphere trade winds that blow from the Australian continent extend northward and strongly influence the wind field over Indonesia. Easterly winds dominate the region from the Banda Sea to the Java Sea. When approaching the equator, wind directions
turn clockwise, becoming northward. North of the equator, southwesterly winds generally blow at 10 m/s. This period is known as the East monsoon in Indonesia. It is also the typhoon season for the Western Pacific region.

In the Java-Banda Seas region the wind speed may reach up to 20 m/s during the monsoonal onset generally lasting 1-3 days. However in general, the winds are rarely above 5 m/s (ASEAN Sub-committee on Climatology, 1982). During monsoon transition periods, winds are weak and variable.

The rainy season for the western Indonesia region comes during the West monsoon period. The precipitation rate has large year-to-year variations but is mostly within the range of 3 to 5 m/year. The rainy season in the eastern region is about the opposite of that in the western region. Precipitation rate is about 1 to 2 m/year (ASEAN Sub-committee on Climatology, 1982).

II.D.2. Intraseasonal period

Within the intraseasonal band, long period planetary events are important to the monsoon climate. These intraseasonal processes seem very complex and involve many controlling conditions both from the atmosphere and ocean. The processes on this time scale are dominated by the west to east propagation of atmospheric perturbations that move at speed of 5 - 6 m/s (Wang and Rui, 1990). These perturbations are strongest during the northern hemisphere winter.

One particular event on the intraseasonal time scale that has attracted a lot of attention recently is the 30-50 day oscillations. This phenomenon, first reported by Madden and Julian (1971, 1972), was later confirmed by many researchers (for examples: Quah,
1984; Murakami et al., 1984; Lau and Chan, 1985; Mehta and Krishnamurti, 1988). These 30-50 day oscillations in the atmosphere are a global process. They were associated with the movement of convection systems within the Walker circulation by Madden and Julian (1972). This phenomenon has also been observed in the extra-tropical regions (Liebmann and Hartmann, 1984), where it was concluded that the disturbances propagated from the mid-latitude region to the Tropics.

In the ocean, 30-50 day oscillations have also been reported by Mysak and Mertz (1984) in the Somali Current, and by McPhaden (1982) in the Central Indian Ocean. The oscillations have also been detected in the sea level records along the west coast of the Americas by Spillane et al. (1987). Kawamura (1988) found 30-60 day variability in the sea-surface temperature in the tropical Western Pacific, coupled with tropical convective activity.

II.D.3. Cyclones

Short period synoptic-scale atmospheric disturbances on the order of days, such as cyclones, are also major features in the monsoonal climate. Many cyclones are generated in the monsoon trough when it is located far enough south or north so that earth rotation becomes important in the dynamics. Furthermore, the wind regime over the Indonesian region consists of three sub-systems; the monsoon westerlies and the north and south trade wind easterlies. The monsoon trough actually represents the frontal zone between the westerlies and easterlies. Imbalance in solar heating due to the sun's position causes one trough line to become more active than the other (Eguchi, 1983).
During the West monsoon, the active monsoon trough is located in the Indian Ocean - Australia region. During the East monsoon, it is over the Western Pacific - South China Sea region.

In 1985 9 cyclones and 13 typhoons were reported on the Singapore Weather Charts in the Southeastern Indian Ocean and the Western Pacific regions, respectively. While cyclone and typhoon both are dynamically the same we are going to use both terms to maintain their geographical association.

Murakami et al. (1984) showed the existence of strong coupling between synoptic-scale disturbances (cyclones/typhoons) and planetary oscillations (intraseasonal) in the region from the Arabian peninsula to the Western Pacific region. They found that several cyclones/typhoons existed contiguously in a short period. This system of cyclones/typhoons propagated eastward while each cyclone/typhoon in the system moved westward. Their finding was important because our present study showed a similar relationship between large scale intraseasonal perturbations in the atmosphere, cyclones/typhoons, currents in the Lombok Strait, and sea-level along the south coast of the Indonesian Archipelago.

II.D.4. Westerly Wind Burst

Another energetic atmospheric disturbance in the Western Pacific region is known as the Westerly Wind Burst or the Cloud Cluster Phenomena (Keen, 1987) which has recently attracted much interest in Western Pacific studies. This atmospheric process is recognized as an important forcing mechanism for upper ocean dynamics in the Western Pacific and is linked to the generation of El-Nino (Lukas, 1987). The influence of the Westerly Wind Burst on circulation in Indonesian region is still unstudied. In addition,
in 1985, when time series records of currents are available from the Lombok Strait, the Westerly Wind Burst events were infrequent and weak (Keen, 1987). Westerly Wind Bursts occur commonly between November and April with 2 to 39 days duration within the area from 5°S to 65°N. Typical Westerly Bursts, however, do not last long enough for associated wind-driven currents to reach their equilibrium states (Lukas, 1987).

II.E. CHAPTER SUMMARY

The oceanographic knowledge of Indonesian Seas is limited to hydrographic data. Lombok Strait connecting the Flores Sea and the Indian Ocean, has been revealed to be a critical component of the Indonesian throughflow system. Despite it's relatively small size, understanding the processes controlling the flow through the Strait will be of great importance in understanding the mechanism of the entire Indonesian throughflow. The dynamical characteristics of the Strait will determine the interaction between the two basins.

Coupling between the atmospheric processes and the ocean dynamics are well observed in many regions especially in the Western tropical Pacific region. The details of air-sea interaction, however, are still not well understood because of their complexity. Observations in the atmosphere persistently show the presence of energetic intraseasonal processes in the tropic region. Although the mechanics of these processes is still under investigation, many of their characteristics have been revealed. The knowledge of intraseasonal processes in the ocean, however, is still very limited. Madden and Julian, 1972, suggested that the 40-60 day period phenomenon is a world wide process in the tropics. Observations of the 40-60 day period oscillation in the Indian Ocean and Pacific Ocean basins confirmed that intraseasonal processes are of large scale. Thus intraseasonal processes are expected to be significant in the Indonesian throughflow region.
III. MATERIALS AND METHODS

III.A. DATA

III.A.1. Current

Current meter data used in this study are from the Lombok Strait Experiment in 1985, conducted by the Coastal Studies Institute, Louisiana State University in cooperation with the Research and Development Centre for Oceanology, Indonesia. The currents were measured by using ENDECO 1174 and ENDECO 174 current meters with a typical sampling rate of 8 or 10 minutes. The model 1174 is a deep water rotor version of the model 174. Some current meters also had temperature and pressure sensors. Temperature and pressure were measured with sampling rates of 8 or 16 minutes. The moored current meters were deployed in three flow regions; the upstream, mid Strait and downstream regions of the Lombok Strait (Figure 2 and 3).

The upstream region mooring (DM-mooring) was located near Kalukalukuang Bank, 500 km north of the Lombok Strait at 600m water depth. The mid Strait location was 1000m deep and instrumented by two current meter moorings (AW and AE moorings). The down-stream region was at the Lombok Strait sill region where two moorings (CE and CW moorings) were installed. The water depth was 225m at the first deployment site and 450m at the second deployment site at the CE and CW sites.

More information concerning the current meter records is given in Table 1. The last two columns are the angle of the rotation of the major axis relative to north (clockwise positive) and the percentage of total kinetic energy associated with currents along the major axis, respectively. Definition and computation of major axis and rotation angle are...
given in III.B.2. For notation simplification the current meter records will be named by their mooring names (DM, AE, AW, CE, or CW) followed by their deployment depth. For example, the current meter from AE mooring at depth 25m is referred as AE25.

<table>
<thead>
<tr>
<th>Site</th>
<th>Water Depth m</th>
<th>Mooring</th>
<th>Nominal Depth (m)</th>
<th>Data Period (0=Jan 1,85) (days)</th>
<th>Rotation degrees</th>
<th>Variance %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mid-strait</td>
<td>1000</td>
<td>AE</td>
<td>25</td>
<td>11 - 295</td>
<td>16.35</td>
<td>93.2</td>
</tr>
<tr>
<td></td>
<td>1000</td>
<td></td>
<td>50</td>
<td>11 - 139</td>
<td>20.11</td>
<td>90.9</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>300</td>
<td>11 - 139</td>
<td>355.89</td>
<td>85.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>800</td>
<td>11 - 255</td>
<td>12.37</td>
<td>89.2</td>
</tr>
<tr>
<td></td>
<td>1000</td>
<td>AW</td>
<td>25</td>
<td>154 - 240</td>
<td>41.25</td>
<td>89.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>300</td>
<td>154 - 255</td>
<td>29.26</td>
<td>89.7</td>
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<tr>
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<td></td>
<td>800</td>
<td>11 - 255</td>
<td>34.96</td>
<td>92.9</td>
</tr>
<tr>
<td>Lombok Sill</td>
<td>450</td>
<td>CW</td>
<td>25</td>
<td>151 - 220</td>
<td>10.07</td>
<td>85.6</td>
</tr>
<tr>
<td></td>
<td>300</td>
<td></td>
<td>50</td>
<td>5 - 50</td>
<td>354.90</td>
<td>95.4</td>
</tr>
<tr>
<td></td>
<td>450</td>
<td></td>
<td>150</td>
<td>5 - 141</td>
<td>340.8</td>
<td>95.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>185</td>
<td>151 - 246</td>
<td>8.22</td>
<td>90.6</td>
</tr>
<tr>
<td>Kalukalukuang Bank</td>
<td>600</td>
<td>DM</td>
<td>25</td>
<td>160 - 274</td>
<td>23.86</td>
<td>74.7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>100</td>
<td>160 - 282</td>
<td>22.75</td>
<td>81.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>300</td>
<td>160 - 274</td>
<td>36.78</td>
<td>91.1</td>
</tr>
</tbody>
</table>

Note: Deployment I started in January 1985 (before day 11)
Deployment II started in June 1985 (day 151 and later)
III.A.2. Sea Level / Pressure Gauge

Two kinds of sea level data are used in this study, sub-surface pressure gauge records and daily mean sea level. The pressure gauge records are from the Lombok Strait Experiment in 1985 where they were deployed in 10 - 15 m water depths. Original sampling rates were 1.725 and 3.5 minutes. The pressure gauge data are converted to sea-level height by water pressure to water column height conversion for water density of 1022 kg/m^-3. The daily mean sea level data are from the TOGA/JASL sea level data archive, from Dr.K.Wyrtki, University of Hawaii, and from Dr.G.W.Lennon, Flinders University of Australia. All records are demeaned. No inverse barometric correction is performed on the sea level / pressure gauge data records because of the large scale character of the atmospheric pressure in the intraseasonal band. Table 2 summarizes the sea level records.

III.A.3. Wind and Atmospheric Pressure

Meteorological data used in this study consist of sea level barometric pressure and wind data from several sources. The daily barometric pressure data at Bali, Lombok, Bawean, Kotabaru, and daily wind data at Bali and Lombok are from the Indonesian Meteorological and Geophysical Agency. The barometric pressure data at other locations are read from 6-hourly weather charts of the Singapore Meteorological Agency. Wind-stress data in the region 90°E-150°E and 30°S-30°N are the ECMWF/TOGA Supplementary Fields Data Set from the European Centre for Medium Range Weather Forecasts (ECMWF) / Tropical Ocean Atmospheric (TOGA) Level III Atmospheric Data
Centre. The Supplementary Fields Data Set are derived from 6-hour forecasts at 00:00 UTC and 12:00 UTC (ECMWF, 1990). This study, however, uses only the data at 00:00 UTC.

<table>
<thead>
<tr>
<th>Site</th>
<th>Region</th>
<th>Period of data in days</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jakarta</td>
<td>Java Sea</td>
<td>-366 to 211</td>
<td>2</td>
</tr>
<tr>
<td>Surabaya</td>
<td>Java Sea</td>
<td>1120 to 1928</td>
<td>3</td>
</tr>
<tr>
<td>Cilacap</td>
<td>Indian Ocean</td>
<td>-281 to 149</td>
<td>2</td>
</tr>
<tr>
<td>Padang</td>
<td>Indian Ocean</td>
<td>690 to 1937</td>
<td>3</td>
</tr>
<tr>
<td>Phuket</td>
<td>Indian Ocean</td>
<td>0 to 924</td>
<td>3</td>
</tr>
<tr>
<td>Cocos Isl.</td>
<td>Indian Ocean</td>
<td>348 to 1319</td>
<td>3</td>
</tr>
<tr>
<td>Darwin</td>
<td>Australia</td>
<td>48 to 729</td>
<td>4</td>
</tr>
<tr>
<td>Port Hedland</td>
<td>Australia</td>
<td>-53 to 364</td>
<td>3</td>
</tr>
<tr>
<td>Davao</td>
<td>Sulawesi Sea</td>
<td>48 to 908</td>
<td>2,3</td>
</tr>
<tr>
<td>Jolo</td>
<td>Sulawesi Sea</td>
<td>-365 to 785</td>
<td>2</td>
</tr>
<tr>
<td>Bitung</td>
<td>Maluku Sea</td>
<td>365 to 1469</td>
<td>3</td>
</tr>
<tr>
<td>Batu Tiga</td>
<td>Lombok St.</td>
<td>156 to 265</td>
<td>1</td>
</tr>
<tr>
<td>Nusa Penida</td>
<td>Lombok St.</td>
<td>11 to 265</td>
<td>1</td>
</tr>
<tr>
<td>Nipa Bay</td>
<td>Lombok St.</td>
<td>157 to 256</td>
<td>1</td>
</tr>
<tr>
<td>Benoa</td>
<td>Lombok St.</td>
<td>1215 to 1896</td>
<td>3</td>
</tr>
</tbody>
</table>

1. Lombok Strait Experiment 1985
2. Courtesy of Dr. K. Wyrtki, University of Hawaii
3. NODC - TOGA/JASL data set
4. Courtesy of Dr. G.W. Lennon, Flinders University, Australia
III.B. METHODS

III.B.1. Interpolation for missing data

Time series data from the Lombok Strait Experiment in 1985 have data gaps between deployment periods. To construct an uninterrupted time series, a linear interpolation scheme is used following

\[
\tilde{y}(t) = \frac{y(t-3\Delta t) + y(t-2\Delta t) + y(t-\Delta t) + y(t+\Delta t) + y(t+2\Delta t) + y(t+3\Delta t)}{6}
\]

with \( t \) and \( \Delta t \) in days. The value of \( \Delta t \) is determined by trial and error until a smooth transition between the two pieces of data is achieved. The effect of interpolation is explored by comparing the actual data and their computed interpolation values according to the above scheme, for 5 to 10 day data blocks. The result shows that the interpolation scheme can restore more than 90% of the actual variance. For the Lombok data set, the best interpolation result was achieved using \( \Delta t \) close to the neap-spring cycle (14 days). For the Nusa Penida pressure data (see Figure 3 for the location), for example, the \( \Delta t \) is 14.035 days (Figure 4A). Furthermore, the data gap between the first and second deployment is about 5% of the total length of the data set. The effect of this interpolation on the whole data set is negligible.

For 6-hourly or daily data such as the barometric pressure, wind, and sea level data, the missing observations were usually scattered along the length of data but limited to a few points. Those missing data were interpolated by using a cubic spline method (Press et al., 1989). In a few cases, for a certain configuration of data, the cubic spline
produced an outlier peak. For those cases, linear interpolation was applied only to the necessary data gap. In cases where the missing data occurred several days consecu­
tively, the data set blocks were treated as individual sets.

Daily sea level data from tide station network often had data gaps for several days. In many cases, the data was truncated to exclude the gap. However, for some locations such as Cilacap (Figure 1), the available data was limited, and all data was essential.
Linear interpolation was performed to fill the gap, so filtering by the FFT method could be carried out. Despite the interpolation, the bandpass filtering procedure still produced a good data profile without much adverse effect by the interpolation block as shown in Figure 4B.
III.B.2. Coordinate rotation

The current in the Lombok Strait is almost a bi-directional flow. Rotating the magnetic coordinate system to a new orthogonal coordinate system which maximizes and minimizes variance of currents along the orthogonal axes often makes the interpretation clearer. The rotation angle of new coordinates relative to magnetic coordinates, $\alpha$, is calculated by

$$\alpha = \frac{1}{2} \tan^{-1}\left( \frac{2 \sum_{i=1}^{n} X_i Y_i}{\left| \sum_{i=1}^{n} X_i^2 - \sum_{i=1}^{n} Y_i^2 \right|} \right)$$

where $X_i$ and $Y_i$ are components of the vector $V_i$ in magnetic coordinate (Preisendorfer, 1988). The rotated coordinates generally follow the topography of the Strait at the current measurement sites. For the AE25 data set the rotation angle is 16.5° clockwise. This is close to the actual topographic orientation of the Lombok Strait. For the other sites, the rotation angle was between -25° and 45° from north. Hereafter, the current components in the rotated coordinate system are referred to as along-strait and cross-strait currents.

