Turbulent mixing of water masses in Selayar Slope - Southern Makassar Strait

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Abstract. During the Java-Makassar-Flores (JMF) Cruise in August 2015 onboard the RV GEOMARIN 3, about 56 hydrographic CTD profiles have been acquired in the JMF confluence region of Makassar Indonesian Throughflow (ITF) and seasonal Java-Flores Current, where three intensive 24-h CTD "yoyo" measurement were carried out in Dewakang sill, Selayar slope and Kangean slope. The aim of this paper is to estimates the turbulent mixing value using Thorpe analysis, particularly in the Selayar slope - southern Makassar Strait. It is found that below surface potential density of \( \sigma_0 \) stratification of water masses is dominated by Makassar ITF with salty (34.6 psu) North Pacific thermocline water (NPSW) and less salty (34.42 psu) North Pacific lower-thermocline water (NPIW). However, above 24.25 \( \sigma_0 \) upper NPSW is drastically dominated by much fresher water (34.17 psu) from Java Sea, where inter-leavings of salinity profiles are obviously seen between surface density of 22.0 \( \sigma_0 \) and 23.50 \( \sigma_0 \). Estimation of turbulent vertical eddy diffusivity in this upper-thermocline are found strong with the order of \( 10^{-3} \) m\(^2\)/s. It is suggested that energetic internal tidal waves may contribute significantly to the vertical mixing in this complex topographic region.

1. Introduction

The Indonesian Throughflow (ITF) is a passage between Pacific Ocean and Indian Ocean. ITF plays an important role in the global ocean and climate change, significantly impacts the thermohaline circulation, because of its strategic position [1]. ITF enters Indonesian sea through two pathways, the west and the east path.

In the western path, ITF flows from the Pacific Ocean, enters to Sulawesi Sea, then anters into Makassar Strait, Banda Sea, part of it flows into Lombok Strait and a part flows into Flores Sea. Makassar Strait is a major ITF and brings North Pacific Water that carried by the Mindanao Current. North Pacific Water mass composed from North Pacific Subtropical Water (NPSW) in the thermocline layer, and North Pacific Intermediate Water (NPIW) in the lower thermocline layer [2, 3]. The minor component that carries by ITF is the South Pacific water [1]. The South Pacific water flows in the eastern path of ITF via the Halmahera Sea and the Maluku Sea to the Seram Sea, and then to the Banda Sea that consists of South Pacific Subtropical Lower Thermocline Water (SPSLTW) in the lower thermocline depths along [4, 5].

The NPSW has characteristic of high (maximum) salinity in the thermocline layer and NPIW has low (minimum) salinity in the lower thermocline layer [4]. Pacific waters flowing through the
Indonesian sea get decreased temperature and salinity [6, 7]. These water mass transformations resulted from vertical mixing that occurs in these waters. They have a significant impact not only through the water column but also on the atmosphere as the cooling of surface waters can influence the onset of deep atmospheric convection [8]. The strong turbulent mixing in the Indonesian sea induced from the sea surface cooling [9].

Selayar Slope - Southern Makassar Strait is one of the important fisheries area because they have high productivity and abundance of marine resources, include the Makassar Strait, Java Sea and Flores Sea (triangle seas). According to [10] in Selayar Sea have a high tidal energy value, so it is suspected that in Selayar Sea there can be happen turbulent mixing. The Indonesian sea is also a region of robust water mass transformation [10, 11]. There are several physical process involved in turbulent mixing that contributes to the transformation of water masses such as surface wind stress, internal tides and sea topographic slope [12]. This study aims to investigate the stratification of water masses and estimates of turbulent mixing using Thorpe analysis, particularly in the Selayar slope - southern Makassar Strait.