III.B.3. Data Filtering

To enhance identification of intraseasonal variability in the data sets, 4-day lowpass and 10 - 100 day bandpass filtering were performed on all data by using the Fast Fourier Transformation (FFT) technique with a box-car window in the frequency domain. All FFT coefficients outside the window area were set to zero and the resulting sequence inverse transformed. Forbes (1988) cautioned that use of the box-car window created 'ringing',
and suggested using other window schemes. For the present data sets, however, comparisons of results using several window schemes did not show significant differences in the output.

In the analysis, the same sampling interval is often needed for all parameters. This demand is fulfilled by decimation of the data set. Decimation of the data set is performed by first lowpass filtering the data sets and then resampling the lowpass filtered data set at a sampling rate half of the cutoff period using the cubic spline method.

III.B.4. Time domain analysis

Data analysis in the time domain is performed by using graphic analysis and statistical analysis. Graphic analysis uses time series plots for visual detection of any trend or episodic events in the data sets. Statistical analysis is used to determine data sets' inter-dependency. It is performed by using Pearson's correlation and regression methods (Press et.al., 1989).
III.B.5. Spectral Analysis

Spectral characteristics of time series data are studied by computing power spectra of time series data based on the Fast Fourier Transform (FFT) technique. The power spectral density function or auto-spectral density \( G_{xx}(f) \) for a stationary record is computed following the procedure explained by Bendat and Piersol (1986).

\[
S_{xx}(f_k) = \frac{1}{N \Delta t} |X(f_k)|^2
\]

\[
X_{xx}(f_k) = \Delta t \sum_{i=0}^{N-1} x_i e^{-2\pi ifk} ; \quad i = \sqrt{-1} \quad k = 0, 1, \ldots, N - 1
\]

\[
G_{xx}(f_k) = 2S_{xx}(f_k) ; \quad k = 0, 1, \ldots, \frac{N}{2} \quad [3a]
\]

where \( f_k = \frac{k}{N \Delta t} \) with \( N \) number of data, and \( \Delta t \) sampling interval. The cross-spectral density \( G_{xy}(f) \) is computed by

\[
S_{xy}(f_k) = \frac{1}{N \Delta t} |X(f_k)Y(f_k)|
\]

\[
G_{xy}(f_k) = 2S_{xy}(f_k) \quad k = 0, 1, \ldots, \frac{N}{2} \quad [3b]
\]

The coherence function of two stationary quantities \( x(t) \) and \( y(t) \), \( \gamma_{xy}^2(f) \), is computed from the auto-spectral and cross-spectral density functions

\[
\gamma^2(f) = \frac{|G_{xy}(f)|^2}{G_{xx}(f)G_{yy}(f)} \quad [4]
\]

The coherence function may be interpreted as a measure of the linear predictability of one data set from the other.
Power spectra are computed on time series data with Hanning windowing performed to reduce sidelobe effects. Further, to statistically stabilize the power spectra estimation, 5 point averaging in the frequency domain was used to estimate the power spectra (Press et al., 1989). For the present analysis this procedure is preferred over the block average method (Bendat and Piersol, 1987) or the Weighted Overlapped-Segment Averaging (Carter, 1987). For long data sets, all methods produced equivalent power and coherence estimations. For limited length time series, however, the information on low frequency signals is eliminated by block averaging.

Assumptions of weak random and normal distribution for time series data sets discussed in this study are validated by Reverse Arrangement and Chi-square goodness-of-fit tests (Bendat and Piersol, 1986). These statistical tests show that the data sets fulfill the hypotheses of weak random stationarity and normal distribution at the 5% significance level.
IV. DESCRIPTION OF PHYSICAL PROCESSES IN THE LOMBOK STRAIT

The research emphasis of this study is on understanding the intraseasonal processes in the Lombok Strait and surrounding region. However, because knowledge of the circulation in the Lombok Strait is extremely limited it is important to first present a description of the present data sets. The characteristics of currents in the Lombok Strait are presented here. Description of tides, and the characteristics of currents at the up-stream mooring site (DM) are given in Appendix B.

IV.A. CURRENTS IN THE LOMBOK STRAIT

IV.A.1. Directional and Vertical Properties

Circulation in the Lombok Strait is a bi-directional flow in a northeast - southwest direction as shown by current rose histograms (Figure 5). The current roses show the frequency distribution of current directions in 10-degree bins. These dominant flow directions are the same as indicated by the principal axes given in Table 1. The current roses in the Lombok Strait are typical for tidal currents confined by lateral boundaries and influenced by strong southward low frequency flow. The asymmetry of the current directions in the upper 300 m depth indicates strong domination by low-frequency currents in the circulation system. In contrast, at depths below 800 m tidal currents dominate the circulation, but flow tendency is still in the southward direction as at the other depths.

Lombok Strait has a strong current system, and the currents have significant spatial and temporal variabilities. In the mid-Strait region (AE and AW mooring locations, Figure 3), the currents in the upper 100 m reached 150 cm/s. The depth of water in this region is about 1000m.
ROSE HISTOGRAM OF CURRENTS IN THE LOMBOK STRAIT

Figure 5. Current rose histograms showing the frequency of current directions in 10-degree bins of hourly currents at the Lombok Strait in January-September 1985. The label on each histogram identifies the current meter. The length of arrow represents a scale for the number of data points in each directional bin. The arrow points to north.
Vertical characteristics of currents in the Lombok Strait are shown in Figure 6. Composite scatter plots from the data of AE and AW moorings display north-south and east-west current components as a function of depth (measured by a pressure sensor in the current meter). The data cover the period from January to September 1985.

The westward components of currents are obviously stronger than the eastward components. In the north-south direction, maximum speed occurs in the north direction. Those high northward currents over 100 cm/s are caused by a single northward flow event on February 19-21, 1985. The currents at depths between 300-350 m are about half of the current magnitude at the 25-100 m with the flow tendency in the south-west direction into the Indian Ocean. At 800-900 m, the magnitude of currents is higher than that at 300-350 m. Note the distinct pattern in Figure 6 (labeled as AW800) at 800-900 m. These data points are from the AW800 in January-June deployment period. The AW-mooring at that time broke off at about 700 m, and left the mooring with insufficient buoyancy to keep it vertical despite weak currents. The vertical distribution of current speeds strongly suggest a baroclinic flow above 200 m with an amplitude of ~75 cm/s superimposed on a barotropic tidal component of amplitude of ~50 cm/s.

The vertical salinity profile in Figure 6 is drawn based on an average of salinity profiles close to the AE- and AW-moorings measured down to 500 m in January, June, and September 1985. It is shown that the strongest currents exist in the low salinity layer above the salinity maximum. The currents at 300 m represent the flow of salinity minimum layer (NPIW).

Current speeds increase significantly toward the sill region as shown in the scatter plot in Figure 7. Figure 7 is similar to Figure 6 except data from the sill region are shown.
The plot consists of the current data from the CW and CE moorings (Figure 3). During the first deployment, from January to May 1985, the moorings were deployed in water about 300m deep. During the second deployment, however, the moorings were deployed in water of about 450m depth. Current speeds in the sill region were about twice the currents speeds in the mid Strait region. In the upper 75 dbar, the north-south components of currents reached 300 cm/s and the west-east components reached 100 cm/s. As shown in Figure 7, the current speeds in the north-south direction decreased considerably with depth. At the 100 - 250 dbar level, the maximum current was 150 cm/s. The east-west current speeds, however, were nearly depth invariant.
The distribution of the current speeds at each depth were almost symmetrical about the zero line. No obvious flow tendency appeared in the sill region from the scatter plot because of domination by tidal currents. However, Table 3 shows the monthly mean flow consistently southward to the Indian Ocean. Note that the spreading of data vertically below the nominal deployment depth (e.g. at speeds over 100 cm/s of the north-south components CW25, Figure 7) is caused by the deflection of the mooring due to large current drag forces.

Figure 7. Similar to Figure 6 but for sill locations. Note at the first deployment (CE50 and CE185) the mooring was deployed at a water depth of 300 m. At the second deployment, the moorings were deployed at water depths of 450 m. Salinity profile is the average of STD-casts in the sill region from January, June, and September 1985.
The salinity profile shown in Figure 7 is the average of CTD-casts from January, June, and September 1985 in the sill region. The salinity maximum layer was observed in some individual profiles, but on average the salinity maximum layer had already dissipated from intense mixing on the sill. Similar to the mid Strait condition, the strongest flow is within the low salinity layer.

IV.A.2. Monthly Currents

Table 3 shows monthly means of east-west currents (upper value) and north-south currents (lower value) of all current records shown in Table 1. Empty values indicate no observation. The monthly mean of the north-south currents above 50m depth flow southward into the Indian Ocean. At the mid-Strait location, the monthly mean currents varied between 5 cm/s in March 1985 (AE25) and 71 cm/s in July 1985 (AW25). The strongest monthly mean current was at CW25 in July 1985 at 87 cm/s. February to April 1985 was the period of weak monthly mean current in the Lombok Strait (as shown by AE25) due to three strong northward events which occurred in these months. The currents increased abruptly from June to July 1985, and strong currents were maintained until October 1985. July to October 1985 was the period of the East monsoon in the Indonesian region when southeasterly winds prevailed. During this period the direction of the wind was the opposite of the currents in the Lombok Strait.

At the 300m depth the monthly means of north-south currents were dominantly southward. The mean values however, were much weaker compared to the currents above 50m. Maximum monthly mean current was 13 cm/s southward at AW300 in August 1985. Weak monthly mean northward currents occurred at AE300 in May, June,
and September 1985 at less than 6 cm/s. Despite the existence of a 300 m sill at the southern opening, southward monthly mean currents were still observed at 800m depth with values less than 6 cm/s (Table 3).

The monthly means of east-west currents in the Lombok Strait were generally weak, below 10 cm/s, Table 3. Exceptionally strong westward monthly mean currents occurred in July and August 1985 at AW25. In these months, the monthly mean currents reached 48 cm/s and 77 cm/s, respectively. This strong westward currents were apparently due to the increase of flow into the Badung Strait. This increase of westward currents during the strong southward flow period was also noted at AE25, but the monthly mean values were still below 20 cm/s. This discrepancy between westward currents at the AW and AE locations seems to be related to AW and AE’s locations relative to the Badung Strait. The AW mooring location was close to the Badung Strait (Figure 3).
Table 3. Monthly mean of currents in the Lombok Strait and DM-mooring

Above: EAST-WEST CURRENTS in cm/sec (- is west)
Below: NORTH-SOUTH CURRENTS in cm/sec (- is south)

<table>
<thead>
<tr>
<th>CURRENTS</th>
<th>JAN.</th>
<th>FEB.</th>
<th>MAR.</th>
<th>APR.</th>
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Power spectra are shown in Figure 8A - 8D for the demeaned along-strait and cross-strait currents at AE25. The AE25 current record of 284 days is the longest current record. The mean values for these 284 days of data are 35 cm/s southward and 3 cm/s eastward for the along-strait and cross-strait currents, respectively. The variance in the currents is distributed over sub-tidal (longer than 1 day period), tidal (diurnal and sub-diurnal periods), and over-tidal or supra-tidal (shorter than 0.5 day period) bands as presented in detail in Table 4. The variance in a given frequency band, $E_{a-b}$, is computed as

$$E_{a-b} = \int_{f_a}^{f_b} G \, df$$

where $G$ is the power spectrum density defined in Eq.3.

The variance of along-strait current is 7.5 times larger than the variance of the cross-strait current. The variance within the sub-tidal and diurnal bands are much larger in the along-strait currents than in the cross-strait currents. Table 4 also shows that the variance within the 0.0 - 0.1 cpd band for along-strait currents, for example, is 62 times that in the cross-strait currents. For the diurnal band (0.8 - 1.2 cpd), the variance in the along-strait current is 4 times the variance in the cross-strait current. In contrast, the variance in semi-diurnal currents (1.8 - 2.2 cpd) is comparable in both current components. Though the variance in each over-tide current (over 2.2 cpd) is 1 to 2 orders of magnitude smaller than in the semi-diurnal current (Figure 8), the total variance within all over-tide bands is significant.
It is about half of the energy in the semi-diurnal and diurnal tides combined (Table 4). Further description of tidal currents is presented in Appendix B.

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<td>2</td>
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<tr>
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For the along-strait current, as shown in Table 4, 33% of variance is within the very low frequency band of 0.0 - 0.01 cycles per day (cpd) band, 42% within the intraseasonal band (0.01-0.1 cpd), and only 19% variance within the tidal band (higher than 0.8 cpd). The variance in the low frequency band increased monotonously without distinctive peaks as shown in Figure 8. The FFT procedure can not resolve any peaks in this wide red spectrum system due to the relatively short time series data for current. In contrast, for the cross-strait current,
only 4% of the variance is within the 0.0 - 0.01 cpd band, 13% within the intraseasonal band, and 78% of the variance within the tidal band. This analysis indicates that along-strait currents at AE25 are an excellent representative for the low frequency circulation in the Strait.
Figure 8. Power spectra of along- and cross-strait components of AE25 current. (A). Along-strait currents, 0.0-10. cpd. (B). Along-strait currents, 0.0-0.2 cpd. (C). Cross-strait currents, 0.0-10. cpd. (D). Cross-strait currents, 0.0-0.2 cpd.
IV.A.4. Sub-tidal Currents at AE25

The AE25 current is dominated by the sub-tidal fluctuation. The energy in this frequency band is mostly confined to in the 0.0 - 0.1 cpd band (Table 4) and the monsoonal flow into the Indian Ocean. The strongest monsoonal flow period is during the East monsoon, between July and August, with the speed more than 60 cm/s southward.

This southward monsoonal flow is significantly modified by the flow events occurring at 10 to 60 day intervals (Figure 9 and 10). In the mid-Strait region, this intraseasonal variability represents 42% of the variance in the current as shown in Table 4. It is close to two times the variance of the tidal currents. Figures 9 and 10 show stick plots of the 10 - 100 day bandpass currents at the mid Strait and the sill regions, respectively. These flow events occur in clusters which are about 50 days apart. Each event within a cluster exists about every 10-20 days lasting for one to two weeks. This flow event often reverses the flow in the Strait northward, but then the flow in the Strait returns to about its pre-event level. These reversal flow events are the main component of the intraseasonal variability in the Lombok Strait circulation.

In the upper 200 m depth, the intraseasonal variability at the AW-mooring is stronger than at the AE-mooring (Figure 9). The intraseasonal current at AW25 and AE25 are in-phase but the peak of the current is reached about one day ahead at AW25 compared to AE25. Below 200m depth, the intraseasonal currents are still present but weak, and in the opposite direction from the upper 200m depth currents.
Figure 9. Stick plots of 10-100 day bandpass filtered currents in the mid Strait region.
Figure 10. Stick plots of 10-100 day bandpass filtered currents in the sill region.
IV.B. SEA LEVEL IN THE LOMBOK STRAIT

IV.B.1. Sea Level Characteristics

The sub-surface pressure gauge records at Nusa Penida (close to the sill), Batutiga (the eastern side of Bali Island), and Nipa Bay (the western side of Lombok island) (Figure 3) are dominated by the tide. The tide is a mixed type and the tidal range varies by location (Figure 11). At Nusa Penida and Nipa Bay the tide range is close to 2m. The tide range at Batutiga is about 1.5m.

IV.B.2. Spectral Characteristics

Power spectra of pressure gauge records at Nusa Penida, Batutiga and Nipa Bay are shown in Figure 12a. The spectra are characterized by peaks in the sub-tidal, diurnal, semi-diurnal, and over-tide frequency bands. The O1, S1, and K1 tidal constituents are the major components of the diurnal tide. The M2 and S2 tidal constituents are the major components of the semi-diurnal tide. The third, fourth, fifth, and sixth diurnal tidal constituents are the major components of the over-tides. The higher frequency over-tide signals, however, are still well developed. The strength of the diurnal and semi-diurnal tides are comparable, and both dominate the records. Their variance is about one order of magnitude higher than the variance of the sub-tidal band (Figure 12b), and about 2 to 3 order of magnitude higher than the variance of each over-tide constituent (Figure 12a).

Figure 12b shows the spectrum in the 0.-0.2 cpd band of pressure gauge records at Nusa Penida and Batutiga. The spectra are characterized by a monotonous increase of energy toward the low frequency band with a consistent energy concentration at all 3 stations in the 0.07 - 0.08 cpd (12.5 - 15 day) band.
Figure 11. 4-day highpass filtered pressure gauge records at Nusa Penida, Batu Tiga, and Nipa Bay sites that represent tidal oscillations.

Note that a spectrum peak exists in the 0.02 cpd (50 day period) of the Benoa spectrum. This energy peak does not appear at the Batutiga and Nusa Penida spectra.
because of their limited length of records. The tidal energy in this frequency band at Nusa Penida, however, is about one order of magnitude lower than at Batutiga and Benoa.

The variance distribution of pressure gauge records over frequency bands is shown in Table 5. The semi-diurnal (1.8-2.2 cpd) and the diurnal (0.8-1.2 cpd) bands dominate the variance in these records, but their proportion varies with locations. At Nusa Penida, the variance in the semi-diurnal and diurnal bands is about the same. At Batutiga and Nipa Bay, the variance of the diurnal signal is more than 2 times the variance in the semi-diurnal band. The contribution of over-tide components to tides in the Lombok Strait is small, less than 1%.

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<th>NIPA BAY VARIANCE</th>
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<td>2.20 - 10.00</td>
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POWER SPECTRAL OF PRESSURE GAUGE RECORDS IN THE LOMBOK STRAIT

Figure 12a. Power spectra of pressure gauge records at Nusa Penida, Batu Tiga, and Nipa Bay sites.
Figure 12b. Power spectra of pressure gauge records at Nusa Penida and Batutiga in the 0.0 - 0.2 cpd frequency band.
V. CURRENT AND SEA-LEVEL SLOPE RELATION IN THE LOMBOK STRAIT

The term sub-Inertial refers to motions with period longer than the inertial flow period. The inertial flow is a free motion controlled by the Coriolis force (Pond and Pickard, 1983). The inertial period \( \omega \) is defined as

\[
\omega = \frac{2\pi}{2\Omega \sin \theta}
\]

\( \Omega \) = rotation rate of earth = \( 7.292 \times 10^{-5} \) s\(^{-1} \) and \( \theta \) is latitude in degrees. The Lombok Strait, located at 8.5°S, has an inertial period of 3.37 days or inertial frequency of 0.269 cycles per day (cpd). For simplicity a upper cut-off frequency for the sub-inertial band is taken as 0.25 cpd (4-day). All records discussed in this Chapter are 4-day lowpass filtered using an FFT technique.

Simultaneous current and pressure gauge records necessary for the following analysis are available only from June to September 1985. Consequently, the analysis in this chapter is performed for only this period. The location map labeled with terms used in the discussion is shown in Figure 13.

For simplification, location names of Nusa Penida, Batutiga, and Nipa Bay are abbreviated by P, B, and N, respectively, in the following text. Labels AW and AE represent mooring sites (Figure 13). Only current records at the 25m depth and their east-west and north-south current components are used in the analysis in this Chapter. Sea level slope directions are shown by lines labeled 1 and 2. Line 1 represents east-west cross-strait direction in the Lombok Strait (N-B) and Line 2 represents north-south cross-strait direction across the northern entrance to the Badung Strait (P-B). Note Line 1 is in an east-west direction and Line 2 is in a north-south direction. Positions of N and P are not exactly east and south of B,
respectively. In this discussion, we assume that N-B and P-B approximate the sea-level slope in the east-west and north-south directions. The coordinate convention here is that positive directions are to the east, north, and up directions.

V.A. CURRENTS - SEA-LEVEL SLOPES RELATIONSHIP

V.A.1. Sub-Inertial Currents and Sea-levels

At sub-inertial time-scales, the sea levels and currents at 25m depth in the Lombok Strait show a high visual coherency. In Figure 14 the 4-day lowpass filtered east-west and north-south currents at AE25 and AW25 are plotted together with sea-levels at Nusa Penida (P), Batutiga (B), and Nipa Bay (N). The data period is from June 9 to September 7, 1985 (day 160 to 250). The sea-level records are demeaned to remove the effect of instrument deployment depths.
Figure 14. Four-day lowpass filtered 25m depth current and sea-level records in the Lombok Strait in June - September 1985 period. Northward and eastward currents are positive.

The flow in the Lombok Strait was characterized by strong southward flow modified by northward flow events. During this June - September 1985 period, the flow consisted of two stages: a stage of northward flow (day 160 - 190) and a stage of strong southward flow (day 190 - 245). The northward flow stage has two northward flow events centered about day 160 and 178 (marked by a and b in Figure 14). Between those events there
was a strong southward flow period about day 165 (c). In the second stage of flow significant oscillation in the current was still present albeit smaller than those associated with the northern flow events.

During this 85 day period, characteristics of currents at AE25 and AW25 were significantly different as shown in Figures 20A and B. Typically the north-south currents at AW25 were stronger than those at AE25 by more than 20 cm/s. The peak northward flows of AW25 and AE25 were practically in-phase. The east-west currents at AE25 and AW25, in contrast, were significantly different. The east-west currents at AE25 were weak, less than 25 cm/s. After day 190, however, the east-west currents at AW25 were strong. They reached more than 75 cm/s westward. The currents even exceeded the speed of southward currents between day 225 and 240. In addition, during reversal flow period (day 160 - 190) the peaks of eastward flow at AW25 were reached about 1 day after the peaks of northward currents at AW25.

Sea-level records at three sites (P, B, and N) also showed variability similar to the currents. Sea-levels rose simultaneously at the three sites during northward flow events, and dropped together during strong southward flow. This finding is significant because it shows that the sub-inertial flow through the Lombok Strait is driven by the along-strait sea-level slope controlled by the Indian Ocean sea-level. This Indian Ocean controlled process is discussed further in Chapter VI. Wyrtki (1961) observed a similar relationship between along-strait slope and current at the Sunda Strait, between Java and Sumatera Islands, and at the Malacca Strait, between Sumatera Island and Malaysia Peninsula.

The variability at the three pressure gauge locations was different. Sea-level at B varied the most followed by sea-levels at N and P. The range of sea level variability at B
was 25 cm, but was about 10 cm at P. The timing of high or low sea-level phase at P, B, and N sites also varied. The sea-level condition in the Strait is strongly influenced by the currents. There is, however, a tendency that the high or low sea-level phase was reached first at P followed by N and B (Figure 14C). This tendency agrees with incoming variability of sea-level from the Indian Ocean.

Even though no current measurements were made in the Badung Strait, strong east-west currents at AW25 suggest an important role for the Badung Strait in the Lombok Strait circulation. From the strong westward current component measured at the AW-mooring (Figure 14) the flow into the Badung Strait is expected to be strong, about 75 cm/s. This current speed is comparable to the southward flow through the Lombok Strait. Different characteristics between sea-level at Nusa Penida and Batutiga also indicate that there is a significant flow into the Badung Strait.

V.A.2. Sea-level Slope in the Lombok Strait

Inside the Strait, from basic principles we expect currents to correlate better with sea-level slope than with sea-level itself. Absolute sea level difference between P and B (P-B, cross Badung Strait), and between N and B (N-B, cross Lombok Strait) cannot be directly determined from our measurements. The relative sea level differences do, however, have strong linear relationships with current speed. The sea-level difference across the north Strait N-B correlates best with the average of north-south currents at AW25 and AE25. Its Pearson’s correlation coefficient is -0.89. This relation suggests a tendency toward quasi-geostrophic adjustment inside the Strait.
Interestingly, there is even a stronger correlation between the sea-level difference N-B (cross-strait at the entrance to the Badung Strait) and the average of east-west currents at AW25 and AE25. Their Pearson's correlation coefficient is -0.96. This high correlation points out the importance of the flow into the Badung Strait in the upper layer circulation of Lombok Strait. Further in this Chapter, the average of the components of AW25 and AE25 currents are referred as the 'observed' currents. Further we take this east-west component of current to represent the flow through the Badung Strait.

These strong linear dependencies, then, are used to estimate the absolute sea-level difference by assuming that currents cease when no sea-level gradient exists. The zero-crossing coefficient of least-square linear regression is taken as the offset in relative sea-level differences. The off-set values are 7.7 cm and 6.7 cm for P-B and N-B, respectively. For data set with more zero-crossings, the estimation of this off-set value will be better.

V.A.3. North-south Currents

Figure 15A shows the temporal variability of north-south currents and N-B sea-level difference after adjustment by the off-set value. The scatter diagram of those data is shown in Figure 15B. Similar plots are also shown for the east-west currents and P-B in Figure 16. The ranges of N-B and P-B are 15cm and 20cm, respectively. Actually, the P-B and N-B sea-level differences are dominated by the large fluctuation at B. Relative sea-level difference between P and N is small compared to P-B and N-B.

As shown in Figure 15A, N-B sea-level difference is almost a mirror image of the north-south currents. Lag-regression analysis reveals that the currents lead N-B by 19
Figure 15. (A). Sea-level difference between Nipa Bay and Batutiga (N-B, cross Lombok Strait), and north-south currents. (B). Scatter plot between N-B and north-south currents. Events marked by 1, 2, 3 and a, b, c are discussed in the text.
hours. This time-lag represents the time scale for sea-level adjustment to the change in current. This value is about two times faster than the inertial time scale which is 40 hours (half of inertial period).

Sea-level difference N-B is sensitive to the change in the current. The sea-level difference N-B responds even for small current fluctuations about 10 cm/s such as shown at day 210-225 (marked by 3 in Figure 15A). During northward flow events between day 160 and 180 (marked by 1 and 2 in the Figure) the N-B becomes negative, or sea-level at B rises more than at N. When the flow returns to southward about day 165-170 (marked by a in the Figure) the N-B becomes more positive which means sea-level at B drops more than at N. During the period of strong steady southward flow that occurred after day 190, N-B stays positive. These sea-level slope settings according to current directions agree with the sense of geostrophic flow.

The strong linear relationship between north-south currents and N-B sea level difference is shown in the scatter diagram Figure 15B. The least-square parametric relation between N-B, $Y$, and north-south currents, $X$, is

$$Y \text{ (cm)} = -0.12X \text{ (cm/s)} \quad [7]$$
V.A.4. East-west Currents

A very similar relationship between the P-B sea-level difference and the east-west currents is shown in Figure 16A. Again the currents lead P-B by about 20 hours. During strong eastward flow events about day 160 and 180 (marked by 1 and 2 in the Figure), P-B becomes negative. This means the sea-level at B rises more than at P. During a period of westward flow at about day 165-170 (marked by a in the Figure), between those two eastward flow events, P-B reaches slightly more than 5 cm. The sea-level at B drops more than at P. During a period of increasing strong westward flow, assumed into the Badung Strait, up to 50 cm/s after day 190, P-B also increases, reaching up to 15 cm.

A strong linear relationship between east-west currents and P-B sea level difference is shown by the scatter diagram Figure 16B. The least-squares parametric relation between P-B, \( Y \), and east-west currents, \( X \), is

\[
Y \text{ (cm)} = -0.25 \times \frac{X}{\text{cm/s}}
\]  

[8]
Figure 16. (A). Sea-level difference between Nusa Penida and Batutiga (P-B, cross Badung Strait), and east-west currents. (B). Scatter plot between P-B and east-west currents. Events marked by 1, 2 and a, b are discussed in the text.
V.B. GEOSTROPHIC CURRENTS

The role of non-linear processes in the Lombok Strait circulation is expected to be important. To estimate the effect of these non-linear processes faces difficulty because of unknown non-linear parameters in the Strait. The magnitude of these processes, however, can be estimated indirectly through a comparison between the geostrophic current estimation and the observed currents.

In a geostrophic balance system, the temporal and spatial accelerations and the friction are neglected. The momentum equation for near surface current by omitting the baroclinic pressure term then reduces to

\[-U'2\Omega \sin \theta = -\frac{1}{\rho} \frac{\partial p}{\partial y} = -\frac{g \Delta h}{\Delta y}\]

\[V'2\Omega \sin \theta = -\frac{1}{\rho} \frac{\partial p}{\partial x} = -\frac{g \Delta h}{\Delta x}\]

where

\(h, \Delta x, \Delta y\) are the sea-level difference and the distance in \(x\) and \(y\) directions.

These near sea surface equations are a very simplified but frequently used form of the full equations. Note the linear relationship between the current speeds \(U', V'\) and sea-level slope perpendicular to the current direction. Note that these equations have a similar form to the parametric equations 7 and 8. In Figure 17, the computed geostrophic approximation of currents is compared to our observed currents calculated from AW25 and AE25. Geostrophic currents are computed from P-B and N-B sea-level differences. The distances between P and B and between N and B are 30 km and 40 km, respectively.
V.B.1. North-south Currents

For north-south currents, Figure 17A, the geostrophic approximation does quite well in explaining the current during the large flow reversal events between day 155 and day 190. The phase shift between observations and geostrophic currents is due to non-linear geostrophic adjustment processes. Note that the observed speed during this period is mostly below 60 cm/s.

During the nearly steady southward flow period (day 190 - 240), the north-south currents fluctuated between 50 cm/s and 80 cm/s. The geostrophic flow approximation overestimates the observed currents most of the time. When the observed currents exceeded 60 cm/s, geostrophic currents deviated significantly, by almost a factor of 2. Between day 190 - 197, 200 - 208, and 226-236, the observed currents are 70-80 cm/s but geostrophic currents are nearly 120 cm/s. It is interesting that when the observed currents return to about 60 cm/s, the geostrophic currents reduce close to the observed currents. The flow is ageostrophic at current speeds more than 60 cm/s. The current speed of 60 cm/s is apparently a threshold above which non-linear processes in the north-south currents become significant.

As suggested by this geostrophic approximation, we expect the role of non-linearities and friction in the flow to increase significantly when the current speed exceeds 60 cm/s. The scaling of the ratio between the Coriolis and the non-linear terms \( \frac{V^2/L_y}{fU} \) increases from about 0.1 to .9 when the current speed increases from 10 cm/s to 75 cm/s. At a current speed of 60 cm/s, the ratio is 0.7. This scaling ratio is known as the Rossby Number. In words, in the Lombok Strait the geostrophic approximation for north-south current appears adequate if the Rossby Number is smaller than 0.7.
The internal Rossby radius of deformation $R$ is the horizontal scale at which the effect of earth's rotation is comparable to the buoyancy effects (Gill, 1982 p.205). It also can be interpreted as the distance an internal wave travels in an inertial period. It is defined as

$$R = \frac{\sqrt{g'}D_o}{|f|}; \quad g' = g\frac{\rho_2 - \rho_1}{\rho_2}$$

with $g'$ is the reduced gravitational acceleration, and $\rho_1, \rho_2, g, f$ are the density of upper and lower layers, the gravitational acceleration ($=9.8 \text{ m s}^{-2}$), and the Coriolis coefficient ($=2.16 \times 10^{-5}$), respectively. Parameter $D_o$ is the vertical length scale of upper layer. In the Lombok Strait, the depth of pycnocline layer is about 100m. The mean density value of the upper pycnocline layer is about 1024 kg m$^{-3}$. The mean density value of the lower pycnocline layer is about 1027 kg m$^{-3}$. The first mode internal Rossby radius of deformation for this two layer system is about 80 km, two times the width of the Strait (40 km).