2. Methods

2.1. Study Area

This research used data were collected from the insitu observation in August 2015 conducted by Makassar – Java - Flores (MAJAFLOX) cruise. MAJAFLOX Cruise is a research conducted by Department of Marine Science and Technology, Faculty of Fisheries and Marine Science of IPB and Marine Geological Institute (P3GL) by using the Research Vessel of Geomarin III. Temperature, salinity and density data were recorded using Conductivity, Temperature, Depth (CTD) Sea Bird Electronic (SBE) V19 plus version with 4Hz frequency rate. There were 26 stations fully recorded and these stations were divided into three main area; i.e Java Sea zone, southern Makassar Striat zone, and Flores Sea zone. This study located in southern Makassar strait at position of 6.81°LS and 119.44°BT (station 22) (figure 1).

Figure 1. Station point of CTD “yoyo” measurement in MAJAFLOX Cruise 2015
CTD was reduced for 11 casts repetition during 24-hours “yoyo” with different depth, starting at 1:11 a.m. until 11.50 p.m. on August 13, 2015, but in the 7th cast there was a technical error so there were only 10 casts repetition. The wind data was downloaded from European Centre for Medium-Range Forecast (ECMWF) through homepage (www.ecmwf.int) with spatial resolution of 0.125° x 0.125°. Wind U10 (zonal) component and wind V10 (meridional) component were used in this study as well as wind data from 10 m above sea level on August 13, 2015. Tidal data in Selayar Sea were obtained from website of Indonesia Agency for Geospatial Information (www.tide.big.id) at 13 August 2015.

2.2. Data Analysis
Movement of up and down fluid to find a stable position is often referred to as buoyancy frequency or Brunt Vaisala (N) frequency. When a high density fluid is above a low density fluid, the fluid will move vertically in search of a stable position. Brunt Vaisala values can be calculated by the equation:

\[
N = \left( -\frac{g \, d\rho}{\rho_0 \, dz} \right)^{1/2}
\]

The first step to understand the mixing of water masses was determining the Thorpe displacement value \( (Td) \). This to identify the presence of overturn. The \( Td \) value was acquired by rearranging the density profile obtained from CTD data into the form of static stability, the density was arranged with the position of low density water mass above the high density water mass. Suppose a vertical density profile with \( n \) sample and density \( \rho_n \) observed at the depth of \( z_a \). If the sample at the depth of \( z_a \) was moved to the \( z_b \) depth to form a stable condition, then \( Td \) can be calculated by the equation (2) [13]:

\[
Td = z_a - z_b
\]

The schematic Thorpe displacement (Td) calculation show in figure 2. The Td value was very essential for observing the vertical distance range from mixing processes. Furthermore, the calculation of Thorpe scale value (LT) used equations (3) [13]:

\[
L_T = \left( \frac{1}{n} \sum_{i=1}^{n} Td_i^2 \right)^{1/2}
\]

where \( Td_i \) is the Thorpe displacement at the depth of \( i \) and \( n \) is the number of samples. Each \( L_T \) value is average of \( n \) samples in the desired depth. The average \( L_T \) value of this study was obtained by dividing the depth of the waters into three layers including the mixed (surface) layer, thermocline layer, and deep layer. Each layer has depth varies depending on the vertical profile of water masses.

Figure 2. The concept Thorpe displacement (Td) calculation
The value of turbulent kinetic energy dissipation (ε) described the turbulent active layer which will experience breakdown into smaller forms (dissipation) which will transfer energy to other media. The rate of turbulent kinetic energy dissipation per unit mass (ε) is calculated using the equation (4) [13, 14]:

\[ \varepsilon = L_0^2 N^3 \]  

(4)

with the value of Ozmidov length \(L_0\) using equation (5) [13, 14]:

\[ L_0 = 0.8L_T \]  

(5)

The turbulent vertical diffusivity value at each depth was subsequently obtained by using the equation (6):

\[ K_z = \frac{\gamma \varepsilon}{N^2} \]  

(6)

where \(\gamma\) is a mixing efficiency \((\gamma)\) is generally set to 0.2 [15] defined as the ratio between buoyancy flux and turbulence production, and \(N\) is the buoyancy frequency.

3. Result and discussion

3.1. Water mass static stability

Mixing water mass can be caused by unstable water columns that can be identified by calculating the Brunt Vaisala frequency value \((N^2)\). The results of the calculation of the Brunt Vaisala value \((N^2)\) for all CTD in Selaya waters showed in figure 3. The vertical profile of \(N^2\) is largely determined by the vertical profile of temperature and \(\sigma_\theta\), where different phenomena occur in each layer of the water column. Mixed layer in replications of 1-11 has an \(N^2\) value that ranges from 0-2x10^{-4} cycle/s with an average of 1.8x10^{-5} cycle/s. The value of \(N^2\) in the mixed layer is relatively lower than the thermocline layer. It can be inferred that the mixed layer has a low density vertical gradient value so that the water mass is less stable which can cause vertical mixing [16].

The value of \(N^2\) in the thermocline layer ranges from 0-1.6x10^{-3} cycle/s where the thermocline layer has the highest \(N^2\) value compared to the mixed layer and deep layer. This indicates thermocline layer that relatively more stable than the mixed layer and deep layer. The value of \(N^2\) is high on thermocline layer because of the presence of pycnocline layer where the density gradient increases sharply with the depth (pressure) [16]. In other words, the thermocline layer is the most stable layer compared to the mixed layer and deep layer.

The deep layer has a relatively lower value of water mass static stability compared with thermocline layer, which is 0-1.4x10^{-4} cycle/s. The lower value of \(N^2\) indicates that the deep layer is unstable and easily experiences turbulence or vertical mixing. It can be concluded that the higher the value of \(N^2\) in a layer, the greater the water mass static stability (low density water mass above high density water mass) of the layer, otherwise if the value of \(N^2\) becomes lower, the water column becomes more unstable or in conditions of static instability.

3.2. Thorpe Scale estimation

Thorpe scale estimation can be used to estimate the magnitude or small vertical mixing that occurs in a waters. Vertical mixing of water masses occurs in unstable water masses or has a structure of water mass instability characterized by low or even negative Brunt Vaisala frequency values. The instability of the water mass occurs in a condition where there is a high density water mass above the low density water mass. The displacement distance of the density value that readjusts to stable conditions is called the displacement of Thorpe or Thorpe displacement \((Td)\).
Figure 3. Brunt Vaisala frequency ($N^2$) profiles from 10 CTD cast measurement in Selayar Sea
Positive \( Td \) value indicates that the water mass moves up to that distance to look for static stability. This happens when a low density water mass is below a high density water mass. On the contrary the negative \( Td \) value shows the mass of water moving downward. The vertical \( Td \) profile of all replications can be seen in figure 4. The mixed layer has an average \( Td \) value ranging from -13 m to 13 m, with a minimum \( Td \) value on the 5th and maximum replications in the 3rd test. The high and low \( Td \) values in the mixed layer are assumed to be related to wind velocity blowing in these waters. The 5th cast was conducted when the waters were in low tide, and the 3rd cast was conducted when the waters transitioned from high to low tide.

The thermocline layer has the lowest \( Td \) value compared to the mixed layer and deep layer. The average \( Td \) value in the thermocline layer ranges from -5 m to 3 m. The lowest \( Td \) value is in the 1st profile, and the highest one is in the 3rd profile. This is presumably caused the thermocline layer is the layer that has the highest level of static stability compared to the mixed layer and deep layer.

\( Td \) value average in deep layers range of -8 m to 5 m. The largest \( Td \) value was found in the third profile with a range of -15 m to 14 m and the smallest in the 9th profile with a range of -3 m to 2 m. \( Td \) value in the deep layer is higher than thermocline layer. This condition is assumed to be due to the low value of the static stability of the water mass in the deep layer.