The effect of geostrophic adjustment at the mid Strait region was also indicated by hydrographic data from January, June, and September 1985. Geopotential at the sea surface relative to 500 dbar, computed from STD data is about 1.0 dynamic-meter ($\text{m}^2\text{s}^{-2}$) higher at the east side than at the west side of the mid Strait region (Arief, 1988). This geopotential difference equals to about 8 - 12 cm east-west sea-level difference.
Figure 17. (A). The average observed north-south currents of AW25 and AE25, and computed geostrophic currents. (B). As (A) but for east-west currents. All data are 4-day lowpass filtered records.
V.B.2. East-west Currents into the Badung Strait

The geostrophic approximation for east-west currents through the Badung Strait estimated from the P-B sea-level difference fails to estimate the observed flow, even when the observed currents fall below 20 cm/s (Figure 17B). Geostrophic currents are about three times stronger than the observed currents. Despite this relationship, temporal variabilities in currents and P-B sea-level difference do have a strong linear relationship as shown in Figure 16B. It indicates a close dynamical link between P-B sea-level slope and east-west currents in mid-Lombok Strait (AW and AE). However, the dynamical relationship is still unclear from the present data sets.

The Badung Strait region is shallow, less than 100m deep. The bottom friction in this Strait is expected to be significant to its circulation. This means that friction and nonlinear terms cannot be omitted from the momentum equation. In addition, Nusa Penida is located at the narrow part of the Lombok Strait due to the protrusion of the Lombok Island (Figure 13). The width of the Lombok Strait at this region is about half of that at the mid-Strait. Geostrophic flow at this narrow part of the Strait is not expected considering that the flow at mid-c Strait region is barely geostrophic. Additional complexities arise because Nusa Penida is also close to the Lombok Sill. The effects of hydraulic flow due to topographic shoaling and contraction in the sill region are expected to affect Nusa Penida sea-level. The effect of topographic contraction and shoaling on circulation has been well observed, for example in the Strait of Gibraltar by Armi and Farmer (1988).

V.B.3. Sea-level in the Badung Strait

Even though the Coriolis force does not dominate the flow through the Badung Strait-Lombok Strait opening, it is still an important to the flow. The influence of the Coriolis
force is indicated by the low variability of sea-level at Nusa Penida compared to the sea-level at Batutiga. The flow in this two-channel system causes opposite effect on the sea-level at Nusa Penida. Westward flow through the Badung Strait would raise sea-level at Nusa Penida. Coincident southward flow through the southern opening of Lombok Strait, in contrast, would lower the sea-level at Nusa Penida. During the reversal flow period, the flow through the Lombok Strait raises sea-level at Nusa Penida. However, the expected eastward flow from the Badung Strait reduces sea-level at Nusa Penida. Thus Nusa Penida will tend to resemble a modal point or zone of minimal sea-level variability as observed.

It has been shown above that the cross-strait sea-level differences are related to the currents as shown in Figures 21 and 22. It appears that non-linear processes amplify cross-strait slopes generated by geostrophic adjustment. There is no current observation in the Badung Strait. During strong currents toward the Indian Ocean, we may expect low sea-level along the Bali Island coast inside the Badung Strait and high sea-level along the west and north Nusa Penida coast. The reversed sea-level adjustments are expected to exist during northward flow events. As observed in the Lombok Strait, sea-level in the Strait (Figure 14) actually drops during southward flow, and rises during northward flow. The effect of these two processes will cause large sea-level fluctuations along Bali coast where they tend to reinforce. Along the Nusa Penida coast, however, sea-level fluctuation will be small because the effect of these processes tend to cancel out.

Sea-level observation in 1988-1990 (Figure 18) at Benoa, located on the southwest coast of Badung Strait (Figure 13) supports the sea-level characteristics described
above. The range of 4-day lowpass filtered sea-level variability at Benoa was 50 cm. Note that this range is two times the sea-level fluctuation at Batutiga (B), and more than about 3 times the range at Nusa Penida (P).

Figure 18. 4-day lowpass filtered sea-level at Benoa, the Badung Strait from May 1988 to March 1990. Note that the range of this sea-level variability is about two times larger then that at Batutiga (Figure 14).
V.C. CHAPTER SUMMARY

Sub-inertial flow in the Lombok Strait is controlled by sea-level variability in the Indian Ocean. The sea-level inside the Strait rises during northward flow periods, and drops during southward flows.

The sea-level variability is also affected by a tendency toward geostrophic adjustment of the currents. Superposition with large scale sea-level variability causes low sea-level variability at Nipa Bay and large sea-level variability at Batutiga and Benoa.

Low sea-level variability at Nusa Penida is due to a double channel effect.

The near surface circulation at mid-Strait region (AW and AE mooring sites) is close to geostrophic when the currents are below 60 cm/s. The flow becomes ageostrophic when the current speed exceeds 60 cm/s.

The Badung Strait is important to upper layer circulation in the Lombok Strait. The flow through its opening into the mid Lombok Strait is ageostrophic. However, sea-level difference across this opening and the east-west current at mid-Lombok Strait have a strong linear relationship in the correct sense of a geostrophic adjustment.
VI. LOMBOK STRAIT CURRENT AND REGIONAL SEA LEVEL RELATIONS

The sea-level difference between Nusa Penida and Batutiga (P-B) has been shown to be highly influenced by dynamics internal to the Strait. Thus P-B sea-level difference will likely not be useful in studying the relationship between along-strait sea-level slope and the flow through the Strait. It is observed that sea-level inside the Strait drops during southward flow, and rises during northward flow. This sea-level change suggests a relationship between the flow through the Lombok Strait and the sea-level difference between the Indian Ocean and the Flores Sea outside the Strait. To study this relationship, we need sea-level records outside the Strait i.e. across the Archipelago, to represent along-strait sea-level slope.

In this Chapter, we explore the relationship between sea-level and currents in the sub-inertial frequency (10-100 day period) band with length scale greater than the length of the Strait. For the period of 1985 current observations in the Lombok Strait however, the closest sea-level records available for this purpose are from Cilacap and Jakarta (see Figure 19B for location). Cilacap is on the south coast of Java about 700 km from the Lombok Strait, and Jakarta is on the north coast of Java about 1000 km from the Strait. For comparative analysis the along-strait currents at AE25, the longest record, is used to represent the Lombok Strait currents. Additionally, the temporal and spatial extent of 10-100 day period variability in sea-level is explored to determine the characteristics of these 10-100 day processes.

Due to the large spatial scale considered here, to reduce high frequency local variability in the data we use a 10-day lowpass filter. The variability of more than 100-day period is also removed to eliminate the monsoonal signal from the data. The length of overlapping data records of Cilacap and Jakarta sea-levels and the AE25 current is about 140 days, from January to May 1985.
VI.A. SEA-LEVEL AT CILACAP AND JAKARTA IN JANUARY-JUNE 1985

VI.A.1. Relation with Currents in the Lombok Strait

Remarkably, despite their separation, high correlation is observed between sea-level variability at Cilacap and the Lombok Strait currents. Maximum Pearson's correlation coefficient ($r$) between the Cilacap sea level and the along-strait currents is 0.87, with the Cilacap signal leading by 1 day. Figure 19 shows 10-100 day bandpass filtered time-series of along-strait current AE25, and of sea-level at Cilacap and Jakarta.

During northward flow events in the Lombok Strait, sea-level at Cilacap rose to 25 cm (see events marked 1 to 5 in Figure 19). This sea-level dropped when the northward flow decreased, and reached a minimum level during strong southward flow. Even fluctuations of a few cm in sea-level were reflected by the change in currents. The peaks reached during a major event in the sea-level and current occurred at different times. The time difference was 1 to 3 days. This time difference may represent propagation of a disturbance along the coast south of the Archipelago which can be eastward (events 2, 3 and 4) or westward (events 1 and 5). These disturbances in sea-level are clearly not the eastward Kelvin waves inferred from drifting buoys in the south of Java region by Quadfasel and Cresswell (1991). The propagation speed of disturbances, estimated from the distance between Cilacap and Lombok Strait (700 km) and the observed time-lags, varies between 3 m/s and 8 m/s. This value is close to the speed of atmospheric perturbations, such as cyclones in the Indian Ocean, a point explored further in Chapter VIII. This propagation speed suggests that these disturbances in sea-level south of the Archipelago propagate as forced-waves.
Figure 19. (A). 10-100 day bandpass filtered of sea-level at Cilacap and Jakarta, and along-strait currents at Lombok Strait. (B). Location map of observation sites. (C). Scatter plot between Cilacap sea-level and Lombok Strait current.
Similarity between temporal variability of sea-level at Cilacap and along-strait currents in the Lombok Strait suggests a close relationship between the two parameters. Parametric relationships between sea-level height, \( h \), at Cilacap and along-strait currents, \( \nu \), in the Lombok Strait, based on a least-square data fit are

\[
h(\text{cm}) = 0.28 \nu (\text{cm/s}) + 0.8
\]  \hspace{1cm} \text{(11)}

with Pearson's correlation coefficient equal to 0.87. The scatter plot between sea-level at Cilacap and currents in the Lombok Strait (Figure 19C) illustrates this linear relationship.

In contrast, there is no significant statistical correlation between Jakarta sea-level with either Cilacap sea-level or Lombok Strait currents. The temporal variability plot (Figure 19A), however, shows some regularity in fluctuations of sea-level at Jakarta, relative to sea-level at Cilacap. When sea-level at Cilacap rose (mark 1), sea-level at Jakarta dropped. Jakarta then rose while Cilacap dropped to its minimum level. A few days before strong sea-level rise at Cilacap during northward flow in the Lombok Strait (mark 2), sea-level at Jakarta dropped by 5 cm. This general pattern appears repeatedly in the rest of the data shown in Figure 19A even though the exact magnitude and timing of Jakarta fluctuations relative to Cilacap varied. It is, however, apparently that sea-level fluctuations at Jakarta are generally the opposite to that at Cilacap, and the change at Jakarta occurs prior to change at Cilacap. Pearson's correlation coefficient between relative sea-level difference Cilacap - Jakarta and along-strait currents in the Lombok Strait is 0.86. This value is similar to the correlation coefficient between Cilacap sea-level and the along-strait currents.
VI.A.2. Dynamical Interpretation

Unfortunately suitable sea-level observations in the region north of Lombok Strait are absent. Based on later data (after 1986) at Surabaya, located on northeast Java, and the Kalukalukuang Bank region (Figure 2, DM-site), the fluctuations of sea-level in the 10-100 day period band were below 10 cm and mostly about 5 cm. For a first approximation, we can then assume that the sea-level fluctuation at the Indian Ocean side of Lombok Strait is the same as that at Cilacap, and the fluctuation at the Flores sea side of the Strait is zero. The empirical linear relationship (Eq.11) may then be considered a relationship between sea-level slope and currents along the Lombok Strait. Dynamically, this relationship resembles a frictionally balanced flow. In this system, the flow is driven by a pressure head generated by the sea-level difference and balanced by the friction force opposing the flow. This system is commonly observed in a narrow or shallow tidal channel system (Pugh, 1987).

From the momentum equation, the system may be expressed as

$$ F_r = -\frac{1}{\rho} \frac{\partial p}{\partial y} $$

where the pressure gradient term on the right is balance by the friction force, $F_r$, on the left side of the equation. The friction force can be parameterized linearly by the surface layer velocity (Neumann and Pearson, 1966; Gill, 1982). Eq.12 becomes

$$ V c_R = -g \frac{\partial h}{\partial y} = -\frac{g}{\Delta y} \Delta h $$

where $V$ and $h$ are the along-strait current and sea-level height. The term $c_R$ is the Rayleigh friction coefficient (Gill, 1982). The terms $\rho$ and $g$ are water density and gravitational acceleration, respectively.
Solving for the sea-level difference

\[ \Delta h = -c_r \frac{\Delta y}{g} V \]  \[\text{[14]}\]

From the linear empirical relation Eq.11, the magnitude of \( c_r \) for the Lombok Strait is estimated by taking \( \Delta y \) equal to 60 km, the length of Lombok Strait.

\[ c_r \frac{\Delta y}{g} = 0.28 \]

\[ c_r = 0.28 \frac{g}{\Delta y} = 4.6 \times 10^{-5} \text{s}^{-1} \] \[\text{[15]}\]

This Rayleigh friction coefficient is about one order of magnitude higher than the value for the open ocean which is on the order of \( 10^{-6} \) to \( 10^{-5} \text{s}^{-1} \) (Neumann and Pearson, 1966). Considering the intense turbulent mixing in the Strait flow over the sill, this value of \( c_r \) appears quite reasonable. Our value is an average for the period from January to June 1985, during West monsoon and early East monsoon periods. In addition, Wyrtki (1961) found a close relationship between monthly along-strait sea-level difference and currents in the Malacca Strait. He computed a friction coefficient for the Malacca Strait equal to \( 3.6 \times 10^{-5} \text{s}^{-1} \).
VI.B. SEA-LEVEL AT CILACAP AND JAKARTA IN 1984

The limited duration of current observation in the Lombok Strait raises a question about the persistence of this low frequency variability. Are these specific features in 1985 or do they occur every year? The only clue to answer this question is from sea-level data from before and after 1985. Sea-level data at Cilacap and Jakarta are available from March 1984 to June 1985. It is interesting to see the characteristics of sea-level in 1984 compared to those in the January-May 1985 period already described. These sea-level records, after 10 - 100 day bandpass filtering, are shown in Figure 20.

This 1984 - 1985 filtered sea-level at Cilacap is characterized by 20 - 40 cm high fluctuations. The magnitude of fluctuation increases from 20 cm in March 1984 (day -300) to 40 cm in February 1985 (day 50) (Figure 20A). The recurrence time for sea-level rise at Cilacap was about 50 days. This Figure shows persistent low frequency variability in Cilacap sea-level over the entire 15 months. The variability is strongest in the West monsoon period.

Similar low frequency variability also exists in Jakarta sea-level (Figure 20B). The magnitude of fluctuations here is less than at Cilacap, typically 15-20 cm. Many sea-level peaks in Cilacap sea-level have a negative counterpart peak in Jakarta sea-level (marked by b in Figure 20), some with apparent time lag of about 7 days. For some events, however, the sea-level at Cilacap and Jakarta rose about the same time (marked by a in Figure 20).
Figure 20. 10-100 day bandpass filtered sea-level record at Cilacap (A) and Jakarta (B), from March 1984 to May 1985. The in-phase events between Cilacap and Jakarta are marked by a, and ~180° out of phase events are marked by b. The phase relationship for the events marked by ? are uncertain.
These sea-level characteristics at Cilacap and Jakarta may be explained by spatial properties of the wind field. Increasing westerly winds along both the southern and northern coasts of Java Island will raise sea-level at Cilacap but lower sea-level at Jakarta due to Ekman transport toward and away from the coasts, respectively. Oppositely, easterly winds will lower sea-level at Cilacap and raise sea-level at Jakarta. Winds south of the Archipelago are normally stronger than the winds in the Java Sea region. This difference in wind force will cause the sea-level response at Jakarta to be smaller than at Cilacap as we observed. Large spatial variability in wind field across the Java Island (about 500 km wide) is common. There are numerous identifiable events where large deflection in Cilacap and Jakarta sea-level are 180° out of phase. These appear to be due to the extent of the large wind system which is spatially coherent across the Archipelago. A detailed discussion on winds is presented in Chapter VIII.

It is concluded that variability in the 10-100 day period band is a regular feature in sea-level data. This consistency in 10-100 day variability in sea-level can be reasonably extended to the currents in the Lombok strait. The strong linear relationship between Cilacap sea-level and along-strait currents in the Lombok Strait in January-May 1985 (Eq.11), even for weak changes in Cilacap sea-level, confirms this extrapolation. In other words, we may expect low frequency variability in currents in the Lombok Strait as an annually recurring feature of circulation in the Strait.
VI.C. SEA-LEVEL AT BENOA, PADANG, AND SURABAYA IN 1988-1990

As shown in the preceding two sections, regional sea-level variability is clearly related to variation of transport in the Lombok Strait. It is instructive to continue this discussions on the longest time period of available data.

Three simultaneous periods of sea-level observations are available from May 1988 to March 1990 at Benoa (Lombok Strait), Padang (west Sumatera), and Surabaya (east Java). These data and a location map are shown in Figure 21. Padang and Benoa are on the Indian Ocean side of Indonesia as is Cilacap. The distance between Benoa and Padang is about 2000 km. Surabaya is on the northern coast of Java and represents sea-level on the southern coast of Java Sea. The distance between Benoa and Surabaya is less than 500 km.
Figure 21. 10-100 day bandpass filtered sea-level record at Benoa (A), Padang (B), and Surabaya (C), from May 1988 to March 1990. Asterisk (*) marks indicate the in-phase sea-level events between Benoa and Surabaya. Locations of sea-level observations are marked on the location map.
VI.C.1. Temporal Variability:

Temporal variability of sea-level records at Benoa, Padang, and Surabaya are shown in Figure 21. Locations of sea-level observations are shown in the location map in the Figure. All data are 10 - 100 day bandpass filtered. The existence of significant low frequency variability at these three locations is shown in the Figure. Benoa sea level has the strongest fluctuations with magnitude varying from 20cm to 50cm. This strong variability at Benoa, as discussed before, is influenced by dynamics inside the Badung Strait. Sea-level events at Benoa (dashed lines in Figure 21) occurred repeatedly at 20 to 80 day intervals during this 2 year period. Sea-level at Padang also showed strong fluctuations of up to 30 cm with similar recurrence times as the sea-level at Benoa. Fluctuations of sea-level at Surabaya were mostly less than 10cm. These fluctuations, however, still showed a repetition of sea-level events as at Benoa even though they were not as distinct as at Benoa. The magnitude of fluctuations at Benoa and Padang are comparable to that at Cilacap, and the magnitude of fluctuations at Surabaya is comparable to that at Jakarta.

Peaks of sea-level at Benoa led those at Padang by between 1 and 15 days. Lag-correlation analysis indicates that sea-level fluctuation at Benoa led sea-level at Padang by 4 days with Pearson's correlation coefficient 0.54. The relationship between sea-level fluctuations between Benoa and Surabaya is similar to that between Cilacap and Jakarta. Only the largest peaks are in phase (marked by * in Figure 21) between Benoa and Surabaya.
From only the two observation points at Benoa and Padang, it is difficult to determine if these sea-level fluctuations are directly related as part of the westward propagation of disturbances in sea-level along the Indian Ocean side of the Archipelago. As discussed previously (Section VI.1.A) analysis of sea-level at Cilacap and the current at the Lombok Strait indicates that disturbances can propagate either westward or eastward as forced-waves. Note that Cilacap, Benoa and the Lombok Strait are at the same latitude, 8.5°S. Padang, however, is located at 1°S. Due to its near equator position, the Ekman transport mechanism is not expected to cause the sea-level setup at Padang.

The variability shown in these three locations from 1988 to 1990 strengthens the previous conclusion that this low frequency variability is persistent on an interannual time scale. That sea-level variability is stronger on the Indian Ocean coast (Benoa, Cilacap, Padang) than at the Java Sea coast (Surabaya, Jakarta) indicates that low-frequency oscillations in the Lombok Strait current originate in the Indian Ocean.

VI.C.2. Spectral Characteristics:

Spectral analysis on the sub-inertial current in the Lombok Strait currents is hindered by the relatively short data record. However, as has been shown, there is a strong relationship between the Lombok Strait currents and Cilacap sea-level. In addition, the characteristics of sea-level at the Indian Ocean coast along the Archipelago (Benoa, Cilacap, and Padang) are similar. These previous results make it possible to infer characteristics of sub-inertial variability in Lombok Strait currents from relatively long sea-level records at Benoa and Padang (over 2 years).
Power spectra of sea-level at Padang, Benoa, and Surabaya for the frequency band of 0.01-0.1 cpd are shown in Figure 22. Within the frequency band of 0.01-0.1 cpd (10-100 day period), there are three spectral regions which are important in representing the variability in sea-level records discussed above. The spectral regions are centered at about frequency 0.025 cpd, 0.057 cpd, and 0.075 cpd (or at periods about 40-days, 17.5-days, and 13.5-days), and marked by A, B, and C in Figure 22, respectively.

The frequency band about 0.025 cpd (40-60 day period band) is known as the Julian-Madden frequency band. The energy within this frequency band dominates sea-level at Benoa and Padang. It is one order of magnitude higher than the peak energy at
the two other bands centered at 0.057 and 0.075. Sub-inertial variability in sea-level at Benoa and Padang and, inferentially, sea-level at Cilacap and currents in the Lombok Strait are mostly related to processes in the Julian-Madden frequency band. In the Indian Ocean, the existence of energetic events at the Julian-Madden frequency has been observed in the Somali Current (Mysak and Mertz, 1984), and in the South Equatorial Currents (Quadfasel and Swallow, 1986). Quadfasel and Swallow (1986) suggested direct meteorological forcing for this oscillation in currents. Kindle and Thompson (1989), in contrast, based on numerical models of the Western Indian Ocean, driven by monthly mean winds, suggested that 40-60 day period oscillations in currents may also be generated by instabilities in the current system itself.

At Surabaya, in contrast, the energy in this Madden-Julian band is relatively insignificant. It is one order of magnitude lower than those at Benoa and Padang. This different energy concentration at 40-60 day periods between Surabaya and Benoa and Padang indicates that sea-level processes in the Indian Ocean and Java Sea are uncoupled in this frequency band.

The second energy peak is centered at 0.057 (about 16-20 day period band) and is significant at Benoa and Padang but not at Surabaya (Figure 22B, peak B). The energy in this band at the two Indian Ocean sites is one order of magnitude smaller than the energy in the 30-50 day period band.

Luyten and Roemmich (1980) observed the existence of a 20-30 day period oscillation in currents along the Equatorial Indian Ocean region. Kindle and Thompson (1989) identified a 26-day period oscillation in currents from a numerical model of the Western Indian Ocean. Furthermore, the NOARL global numerical model showed a propagation
of Equatorial Waves from the Equatorial Indian Ocean to the southwest Sumatera Coast. The waves then propagated southeast along the south coast of Java as coastal Kelvin waves (John Kindle, personal Communication). From the present study, based only on sea-level observations at Benoa and Padang, it is not clear if the sea-level variability in the 16-20 day period band represents coastal Kelvin waves or some other phenomena.

Another significant energy peak in sea-level records at the Benoa, Padang, and Surabaya sea levels is centered at 0.075 cpd (about 12.5 to 15 day period) band. This energy peak was also observed in the currents and sea levels in the Lombok Strait (Figure 8 and 14b), and at sea-level stations in the Indian Ocean - Western Pacific region. This energy peak apparently related to low frequency shallow water tidal constituents (Pugh, 1987).

VI.C.3. Coherence Analysis:

Further investigation of sea-level variability at Benoa and Padang utilized coherence and phase-lag analysis between the two data sets. The coherence between sea level at Benoa and Padang is shown in Figure 23A. As expected from power spectra curves, the coherence between Benoa and Padang is high at the three energy peak bands (A, B, and C) shown in Figure 22. This high coherency can be interpreted as meaning that processes within those frequency bands at Benoa and Padang are related.

The phase-lag between Benoa and Padang, shown in Figure 23B, shows that sea-level variability at Benoa leads Padang. The mean values over the 40-60 day (peak A), the 16-20 day (peak B), and the 12.5-15 day (peak C) period bands are 30°, 180°, and
330°. For peak A, this phase-lag equals to about 5 days time-lag. The lagged correlation analysis discussed earlier indicated about a 4 days time-lag between Benoa and Padang sea-levels.

Figure 23. Coherence and phase-lag spectra between sea level at Padang and Benoa. Note in phase spectra, Benoa leads Padang.
VI.D. VARIANCE DISTRIBUTION OF 40-60 DAY BAND

Having established the importance of energy in the 40-60 day period band along the Indian Ocean coast of the Archipelago, we extended the investigation to wider spaced locations. The objective here is to explore the possible source of the 40-60 day phenomenon.

Variances of sea-level within the 40-60 day period band at 16 locations are presented as scaled circles in Figure 24. The sea-level records are from the Indian Ocean coast of Java and Sumatera islands (Benoa, Cilacap, and Padang) and Thailand (Phuket), at the north and west Australian coast (Darwin, Port Hedland, Freemantle), in the Indian Ocean (Cocos Island), in the Sunda Shelf region (Surabaya, Jakarta, and Kuantan), and in the Sulawesi and Maluku Seas (Jolo, Davao and Bitung), and Western Pacific (Guam and Yap Islands). The variances plotted in Figure 24 is obtained by integrating the autospectrum of each sea level record over the stated frequency band. The data sets are from TOGA Sea Level Center.

Variance in the 40-60 day band is strongest at the stations south of Java (about 40 cm²). The variance values between 10 to 20 cm² are found in the Indian Ocean from Freemantle, Cocos Island to Padang, and in the South China Sea (Kuantan). Low variance regions, mostly below 5 cm², are located in northern Australia (Port Hedland and Darwin), inside the Archipelago, and in the Western Pacific. This variance distribution strongly suggests that the 40-60 day variability in the sea-level is an Indian Ocean process. Significant variance in this 40-60 day band at Cocos Island, a small island in the Indian Ocean, suggests that the 40-60 day process in the Indian Ocean is not restricted to the coastal region.
Figure 24. Spatial distribution of variance in 40-60 day period band in the Eastern Indian Ocean to Western Pacific Ocean region. Data are from TOGA Sea Level Center. The number inside the circle indicates the variance in cm$^2$ within 40-60 day period band.
VI.E. CHAPTER SUMMARY

Along-strait currents in the Lombok Strait in the 10-100 day period band linearly correlate with the variability in sea-level at Cilacap. The currents are driven by variability of sea-level at the Indian Ocean side. The circulation through the Strait represents a friction dominated flow system with Rayleigh friction coefficient of $4.6 \times 10^{-5}$ s$^{-1}$.

The intraseasonal variability in the sea-level along the south coast of the Indonesian Archipelago and in current in the Lombok Strait occurs persistently on a year to year basis. The 40-60 day period processes dominate the variability and are limited to the Indian Ocean region.
VII. ATMOSPHERIC PRESSURE AND CURRENT RELATIONS

It has been shown in Section V.1 that in the 10-100 day period band currents in the Lombok Strait are driven by sea-level variability south of the Archipelago. An investigation into the mechanisms driving this variability in sea-level is presented in this and the following Chapter. In this Chapter, we will show the linkage between sea-level variability and both large scale atmospheric processes and cyclone/typhoon occurrences. First, characteristics of surface barometric pressure in the 10-100 day period band in the Lombok Strait are described and related to currents in Lombok Strait. Then, the study area is extended over the Eastern Indian Ocean to the Western Pacific to cover large scale processes in the atmosphere.

VII.A. ATMOSPHERIC PRESSURE IN THE LOMBOK STRAIT

There are two meteorological stations in the Lombok Strait region, at Ngurah-Rai, Bali and at Ampenan, Lombok. Surface barometric pressure records from these two locations were identical in the 10-100 day period band. So, only data from Bali is shown in Figure 25. Figure 25 is the 10-100 day bandpass filtered Bali barometric pressure record overlaid with along-strait currents.

In this Figure the continuous line represents along-strait currents, and the connected dots represent atmospheric pressure. There is an apparent connection between the occurrence of troughs in atmospheric pressure and northward flow in the Lombok Strait currents. The current events followed the pressure troughs by 6 to 18 days. Short lines above the curves in Figure 25 indicate the elapsed time in days between the appearance of the atmospheric trough and the maximum northward flow events.
Figure 25. (A). 10-100 day bandpass filtered surface atmospheric pressure at Ngurah-Rai (Bali) and along-strait currents in the Lombok Strait in 1985. Note that pressure troughs lead northward flow by 6-18 days. (B). Power spectra of surface atmospheric pressure at Ngurah-Rai, Bali. A, B, and C are positions of sea-level spectrum peaks at Benoa as shown in Figure 22.
Magnitudes of variability in pressure and currents also show similar trends. They are strong in the January - June period, and small in the July-September period. Note a secondary atmospheric pressure trough about day 50 (marked by an arrow) was the pressure trough due to Cyclone Jacob in the Indian Ocean. Cyclone Jacob was the most intense cyclone in the Indian Ocean in 1985 (Kuuse, 1985) and its track was about 500km from the Lombok Strait.

Pearson's correlation coefficient between atmospheric pressure and currents is relatively low ($r = 0.5$) for this 300 day long record. The irregularity of the time-lag between events in the atmosphere and currents seems the cause for this low correlation. This time-lag indicates that connectivity between atmospheric processes and the Lombok Strait currents is not a direct one. However, this relationship show that the variabilities in currents as well as in sea-level south of Java are led by changes in the atmospheric system.

The spectrum of surface atmospheric pressure at Ngurah-Rai, Bali, for year 1985 is shown in Figure 25B. The variance increases toward low frequency dominated by the monsoon (longer than 100 day period). Within the 10-100 day period band, the 50-100 day band has the highest variance but the peak is not significant. This concentration of variance is at a little longer period than that in the sea-level data as shown in Figure 22. The positions of spectral peaks in Benoa sea-level are marked by dashed-lines in Figure 25B. The variance in atmospheric pressure is also high at the other two peaks in the sea-level spectrum at about 17.5 day and 13.5 day periods (0.057 cpd and 0.075 cpd).
Note that one mbar atmospheric pressure is equivalent to about 1 cm of sea-level height. The variances in sea-level at Benoa and Padang (Figure 22) are about 1-2 orders of magnitude higher than the variance in the atmospheric pressure at Ngurah-Rai, Bali. This difference means that the effect of inverse-barometric changes on sea-level data is small and can be neglected in the sea-level data.

VII.B. REGIONAL ATMOSPHERIC PRESSURE

Because of this strong relation between atmospheric pressure and currents in the Lombok Strait, it is significant to this study to investigate the 10-100 day period phenomenon in the atmosphere over a larger domain. In Figure 26, 10-100 day bandpass filtered surface atmospheric pressure data from 13 stations over the Eastern Indian Ocean to Western Pacific region in 1985, including the data from Bali, are plotted. The observation stations are all in the tropical latitudes band, and their locations are given in Figure 26B.

Figure 26 shows clearly that variability at 12 different locations are essentially similar in magnitude and phase to the variability at Ngurah-Rai, Bali. Pressure trough events occurred almost simultaneously from Palau Island (West Pacific) to Dili (North Australia) to Andaman Island and Cocos Island (Indian Ocean) covering a region over 5000 km long. Variability is strong in the January-June period (6 mbars), and weakens in the July-November period (4 mbars). These pressure troughs occurred repeatedly every 20-60 days. The data from these 13 locations strongly point out that the 10-100 day period phenomenon in the atmosphere is a large scale process within the tropics. This finding agrees with studies of intraseasonal atmospheric processes which concentrated in the 40-50 day period band, e.g., Madden and Julian (1972); Wang and Rui (1990).
Madden and Julian (1972) using pressure and upper-air data set of 5 - 10 years duration, detected a global scale for the 40-50 day events moving eastward from the Indian Ocean to the East Pacific. They suggested this atmospheric phenomenon corresponded to the movement of the Indian Ocean and Pacific Ocean Walker Circulation Cells.
Wang and Rui (1990) studied the climatology of the Tropical Intraseasonal Convection Anomaly (TICA) from 10 years of outgoing long wave radiation data. They identified 122 events which propagated in three directions: eastward (77 strong to moderate events), northward (27 weak events), and westward (18 weak events). They found the eastward propagating events moved along the equator from Africa to the mid-Pacific. The major formation region of these events was in the Western Central Indian Ocean and the west coast of Equatorial Africa. Eighty percent of the events took place in the period between December and May. The northward and westward propagation events mostly occurred between May and October. They also suggested a close association between the meridional movement of eastward propagating events and the annual movement of the monsoon trough.