3.3. Analysis of vertical turbulent mixing
Vertical turbulence intensity can be seen from two parameters, turbulent dissipation kinetic energy (\( \varepsilon \)) and vertical diffusivity (\( K_z \)). The dissipation of turbulent kinetic energy value (\( \varepsilon \)) describes the turbulent active layer which will experience breakdown into smaller forms (dissipation) which will transfer energy to other media. The high value of turbulent energy dissipation indicates the release of a number of turbulent kinetic energies that modify the structure of the water mass in the mixing process. The estimated turbulent dissipation energy kinetic average value (\( \varepsilon \)) is presented in table 1.

Results from the average of each 50 meter depth interval indicate that the increasing depth of turbulent dissipation kinetic energy value (\( \varepsilon \)) decreases. The highest \( \varepsilon \) value in the mixed layer is \( 1.97 \times 10^{-6} \) Wkg\(^{-1}\). In the thermocline layer the value of \( \varepsilon \) is \( 2.37 \times 10^{-7} \) Wkg\(^{-1}\) and in the deep layer is \( 9.75 \times 10^{-8} \) Wkg\(^{-1}\). The low value of \( \varepsilon \) in the deep layer indicates that less kinetic energy is in turbulent flow which will experience breakdown into smaller forms (dissipation) to transfer energy. The highest kinetic energy value is in the mixed layer (\( 1.97 \times 10^{-5} \) Wkg\(^{-1}\)), this shows that the mixed surface layer is the layer where kinetic energy experiences the highest solution that contributes to the mixing process. The high dissipation of turbulent kinetic energy in the mixed layer is an indication of the strong vertical mixing of water masses. The average value of turbulent vertical diffusivity of all layers is \( 2.17 \times 10^{-9} \) Wkg\(^{-1}\). The average value of \( \varepsilon \) in these waters is almost the same as that obtained by [17] in the Ombai Strait by O (\( 10^{-6} \)) Wkg\(^{-1}\) and [18] in Dewakang waters which is equal to O (\( 10^{-7}-10^{-6} \)) Wkg\(^{-1}\). The highest \( \varepsilon \) value in the mixed layer is around \( 1.97 \times 10^{-5} \) Wkg\(^{-1}\). In the thermocline layer the value of \( \varepsilon \) is \( 2.37 \times 10^{-6} \) Wkg\(^{-1}\) and in the deep layer \( 0.09 \times 10^{-6} \) Wkg\(^{-1}\).

The mixing process that occurs in the surface layer of the sea can be caused by the influence of wind stress. The average wind stress at the time of data collection was 8 ms\(^{-1}\). The value of wind stress from 00.00 has increased and reached a maximum at 09.00-12.00 then experienced a decline again and increased at 21:00. The correlation between dissipation of turbulent kinetic energy and wind speed (Figure 5) shows that there is a high correlation value that is equal to \( r = 0.66 \). The correlation values obtained are categorized quite high, so it can be inferred that turbulent mixing that is strong enough in the mixed layer is caused by the velocity of the wind blowing on the waters. This is similar to results obtained by [18] and [19] where wind pressure is very influential on the turbulent mixing process on the surface layer.
Figure 4. Thorpe Displacement profiles from CTD cast measurement
Table 1. Average value dissipation of turbulent kinetic energy ($\varepsilon$) on every 50 meter depth from 10 CTD profiles.