Chu and Sikdar (1983) also observed near-in-phase surface atmospheric pressure systems from Central Asia to the northwest Borneo (Kalimantan) region during December 1978. They estimated the propagation speed of perturbations in atmospheric pressure as 10° longitude per day (about 12.5 m/s). Contrary to Madden and Julian (1972), Chu and Sikdar (1983) associated this large scale perturbation to the variability in the East Asia local Hadley circulation. The East Asia Hadley circulation is the most dominant atmospheric feature during the northern winter in this region (Chang et al., 1979).

Imbedded in the large scale atmospheric pressure trough were other pressure troughs more limited in their spatial scale but of considerable importance to this study. These pressure troughs were caused by cyclones, and recorded only at observation stations within the range of the cyclones, usually about 500 km or less. In Figure 26, the cyclonic troughs may appear as a distinct secondary trough. In the Figure, Jacob, Kirsty and Margot are the
name of cyclones in the Southeastern Indian Ocean, and the other are the name of typhoons in the Western Pacific. Typically at least one cyclone or typhoon occurred between 3 to 15 days after the occurrence of a large scale pressure-trough.

VII.C. CYCLONES AND TYPHOONS

The tropical cyclone season in the Southeast Indian Ocean region is normally in the December - April period. The typhoon season in the Western Pacific region, occurs in the June - October period. In 1985, there were 10 cyclones in the Southeast Indian Ocean (Kuuse, 1985) and 26 typhoons in the Western Pacific region (Steinbruck et al., 1985). It is common for two or three cyclones/typhoons to exist together, or to occur one after the other within a 5-10 day period. Murakami et al. (1984) showed that this clustering of cyclones is characteristics of cyclonic disturbances from the Arabian Peninsula to the West Pacific.

During the 300 day interval in 1985, 10 out of 12 events of northward flow in the Lombok Strait occurred at the same time as the cyclones in the Indian Ocean or typhoons in the West Pacific Ocean. The exceptional 2 events occurred at day 160 and day 220 marked by ??? in Figure 27A. The northward flow events were strong, reaching more than 40 cm/s during the Indian Ocean cyclones. Coincident appearance of northward flow in the Lombok Strait and the cyclones/typhoons also occurred in 1984. Figure 27B shows the episodes of sea-level raise at Cilacap and their related cyclones and typhoons events in the 1984 and 1985. In Figure 27B the events marked by ??? indicate that no cyclone/typhoon occurred at that time. This figure demonstrates that the relation between the north-
ward flow in the Lombok Strait and a cyclone / typhoon is a persistent phenomena. Occasionally episodes of northward flow are not associated with generation of a typhoon or cyclone.

The occurrence of the large scale atmospheric troughs typically were followed by a change in large scale wind field, i.e., the westerly winds south the Indonesian Archipelago intensified. In addition, some cyclones or typhoons were generated. Details about wind characteristics are discussed in the following Chapter.

The combined number of cyclones in the Southeast Indian Ocean and typhoons in the Western Pacific was more than the number of northward flow events in the Lombok Strait. The cyclones and typhoons that related to large scale pressure troughs in 1985 as shown in Figure 25 were generated in the monsoon troughs (Kuuse, 1985; Steinbruck et.al., 1985). The monsoon troughs typically extended in an east-west direction, one located about 10°S and one located about 10°N. These cyclones and typhoons (Figure 28) exist east of 90°E in the Indian Ocean, and south of 20°N between the Philippines and Guam in the Western Pacific. The cyclone and typhoon tracks (Kuuse, 1985; Steinbruck et.al, 1985), including their tropical depression phase, are given in Figure 28.

The cyclones (Jacob, Kirsty, and Margot) started from tropical storms close to the Indonesia Archipelago, and reached their cyclone phase between 10°S and 15°S. Most of the Southeast Indian Ocean cyclones moved southward and dissipated over cold water regions or over the Australian continent. The typhoons (Irma, etc.) developed at about 10°N and reached their maximum strength typically in the region west of the Philippines. These typhoons commonly propagated westward toward the continent of Asia.
The typhoon tracks in Figure 28 show large distances of more than 3000 km between the typhoons and the Lombok Strait. This distance is more than 10 times the radius of the typhoon generated wind field even for a super-typhoon. Direct influence of these typhoons
on the intraseasonal variability of the Lombok Strait circulation is not expected. The propagation of typhoon generated waves from the Western Pacific to the south of Java is not observed in the 40-60 day period band in the sea-level data (Figure 24). The distance between cyclones in the Indian Ocean and the Lombok Strait is closer, but still more than 500 km. These data suggest strongly that the northward flow events in the Lombok Strait and cyclones/typhoons are mutually related to a large scale atmospheric phenomenon as indicated by large scale pressure troughs.

Figure 28. Tracks of cyclones and typhoons that occurred at the same time as northward flow events in the Lombok Strait (Kuuse, 1985 and Steinbruck et al., 1985). Note distance between typhoons and Lombok Strait is more than 3000km.
VII.D. CHAPTER SUMMARY

Large scale atmospheric phenomena appear as successive atmospheric pressure troughs every 20-60 days over 5000km from the Indian Ocean to the Western Pacific region.

The variability in the 10-100 day period band of sea-level along the south of Java and of the currents in the Lombok Strait are associated with this large scale atmospheric phenomena.

The occurrence of large scale pressure troughs led the episodes of northward flow in the Lombok Strait, sea-level rise south of Java, and some cyclones in the Southeast Indian Ocean and typhoons in the Western Pacific Ocean which were generated in monsoon troughs.

The typhoons in the Western Pacific do not drive the northward flow in the Lombok Strait. The typhoons and the flow, however, are associated with the same large scale atmospheric phenomena indicated by large scale atmospheric pressure troughs.
VIII. WIND AND CURRENT RELATIONS

It has been shown in Chapter VII that in the 10-100 day period band the sea-level rise at Cilacap and the northward current in the Lombok Strait are related to large scale atmospheric pressure processes. The atmospheric pressure values, however, are too small to drive directly these coastal events. It is hypothesized that these sea-level and current events are caused by the intensification of westerly (eastward) wind along the south coast of the Archipelago. Due to the effect of the earth's rotation, this westerly wind will displace water northward, causing setup of sea-level along the south coast of the Indonesian Archipelago. In this Chapter, we will investigate this hypothesis. First, characteristics of observed winds at the Lombok Strait are discussed and compared with the wind-stress data from the Supplementary Fields Data Set of the ECMWF/TOGA Advanced Operational Analysis data sets. The Lombok Strait current relationships will be examined further in three case studies which represent the response type relationship observed in 1985.

VIII.A. CHARACTERISTICS OF WIND

In 1985-86, winds in the Lombok Strait were measured 6-hourly at Ngurah-Rai, Bali, and at Ampenan, Lombok (Figure 3). The distance between the two locations is about 100 km across the Strait. The wind observations were archived as daily wind mean with its direction in 45° bins, and as daily maximum wind with its direction in 10° bins. In this study, the daily maximum wind data is used because of its better direction resolution. The data are 450 days long from January 1985 to March 1986. Another source of wind data used in this study is the wind-stress from the Supplementary Fields Data Set of the ECMWF/TOGA Advanced Operational Analysis data sets. These data are derived from the ECMWF's 6-hour forecasts. The length of wind-stress data available for this study is 450
101 days long from January 1985 to March 1986. Using these data we compare the wind-stress offshore the Lombok Strait and offshore Cilacap with the maximum wind observation at Ngurah-Rai for year 1985. All wind data used are 10-100 day bandpass filtered using the FFT technique.

VIII.A.1. Observed Winds

Bali and Lombok Islands are mountainous and the effect on the land-based wind observations is expected to be significant. It is instructive to evaluate to what extent the wind data from these locations may represent large scale winds in the 10-100 day period band.

In the 10-100 day period band, the wind at Ngurah-Rai was similar to that at Ampenan as shown in Figure 29. In the Figure, positive values are the northward and eastward components. The east-west wind components were dominant with speeds in the range of -6 m/s to 6 m/s, while the north-south wind components were usually less than 2 m/s. The wind records show that significant eastward and westward events occurred at about 20-40 days interval.

This wind similarity between these two locations separated by about 100 km shows the existing large scale wind in the 10-100 day bandpass despite the topographic effect. This means that the length scale of this wind phenomenon is larger than the topographical effect scale. There were, however, some wind events with significant wind speed differences observed between the locations. Those events are marked by (*) in the Figure. Geographic location of the observation stations and the extent of the wind field are apparently the causes for these discrepancies. Ngurah-Rai (Figure 3) is at the
southern tip of Bali Island, and Ampenan is about the mid-Strait latitude. The location at Ngurah-Rai is more exposed to the Indian Ocean winds than at Ampenan, so the Ngurah-Rai winds are considered to best represent observed wind in the Lombok Strait.

The magnitude of observed wind at Ngurah-Rai was weak compared to regular wind conditions over the ocean. Weaker wind on the coast than offshore is commonly observed (Hsu, 1981, 1986, 1988) because of the influence of land and the existence of Atmospheric Planetary Boundary Layer. The monthly mean wind speeds offshore are between 1.5 to 2.5 times stronger than the winds at the coast. This factor is a function of height of the planetary boundary-layer and drag coefficient (Hsu, 1988).

Compared to the atmospheric pressure data (Figure 25), the fluctuations in the Lombok Strait winds occurred at higher frequency. This characteristic may indicate that the winds were more affected by mesoscale processes than the atmospheric pressure. Power spectra of Ngurah-Rai north-south and east-west wind components are shown in Figure 30. The monsoonal band (longer than 100 day period) dominates the wind spectra. The variance of the east-west winds is about 1 order of magnitude higher than that of the north-south wind. In the 10-100 day period band, the spectra do not have any distinctive peaks. There is a tendency, however, for concentration of wind energy in the 20-50 day period band for the east-west winds similar to the spectrum of atmospheric pressure shown in Figure 25. The characteristics of these spectra differ from the spectrum of sea-level at Benoa (Figure 22). This suggests that the variabilities of currents and sea-level in the 10-100 day period band are not a consequences of local processes.
Figure 29. 10-100 day bandpass filtered winds at Ngurah-Rai, Bali, and at Ampenan, Lombok, in 1985. Note the wind events marked by (*) are not observed at the both locations. Winds toward east and north are positive.
VIII.A.2. ECMWF Wind-Stresses

Wind-stress offshore from the Lombok Strait (115.5°E/9°S) and Cilacap (109°E/9°S) were extracted from the Supplementary Fields Data Set of the ECMWF/TOGA Advanced Operational Analysis data sets. The data are the averaged values in a 3.375° square (average of nine points of the ECMWF grids) and thus they represent wind characteristics with length scale of about 350 km. The 10-100 day band-pass filtered time series of these east-west wind-stresses are shown in Figure 31 together with local wind-stress at Ngurah-Rai, Bali, for comparison. As shown in the previous Section, the Ngurah-Rai wind may represent the Lombok Strait region. Its length scale is about 100 km, the distance between Ngurah-Rai and Ampenan.
The wind-stress is commonly expressed as

\[ \tau = C_D \rho_\alpha |\overrightarrow{\vec{u}}| |\overrightarrow{\vec{v}}| \]  

where \( \tau \) is wind-stress (Nm\(^{-2}\)), \( C_D \) is drag coefficient with a value ranging from 1.1x10\(^{-3}\) to 2.1x10\(^{-3}\) (Gill, 1982), \( \rho_\alpha \) is air density (1.3 kg\(m^3\)), and \( \overrightarrow{\vec{v}} \) is wind velocity (m/s). The wind-stress at Ngurah-Rai is computed by using \( C_D = 1.3 \times 10^{-3} \).

Temporal variability of wind-stress offshore Lombok Strait (Figure 31A) and offshore Cilacap (Figure 31B) are similar, while their magnitudes are slightly different. Their Pearson's correlation coefficient is 0.84 with zero-lag. This indicates that the fetch of this wind is more than 700 km, the distance between Cilacap and the Lombok Strait. There are 12 events of westerly (eastward) wind with wind-stress more than 0.1 N/m\(^2\) as shown in Figure 31. Between two consecutive westerly wind events there is usually a strong easterly wind event.

The two strongest wind events in 1985 were during cyclones Jacob (event 2, day 50) and Margot (event 4, day 106). The peak values in this 10-100 day bandpass wind-stress are equal to about 25 m/sec (50 knot) of westerly wind. For cyclone Margot, its maximum wind speed was estimated at about 50 m/sec while the center of the cyclone was about 500 km from Cilacap (Kuuse, 1985). In general, the wind speed in this 10-100 day band is less than 15 m/sec.
Figure 31. The 10-100 day lowpass filtered east-west wind-stresses of the ECMWF data offshore the Lombok Strait and Cilacap, and the east-west wind at Ngurah-Rai, Bali, in 1985. Dashed-lines indicate the peaks of eastward wind events.
Compared to the wind-stress at Ngurah-Rai, the ECMWF wind-stress is about 13.8 times stronger than the Ngurah-Rai wind-stress (Figure 32). Figure 32 shows the scatter diagram between the ECMWF wind-stress offshore the Lombok Strait and at Ngurah-Rai. The ECMWF offshore wind is about 3.7 times stronger than the observed wind at Ngurah-Rai. Hsu (1981) found off Kodiak Island and Cape Hatteras that the offshore monthly winds were up to 2.5 times stronger than winds at the coast.

The Pearson's correlation coefficient between these two data sets is 0.56 which is statistically significant at the 95% confidence level. Despite strong land effects, the Ngurah-Rai wind data reflect the of strong east-west wind events offshore the Lombok Strait as shown in Figure 31. Note the vertical scale of Figure 31C is one tenth of that Figure 31A and 31B.

Figure 32. Scatter plot of 5-day mean of the 10-100 day bandpass filtered east-west wind-stresses from the ECMWF data offshore the Lombok Strait and the observed data from Ngurah-Rai, Bali, in 1985.
VIII.B. WIND AND CURRENT RELATIONSHIP IN THE LOMBOK STRAIT

The relation between wind and northward flow in the Lombok Strait is shown in Figure 33. In this Figure the east-west component of wind-stress offshore Lombok Strait is compared to the long-strait currents at AE25 in the 10-100 day period band in 1985. Dashed lines in the Figure mark the peaks of the northward flow.

The linear relationship between the east-west wind-stress offshore Lombok Strait and the along-strait current in the Lombok Strait is weak. The maximum Pearson's correlation coefficient is 0.41 when the wind led the current by 4 days. This relationship indicates that the flow in the Lombok Strait is not simply driven by local wind. Despite this weak relationship, the events in the wind and in the current are matched quite well (Figure 33A and B). This relationship hints that the current is linked to the wind but influenced strongly by other factors, a subject which will be elaborated further in the next Section.

There are three different responses of the Lombok Strait current on the offshore wind. The first response, northward flow in the Lombok Strait, occurs when the westerly wind prevails (events 1 to 8 in Figure 33). The second response, again northward flow occurs during easterly wind periods (events a - d). The third response, is when westerly wind is not accompanied by a northward flow in the Lombok Strait (events A - D).

Response Type 1:

In the first response type (event 1 to 8 in Figure 33), the peak wind led the peak current up to 6 days except for event 5. For event 5, the peak wind was about 4 days after the peak current. The deceleration of the southward flow in the Lombok Strait, however, occurred soon after the westerly wind began to blow. This suggests that the response time of the Lombok Strait current to the offshore wind is short, less than 1 day. The observed time lag
between the peak wind and current thus indicates the existence of other processes besides the local wind. As shown in Figure 19, the sea-level rise at Cilacap led the northward current in the Lombok Strait from 1 to 4 days during events 2, 3, and 4 in Figure 33. This sea-level - current relation supports the arrival of a sea-level disturbance to the Lombok Strait from a remote area. In addition, this remote forcing may explain the disparity of the wind-stress and the northward flow relationship as shown in Figure 33C. The origin of the remote forcing, however, cannot be pinpointed from the present data set. The fact
that the sea-level at Cocos Island shows energy in the 40-60 day period band (Figure 24) opens the possibility that the disturbances are generated in the Southeastern Indian Ocean as well as along the south coast of Java.

Response Type 2:

This remote forcing concept explains also the apparent paradox of the second response type (events a to d in Figure 33). The easterly wind offshore the Lombok Strait will decrease sea-level south of the Lombok Strait. This will enhance the southward flow in the Lombok Strait rather than the northward flow as we observed. Sea-level observation in the Java Sea and Kalukalukuang Bank (see Figure 2 for locations) support this rejection. As expected during these events, remote forces overpower the effect of local wind on the circulation in the Lombok Strait.

Response Type 3:

In the third response type (event A - D), the westerly wind blows offshore the Lombok Strait. The remote forcing is apparently absent or negates the local wind effect on the Lombok Strait circulation. This type three suggests that the effect of local wind is actually too weak to overturn the inherent southward pressure gradient force that drives the southward flow in the Lombok Strait.

In summary, this qualitative analysis strongly indicates that the northward flow mechanism is not solely a local process. The circulation in the Lombok Strait is influenced strongly by remote processes occurring in the Southeastern Indian Ocean and in the coastal region south of Java. The effect of local wind on the Lombok Strait flow is expected to be small compared to the effect of remote forcing.
VIII.C. CHARACTERISTICS OF THE LARGE SCALE WIND FIELD

In the previous Section remote forcing was shown to largely drive the intraseasonal
circulation in the Lombok Strait. In this section, we investigate the regional wind field
extending from the Eastern Indian Ocean to the Western Pacific for this remote forcing
based on the Supplementary Fields Data Set of the ECMWF/TOGA Advanced Operational
Analysis data sets. In particular we study the evolution of the wind-stress field for each of
the three types of wind - current relationship identified previously. In the following sub­
sections we discuss the characteristics of the wind field during events 2, b, and A in Figure
33. Event 2 is during the cyclone Jacob period. Event b is during the typhoon Irma period.
Event A occurs during a westerly wind period, but when no northward flow is observed in
the Lombok Strait.

VIII.C.1. Response Type 1 - Event 2

The wind-current event 2 (Figure 33) occurs during the onset period of Cyclone
Jacob. Cyclone Jacob was formed in the Timor Sea on 15 February 1985, and dissi­
pated over southwest Australia on 27 February 1985 (Figure 28). It maintained a
minimum pressure of 950 mbars for 5 days, from 15.5°S/113°E to 16°S/106°E with
maximum surface wind estimated at 53 m/sec (98 knots) (Kuuse, 1985). Figure 34
shows a snapshot of the 10-100 day bandpass filtered ECWMF wind-stress field every 3
days from February 14 to February 23, 1985. Letters J and H are the center locations of
cyclones Jacob and Hubert. Cyclone Hubert developed from the Joseph Bonaparte
Gulf, and existed from February 10 to 19, 1985.

Prior to cyclone Jacob, on February 14, 1985, the 10-100 day period bandpass
wind-stress field in the Indian Ocean was weak (Figure 34). Southeasterly wind prevailed
over the northern Australian ocean areas, and westerly wind existed at about 90°E/7°S. The wind pattern changed significantly in the following 12 days. In the Indian Ocean, the westerly wind increased in strength and extended eastward reaching the Timor Sea on February 17. The westerly wind, then, retreated again westward, and covered only a region east of 110°E and south 14°S on February 26. This evolution of the westerly wind extension was in agreement with the southwest movement of cyclone Jacob from the Timor Sea to the west of the Australian continent.

Figure 34 also shows that the change of wind pattern in this 10-100 day period band occurred simultaneously in the Indian Ocean and the Western Pacific region during this February 14 - 26 period. A cyclonic wind pattern east of the Philippines on February 14 was replaced by southwesterly winds on February 20. The cyclonic wind pattern developed again on February 26.

This regional wind evolution shows that the westerly wind is a large scale phenomenon in the Southeastern Indian Ocean. The westerly wind has already developed prior to cyclone Jacob reaching its maximum stage. The westerly wind intensified during the development of Jacob, but eastward extension of the wind decreased. It is expected that the Ekman transport from this large scale westerly wind will significantly displace sea water northward and cause setup of sea-level along the south coast of Java. The large area coverage and long duration of this westerly wind is strongly related to the strongest northward flow (86 cm/s) in the Lombok Strait in 1985 (Figure 33). The peak of this northward event, however, occurred when the westerly wind offshore the Lombok Strait had already ceased (Figure 33).
Figure 34. Evolution of 10-100 day bandpass filtered wind-stress field during the cyclone Jacob. The wind-stress is filtered from the ECMWF data set.
In summary, the northward flow begun prior to the onset of cyclone Jacob, reached its peak on February 20, and still existed on February 26 when easterly wind already prevailed along the coast of the Indonesian Archipelago. The evolution of the northward flow, however, followed the change in the large scale westerly wind. This sequence of events shows that the northward flow is not caused by the cyclone nor the local wind, but by remote large scale westerly wind.

VIII.C.2. Response Type 2 - Event b

The wind-current event b (Figure 33) occurs during the onset period of typhoon Irma. Typhoon Irma was generated at the eastern extension of the monsoon trough and reached its typhoon level on June 27, 1985 (Steinbruck et.al., 1985). Irma had minimum pressure of 957 mbars with maximum wind of 46 m/s (90 knots). The typhoon moved in the north-northwest direction and dissipated on July 01, 1985 west of Japan (Figure 28). Prior to Irma, typhoon Hal transited the area north of the Philippines from June 20 to June 26, 1985. Figure 35 shows the evolution of the 10-100 day bandpass filtered ECWMF wind-stress field during typhoon Irma. The Figure shows a snapshot every 3 days from June 19 to July 03, 1985. Letters I and H are the center locations of typhoons Irma and Hal.

During the period between June 19, 1985 and July 1, 1985 there was a significant change in the wind field over the Southeastern Indian Ocean. The development of a cyclonic wind pattern begun on June 19, reached its maximum on June 22 and is still recognizable on June 25. It was replaced by an anti-cyclonic wind pattern on June 28. During this June 19 - July 1 period the westerly wind south of the Indonesian Archipelago changed to easterly wind which blew strongly over the region between Bali and
Figure 35. Evolution of 10-100 day bandpass filtered wind-stress field during the typhoon Irma. The wind-stress is filtered from the ECMWF data set.
Australia. The duration of the easterly wind, however, was only about 4 days. On July 1, the easterly wind had practically disappeared from the region south of Java - Timor Islands. In addition, the fetch of this strong easterly wind was limited from the Timor Sea to just about the Lombok Strait. In the region west the Australian continent and south of 10°S, strong southeasterly wind was replaced by strong northwesterly wind. The reversal of these winds occurred at the time when typhoon Irma reached its maximum on June 27, 1985.

During this short period of limited fetch easterly wind in the Indian Ocean, there was a 37 cm/s northward flow event in the Lombok Strait. This northward flow reached its peak on June 27, 1985 when the easterly wind reached its maximum offshore of the Lombok Strait. This suggests that this northward flow is related to the development of extensive westerly wind associated with the cyclonic circulation south of the Archipelago a week before the easterly wind. The local easterly wind showed little influence on the flow through the Lombok Strait.

During the typhoon period of June 19 - July 1, the change in the 10-100 day band wind field in the northern hemisphere also occurred. The southwesterly wind over the South China Sea increased in strength, while the easterly wind in the Western Pacific north 10°N weakened during the development of typhoons Hal and Irma.

In brief, there is similarity between response type 1 and response type 2 in that significant change of the wind field occurs simultaneously in the Indian Ocean and in the Western Pacific during the development of both cyclone and typhoon. In both response types the development of extensive westerly wind led the northward flow in the Lombok Strait by about 3 days. The wind directly offshore of the Lombok Strait does not have a
significant influence on the flow through the Lombok Strait. There is also, however, a fundamental difference between these two response types. Namely that the westerly wind strengthens during the cyclone development, but during the typhoon the westerly wind is replaced by the easterly wind but with a limited fetch and duration.

Comparing the effect of wind during response type 1 and 2 on the flow through the Timor passage we may speculate that opposite results may occur. For response type 1, the flow through the Timor passage into the Indian Ocean will decrease or reverse as in the Lombok Strait because of the Ekman transport generated by the westerly wind. For response type 2, however, unlike the Lombok Strait we may expect an increase of the flow into the Indian Ocean through the Timor - Savu passages due to the extensive strong southeasterly wind.

VIII.C.3. Response Type 3 - Event A

The event A (Figure 36) is different from to the two events discussed above. During event A, response type 3, westerly wind blew offshore of the Lombok Strait but no northward flow was observed in the Strait (Figure 33). Figure 36 shows the evolution of the 10-100 day bandpass wind-stress in the Southeastern Indian Ocean and Western Pacific region from day 120 to day 132. The peak westerly wind-stress is at day 129.

On May 1, 1985 (day 120), the wind field in the Indian Ocean - Western Pacific in general is calm (Figure 36). From May 4 to May 13, 1985 the wind field over the Southeastern Indian Ocean was characterized by strong southeasterly wind in the region south of 5°S, west of the Australian Continent. Along the south coast of Java to the Timor Islands the wind was weak and with a fetch less than 1000 km. It developed from May 4 to May 10 (Figure 36).
Figure 36. Evolution of 10-100 day bandpass filtered wind-stress field from day 120 to day 132. The wind-stress is filtered from the ECMWF data set.
During this event a period, in the Indian Ocean (Figure 36) although the westerly wind blew for several days, no northward flow occurred in the Lombok Strait, instead the southward flow increased (Figure 33). It is suggested that this wind was too weak to significantly affect the Lombok Strait flow. In contrast, strong southeasterly wind persistent for about 10 days in the region west of 110°E, apparently countered the possible effect of this westerly wind, by removing water from the region between the Archipelago and Australia via the Ekman transport. In other words, this southeasterly wind reduced the sea-level in this region and enhanced the flow into the Indian Ocean as observed in the Lombok Strait current.

Comparing the three response types, there are similarities in the relationship between the northward flow in the Lombok Strait and the large scale wind. The wind offshore of the Lombok Strait does not show any significant role in the flow through the Strait. The wind in the Southeastern Indian Ocean between 90°E - 125°E, in contrast, related to the northward flow in the Strait which reaches peak flow 2 to 4 days after the maximum development of the wind. The duration and fetch of the wind are important factors in determining the mode of flow through the Lombok Strait.
VIII.D. CHAPTER SUMMARY

Wind observations at Ngurah-Rai, Bali, are affected significantly by land. The magnitude of observed wind is expected to be about one-third of the magnitude of wind offshore. The observed wind, however, still provides source limited information about the east-west wind events in the intraseasonal band.

The wind and the flow through the Lombok Strait show three types of response. First, northward flow may occur during large scale westerly wind events. Second, northward flow may occur during local easterly wind. Third, no northward flow may occur during local westerly wind. These three types of response are explained by the spatial and temporal properties of the large scale wind field.

The east-west wind component offshore of the Lombok Strait region affects the flow through the Lombok Strait. The effect, however, is weak compared to the effect of remote forcing due to the wind activity in the Southeastern Indian Ocean.

The northward flow events in the Lombok Strait are forced by westerly wind in the Southeastern Indian Ocean. The fetch and duration of the westerly winds are important contributors to the magnitude of the northward flow episodes in the Strait.
IX. SUMMARY, CONCLUSIONS, AND SUGGESTIONS

IX.A. SUMMARY

The Lombok Strait has a strong bi-directional current system oriented in the northeast-southwest direction. Above 200m depth, the currents reach 150 cm/s in the mid Strait region, and 300 cm/s in the sill region. The monthly mean current varies between 5 cm/s in March 1985 and 70 cm/s in July 1985, southward into the Indian Ocean. Strong tidal circulation still exists at 800m depth where the currents reach 75 cm/s, and their monthly mean shows a tendency for southward flow up to 5 cm/s.

In the sub-inertial frequency band, inside the Strait there are strong relationships between currents and sea-level differences in the mid Strait region. The parametric relationship between the north-south current \( V \) and the Nipa Bay - Batutiga sea level difference \( H_{N-B} \) is

\[
H_{N-B} \text{ (cm)} = -0.12V \text{ (cm/s)}
\]

There is also a parametric relationship between the east-west current \( U \) at mid Strait and the Nusa Penida - Batutiga sea level difference \( H_{P-B} \)

\[
H_{P-B} \text{ (cm)} = -0.25U \text{ (cm/s)}
\]

apparently due to a significant transport westward into the Badung Strait.

Despite its low latitude (8.5°S), the effect of the earth's rotation on the circulation in the Lombok Strait is important. The along-strait flow in the Lombok Strait is near geostrophic when the flow is less than 60 cm/s. This geostrophic flow is expected to occur only in the upper 50 m depth, the thickness of the density mixed-layer.
In the Lombok Strait surface circulation, the Badung Strait is expected to have a significant role. The surface layer speed through the Badung Strait is apparently comparable to that in the Lombok Strait proper. The flow through the Badung Strait, however, is ageostrophic even during the low current period.

This sub-inertial flow through the Lombok Strait is driven by the sea-level differences outside the Lombok Strait. This flow through the Strait is consistent with a frictional balance with a Rayleigh friction coefficient equal to $4.6 \times 10^{-5}$ sec$^{-1}$. This friction coefficient is nearly the same as the friction coefficient for the Malacca Strait (Wyrtki, 1961).

The flow through the Lombok Strait is strongly characterized by intraseasonal episodic northward flows of 20 to 100 cm/s that occur about every 20 to 60 days. In the mid-Strait region, the variability in this intraseasonal band (10 - 100 day period) represents 42% of the total variance in the current. It is about two times of the variance of the tidal currents.

This intraseasonal variability in currents appears driven by large scale intraseasonal atmospheric processes associated with atmospheric pressure troughs. These pressure trough episodes extend in phase over the Eastern Indian Ocean to the Western Pacific region and occur every 20 to 60 days. Following these pressure trough episodes, there are cyclones and typhoons frequently generated in the monsoon trough. These processes are also evident in the wind field over the Southeastern Indian Ocean - Western Pacific region. The intraseasonal wind over the Indonesian Archipelago is weak almost all the time.

One most important feature of the change in the wind field in the Southeastern Indian Ocean is the onset of episodes of large scale westerly wind south of Java. This westerly wind generates an Ekman transport toward the south coast of the Indonesian Archipelago causing sea-level setup of 10 to 30 cm at Cilacap and eastward to the Lombok Strait. The
extension of these westerly winds into the Java Sea causes sea-level drops (mostly less than 15 cm) along the northern coast of Java as observed at Jakarta and Surabaya. These intraseasonal episodes of sea-level rise south of the Archipelago overcome the inherent southward pressure gradient from the Western Pacific to the Indian Ocean and drive the northward flow events.

The effect of local wind offshore of the Lombok Strait on the sea-level rise is small compared to the effect of remote forcing from large scale wind fields in the Southeastern Indian Ocean. In addition, winds directly associated with cyclones and typhoons do not drive the northward flow in the Lombok Strait.

From the relationship between the wind offshore the Lombok Strait and the northward flow in the Strait, three flow responses in the Lombok Strait are recognized. They are: first, westerly wind with northward flow; second, easterly wind with northward flow; and third, westerly wind without northward flow. These are all explained in term of spatial and temporal variability in the large scale wind field.

The sea-level data from 1984 to 1990 in the Eastern Indian Ocean and Western Pacific region show the intraseasonal variability in sea-level along the southern coast of the Indonesian Archipelago to be a persistent phenomenon. Thus we conclude the same about northward flow events in the Lombok Strait. The energy of this variability concentrates in a 40-60 day period band, the Julian-Madden frequency band.

Although much remains to be clarified in detailed future studies we have gained a considerable understanding of the mechanics governing intraseasonal processes in the Lombok Strait region.
IX.B. CONCLUSIONS

Circulation Inside the Lombok Strait:

1. The Lombok Strait is a strong bi-directional current system in the northeast - southwest direction. Above 200m depth, the currents reach 150 cm/s in the mid Strait region, and 300 cm/s in the sill region. The monthly mean of currents varies between 5 cm/s in March 1985 and 70 cm/s in July 1985, southward into the Indian Ocean.

2. In mid Lombok Strait region, the circulation at 800m depth is dominated by tidal currents which reach 75 cm/s. The monthly mean of the flow at 800m show a tendency for southward flow up to 5 cm/s.

3. The current system in the Lombok Strait is a baroclinic flow of ~75 cm/s above 200m depth superimposed by a barotropic tidal flow of ~50 cm/s.

4. The effect of earth rotation on the circulation within the Strait can not be neglected. During period of flow less than 60 cm/s, along-strait flow in the upper 50m depth in the mid Strait region is near geostrophic flow. Parametric relationship between the current $V$ and the Nipa Bay - Batutiga sea level difference $H_{N-B}$ is

$$H_{N-B} \text{ (cm)} = -0.12V \text{ (cm/s)}$$

5. The Badung Strait is important to upper layer circulation in the Lombok Strait. The flow through its opening into the mid Lombok Strait is ageostrophic.
6. Sea-level difference across the northern opening into the Badung Strait and the east-west current at mid Lombok Strait have a strong linear relationship in the correct sense of a geostrophic adjustment. Parametric relationship between the current $U$ and the Nusa Penida - Batutiga sea level difference $H_{P-B}$ is

$$H_{P-B} (\text{cm}) = -0.25U (\text{cm/s})$$

Circulation through the Lombok Strait:

1. The flow through the Lombok Strait is strongly characterized by intraseasonal episodic northward flows of 20 to 100 cm/s that occur about every 20 to 60 days. In mid Strait region, the variability in this intraseasonal band (10 - 100 day period) represents 42% of total variance in the current. It is about two times of the variance of the tidal currents.

2. Sub-inertial flow through the Lombok Strait is controlled by sea-level difference between the Indian Ocean and Flores Sea entrances. Most of this sea-level variability is in the Indian Ocean signal. The sea-level inside the Strait rises during northward flow period, and drops during the southward flows.

3. The flow through the Strait is in a frictional balance with a Rayleigh friction coefficient for the Lombok Strait of $4.6\times10^{-5}$ s$^{-1}$. This friction coefficient is nearly the same as the friction coefficient for the Malacca Strait from Wyrtki (1961).
Intraseasonal Variability in the Lombok Strait Current

1. The intraseasonal variability in the currents is dominated by a 40-60 day period oscillation. Intraseasonal flow in along-strait current component, $V$, in the Lombok Strait can be empirically estimated from sea-level at Cilacap (south Java coast), $h$, as

$$h(\text{cm}) = 0.28* V(\text{cm/s}) + 0.8$$

2. The intraseasonal variability of the flow through the Lombok Strait driven by sea-level changes at the Indian Ocean, is generated by episodes of intensification of westerly winds in the Southeastern Indian Ocean and resultant Ekman convergence toward the coast.

3. Superposition of large scale sea-level variability and internal sea-level adjustments inside the Lombok Strait causes low sea-level variability at Nipa Bay or the east side of the Strait but large sea-level variability at Batutiga and Benoa or the west side of the Strait.

4. The intraseasonal variability in sea-level over 2000 km stretch along the south coast of the Indonesian Archipelago and in the current in the Lombok Strait is dominated by 40-60 day period processes. These processes in the ocean are limited to the Indian Ocean region.

5. The occurrence of large scale atmospheric pressure troughs extending from the Eastern Indian Ocean to Western Pacific region consistently led the episodes of northward flow in the Lombok Strait, sea-level rise south of Java, and cyclones in the Southeast Indian Ocean and typhoons in the Western Pacific Ocean by 5-20 days. These cyclones and typhoons are generated in the monsoon troughs.
Wind field:

1. Wind observations at Ngurah-Rai, Bali are affected significantly by land. The magnitude of the observed wind is about one-third of the magnitude of the offshore wind. This observed wind, however, still provides useful information about the east-west wind events in the intraseasonal band.

2. The influence of the wind offshore of the Lombok Strait on the along-strait flow in the Lombok Strait is weak compared to the effect of remote forcing due to the wind activity in the Southeastern Indian Ocean. We cannot use this offshore wind to predict the intraseasonal flow in the Lombok Strait.

3. Even though most northward flows in the Lombok Strait occur at the same time as the onset of cyclones in the Indian Ocean, these cyclone do not drive the northward flow events. The cyclones, however, enhance the strength of large scale wind over the Southeastern Indian Ocean.

4. The typhoons in the Western Pacific do not directly drive the northward flow in the Lombok Strait. The typhoons and the flow, however, are related to the same atmospheric phenomena marked by large scale atmospheric pressure troughs which (a) induce the westerly winds that (b) raise the sea-level along the coast of the Indonesian Archipelago resulting in (c) episodic northward flow reversals in the Lombok Strait.
IX.C. SUGGESTIONS

Future studies on the intraseasonal variability in the Lombok Strait circulation, or more generally on the Indonesian throughflow, need better observational data for at least one year period. Sea-level measurements along the south coast of the Indonesian Archipelago and in the Java-Banda Sea are vital in detecting the characteristics of propagation of perturbations in sea-level that drive the northward flow in the Lombok Strait.

These sea-level observations should cover the Sunda Strait (between Sumatera and Java), south of Java, the Lombok Strait, and the Timor-Savu passages. The location of sea-level stations should be away from the Strait opening to avoid local dynamics of the Straits.

Current meter moorings in these openings to the Indian Ocean are necessary, and should be located in the deepest and widest part of the Straits in order to reduce any significant local effects. The current measurements, however, should be concentrated in the upper 200 m depth. The use of Acoustic Doppler Current Meters will improve significantly the quality of data in order to understanding the dynamics of flow through the straits.

As shown in this study, wind observations close to the shore may not provide enough information to understanding the intraseasonal flow through the Straits. Several wind observations using buoys along the southern Indonesian Archipelago and in the Java-Banda Seas region are necessary. The availability of wind field data such as the ECMWF data set are important.

Studies of this intraseasonal phenomena using a numerical model approach need at least a basin wide model domain to accommodate the effect of large scale wind in the Indian Ocean. Because the intraseasonal variability in currents is driven by sea-level, the model should be designed to simulate the sea-level variability along the south of the Indonesian
Archipelago. The availability of a realistic wind data set to drive the model seems the major obstacle in an effort to simulate this intraseasonal variability phenomenon. The use of the ECMWF data set seems encouraging. Qualitatively, this ECMWF wind data set shows good agreement with the observed northward flow events in the Lombok strait.

In addition, the South Equatorial Current which flows northward west of the Australian continent and is deflected to the west in the region south of Java, is a major current system in this Southeastern Indian Ocean. The role of this current system in the intraseasonal variability is still unknown, but needs to be considered.
REFERENCES


APPENDICES

A. TIDAL CURRENTS IN THE LOMBOK STRAIT

The tidal currents in the upper layer of the Lombok Strait account for about 30% of the total variance. In general, the tidal currents rotate clockwise, and flow southward during ebb. The O1 (Principal lunar) tidal constituent is the main component in the diurnal tidal currents. The M2 (Principal lunar) tidal constituent is the main component in the semi-diurnal currents. Figure A1 and A2 show the tidal currents in the Lombok Strait. Figure A3 to A8 show the diurnal tidal currents (0.92-1.09 cpd or 22-26 hour period band), the semi-diurnal tidal currents (1.84-2.18 cpd or 11-13 hour period band), and the over-tide currents (>2.4 cpd or <10 hour period band). Table A1 lists the rotation angle of the major axis (relative to north) and the percentage of variance carried by the tidal currents. Table A2 is similar as Table A1 except for the tidal current components.

In the Lombok Strait, the strength of diurnal, semi-diurnal, and over-tide tidal currents vary significantly with time and location. The tidal currents in the sill region are about two times stronger than in the mid Strait region. In the mid Strait region, the diurnal tidal currents are strongest in the upper layer, but the semi-diurnal tidal currents are strongest at 800 m. In contrast, in the sill region both diurnal and semi-diurnal tidal currents are strongest in the upper layer.
TIDAL CURRENTS ALONG MAJOR AXES - MID-STRAIT REGION

Figure A1. Tidal currents in the mid Strait region along the major axes.
Figure A2. Tidal currents in the sill region along the major axes.
Figure A3. Diurnal tidal currents in the mid Strait region along the major axes. Note the interference between diurnal O1 and K1 constituents produces wave-envelope with period about 14 days.
Figure A4. As Figure A3 but for the sill region.
Figure A5. Semi-diurnal tidal currents in the mid Strait region along the major axes. Note the interference between semi-diurnal M2 and S2 constituents produces wave-envelope with period about 14 days.
Figure A6. As Figure A5 but for the sill region.
Figure A7. Over-tide tidal currents in the mid Strait region along the major axes.
Figure A8. As Figure A7 but for the sill region.
Table A1. Major Axes Rotation Angle Relative to North and the Percentage Variances of Major and Minor Axes of tidal currents

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B. UP-STREAM CURRENTS (DM-mooring)

DM-moored current meter was deployed between the Masalima Island and Kalukalu-kuang Bank during May - September 1985 period (Figure 2) to gain information about the up-stream condition of the Lombok Strait circulation. The location of the DM-mooring is about 500 km from the Lombok Strait. It had 3 current meters at depths of 25m, 100m, and 300m (DM25, DM100, and DM300), and was deployed in a topographic trough that connected the Makassar Strait and the Lombok Strait. The results of water mass analyses showed this topographic trough was one of the main paths for flow of the Western Pacific water mass that flows southward through the Makassar Strait (Arief, 1988). The upper meter at the DM-mooring was exposed to the eastward - westward reversing currents of Java - Flores Sea monsoonal currents.

B.1. Current characteristics

The currents at the DM-mooring site are not as strong as in the Lombok Strait. The instantaneous currents reach over 80 cm/s with the strongest current at 100m depth (DM100). Major axes of the flows, that explained over 74% of the variances, follow the topographic setting of the east boundary of the Sunda Shelf in a northeast-southwest direction (22° to 37°, Table 1).
The monthly mean north-south flow (Table 3) is between -10 cm/s and 2 cm/s, between -18 cm/s and 35 cm/s, and between -19 cm/s and -29 cm/s, at DM25, DM100, and DM300, respectively, during the June-September period. Negative value means a southward flow toward the Lombok Strait. The monthly mean east-west flow at the three current meter depths are less than 20 cm/s during the June-September period.

Bi-directionality flow at the DM-mooring site is clearly shown by current rose histograms (Figure B1). The current roses show the frequency distribution of current directions in 10-degree bins. These modal flow direction are the same as indicated by the currents principal axes directions. The characteristics of current roses are typical for tidal current with a lateral boundary and a strong low frequency flow, similar to the currents characteristics in the Lombok Strait. Tidal current rotation is significantly constrained by the southward main flow even at the upper current meter (DM25). A distinct eastward flow at DM25 (commented in Figure B1) is apparently caused by flow through a gap at the Kalukalukuang Bank.

Figure B1. Current rose histograms showing the frequency of current directions in 10-degree bins for hourly currents at DM-Mooring in June-September 1985. The label on each histogram identifies the current meter. The length of arrow represents a scale for the number of data points in each directional bin. The arrow points to north.
Vertical structure of currents at DM-mooring is shown by a scatter diagram plotting speed versus pressure (depth) in Figure B2. Strongest currents are located at mid-depth, about 100 dbars. Even though the distribution of flow in the upper 150 dbars appears symmetrical to the zero line, most current observations are toward the southwest. The tendency of the flow in the southwest direction is obvious at 300 dbars, about the position of the North Pacific salinity minimum core layer. The currents at this depth had a consistent flow at 30 cm/s southwestward during this 3 month period. An average vertical salinity profile from STD-casts during June 1985 about the DM-mooring site (Figure B2) shows that the currents at 25 dbars are related to the flow of low salinity upper layer water. The currents at 100 dbars are related to the flow at the upper part of salinity maximum layer (Northern Central tropical Water). The currents at 300 dbars are related to the flow of the salinity minimum layer (North Pacific Intermediate water).
Figure B2. Composite plot of current components vs depth at DM-mooring site during June-September 1985 (East monsoon). Salinity profile curve is the average of STD-casts around DM-mooring from June 1985.
B.2. Spectral Characteristics

Power spectra are shown in Figure B3 for current components along major axes at DM25, DM100, and DM300. The current records are 114 to 122 days long from June to September 1985 (see Table 2). The spectra are characterized by energy peaks in the sub-tidal, the diurnal, and the semi-diurnal frequency bands. The over-tide currents are not well developed. Due to short data records, spectral characteristics of the low frequency band are not well resolved. The $S1$ and $O1$ tidal constituents are the major components in the diurnal currents, except at DM25 where the $O1$ constituent is weak. The $M2$ and $S2$ tidal constituents are the major components in the semi-diurnal currents. The over-tide currents are the strongest at DM100, but they are still two orders of magnitude smaller than the diurnal current.

The variance of current components along major axes is dominated by the sub-tidal frequency band at scales longer than the 10 day period (Table B1). This frequency band explains over 60% of variance. The diurnal (0.8-1.2 cpd) and semi-diurnal (1.8-2.2 cpd) bands contribute between 20 to 30% of the total variance. The variance of each band is about the same, except at DM25. At DM25, the variance of the semi-diurnal currents is 2.5 times the variance of the diurnal currents.
Figure B3. Power spectra of currents components along major axes of DM-mooring current meters between 0.0 and 10. cpd ranges.
Table B1. Variance distribution at selected frequency bands

<table>
<thead>
<tr>
<th>FREQUENCY RANGE</th>
<th>CPD</th>
<th>DM25</th>
<th>DM100</th>
<th>DM300</th>
</tr>
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<tbody>
<tr>
<td>FREQUENCY</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RANGE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>FREQUENCY</td>
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</tr>
<tr>
<td>RANGE</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>FREQUENCY</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>R ange</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PERCENTAGE OF VARIANCE (%) for major axes currents</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DM25</td>
<td>DM100</td>
<td>DM300</td>
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<td></td>
</tr>
<tr>
<td>0.0 - 0.1</td>
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<td>68.2</td>
<td>61.8</td>
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<td>0.1 - 0.8</td>
<td>1.5</td>
<td>4.8</td>
<td>5.1</td>
<td></td>
</tr>
<tr>
<td>0.8 - 1.2</td>
<td>7.0</td>
<td>10.0</td>
<td>12.7</td>
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</tr>
<tr>
<td>1.2 - 1.8</td>
<td>0.3</td>
<td>0.9</td>
<td>1.0</td>
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</tr>
<tr>
<td>1.8 - 2.2</td>
<td>17.2</td>
<td>10.5</td>
<td>15.9</td>
<td></td>
</tr>
<tr>
<td>2.2 - 10.0</td>
<td>2.1</td>
<td>5.5</td>
<td>3.6</td>
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<tr>
<td>VARIANCE (CM/S)^2</td>
<td>615.7</td>
<td>786.1</td>
<td>277.4</td>
<td></td>
</tr>
</tbody>
</table>

B.3. Sub-tidal Currents

Figure B4 shows a stick plot of 10-day lowpass filtered currents at DM25, DM100, and DM300. Temporal variability of currents at the three depths are significantly different. The upper current meter (DM25) is dominated by the southwest flow. The flow is characterized by speed fluctuations with about a 25 day interval and weak flow after day 225 (mid-August). Surprisingly, during this East monsoon period no monsoonal westward current was observed at DM25. The position of DM25 is above the depth of the Sunda Shelf (60 m). During this period, westward surface flow is supposed to reach its maximum flow (Wyrtki, 1961). The Kalukalukuang Bank that was located east of mooring apparently has an important role in the circulation in the DM-mooring region.
The currents at DM100, about 75m below DM25, are about two times stronger than those at DM25 (Figure 18). The currents are still dominated by the southwest flow. Intra-seasonal fluctuations of the currents are even larger than that at DM25. In contrast, at depth 300m (DM300) the current is dominated by nearly steady strong southwest flow.

Figure B4. Stick plots of 10-day lowpass filtered currents at DM-mooring.
VITA

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DOCTORAL EXAMINATION AND DISSERTATION REPORT

Candidate: Dharma Arief

Major Field: Marine Sciences

Title of Dissertation: A Study on Low Frequency Variability in Current and Sea-level in the Lombok Strait and Adjacent Region.

Approved:

[Signatures]

Major Professor and Chairman
Dean of the Graduate School

EXAMINING COMMITTEE:

[Signatures]

Date of Examination: February 3, 1912