| Depth (m) | Dissipation of Turbulent Kinetic Energy ($\varepsilon$) Wkg$^{-1}$ | Standard Deviation |
|-----------|---------------------------------------------------|-------------------|
| 25        | 1.97 x 10$^{-5}$                                   | ±3.51 x 10$^{-5}$ |
| 75        | 9.41 x 10$^{-6}$                                   | ±7.63 x 10$^{-6}$ |
| 125       | 1.97 x 10$^{-6}$                                   | ±1.29 x 10$^{-6}$ |
| 175       | 1.44 x 10$^{-6}$                                   | ±1.09 x 10$^{-6}$ |
| 225       | 6.39 x 10$^{-7}$                                   | ±4.53 x 10$^{-7}$ |
| 275       | 4.51 x 10$^{-7}$                                   | ±1.07 x 10$^{-7}$ |
| 325       | 2.77 x 10$^{-7}$                                   | ±1.86 x 10$^{-7}$ |
| 375       | 3.33 x 10$^{-7}$                                   | ±2.88 x 10$^{-7}$ |
| 425       | 1.26 x 10$^{-7}$                                   | ±8.01 x 10$^{-8}$ |
| 475       | 6.62 x 10$^{-8}$                                   | ±4.55 x 10$^{-8}$ |
| 525       | 3.23 x 10$^{-8}$                                   | ±3.20 x 10$^{-8}$ |
| 575       | 5.44 x 10$^{-8}$                                   | ±7.57 x 10$^{-8}$ |
| 625       | 3.32 x 10$^{-6}$                                   | ±1.13 x 10$^{-8}$ |
| 675       | 1.29 x 10$^{-7}$                                   | ±1.94 x 10$^{-8}$ |
| 725       | 5.03 x 10$^{-8}$                                   | ±1.03 x 10$^{-8}$ |

Figure 5. The correlation between dissipation of turbulent kinetic energy and wind speed on the surface layer in Selayar sea.

Table 2 shows the average value of vertical turbulent diffusivity with an average of 50 meters. In this Selayar waters the highest diffusivity value is in the mixed layer and by the depth increase the diffusivity value is decreases. The average value of vertical turbulent diffusivity of all layers is 5.28 x 10$^{-3}$ m$^2$s$^{-1}$. Mixed layer has the highest vertical turbulent diffusivity value (5.78 x 10$^{-2}$ m$^2$s$^{-1}$), then the thermocline layer is lower with 2.4 x 10$^{-3}$ m$^2$s$^{-1}$ and the deep layer is 1.35 x 10$^{-3}$ m$^2$s$^{-1}$. The $K_z$ value obtained is almost the same as that obtained by [20] of 6.0 x 10$^{-3}$ m$^2$s$^{-1}$ at the Dewakang threshold, [21] in the Labani Canal obtaining turbulent strength with high eddy diffusivity values obtained by O (10$^{-6}$ - 10$^{-2}$) m$^2$s$^{-1}$. The $K_z$ value is slightly larger than that obtained by [18], which is 2.83 x 10$^{-4}$ m$^2$s$^{-1}$ in.
Dewakang Threshold and greater than that obtained by [10] which is $1.5 \times 10^4 \text{ m}^2 \text{s}^{-1}$ in the Indonesian fishery region. The high value of $K_z$ in Dewakang Threshold is caused by a very intensive presence and an interaction between surface waves and lee waves [20].

Table 2. Average turbulent vertical diffusivity ($K_z$) on every 50 meter depth from 10 CTD profiles.

| Depth (m) | Turbulent Vertical Diffusivity ($K_z$) (m$^2$s$^{-1}$) | Standard Deviation |
|-----------|-------------------------------------------------------|--------------------|
| 25        | 5.78 x 10^{-2}                                       | ±7.97 x 10^{-4}    |
| 75        | 4.56 x 10^{-3}                                       | ±2.66 x 10^{-3}    |
| 125       | 2.22 x 10^{-3}                                       | ±1.22 x 10^{-3}    |
| 175       | 2.74 x 10^{-3}                                       | ±2.38 x 10^{-3}    |
| 225       | 1.61 x 10^{-3}                                       | ±1.04 x 10^{-3}    |
| 275       | 1.63 x 10^{-3}                                       | ±6.42 x 10^{-4}    |
| 325       | 1.74 x 10^{-3}                                       | ±1.24 x 10^{-3}    |
| 375       | 2.26 x 10^{-3}                                       | ±1.63 x 10^{-3}    |
| 425       | 1.26 x 10^{-3}                                       | ±7.37 x 10^{-4}    |
| 475       | 8.27 x 10^{-4}                                       | ±3.86 x 10^{-4}    |
| 525       | 9.14 x 10^{-4}                                       | ±6.39 x 10^{-4}    |
| 575       | 2.04 x 10^{-3}                                       | ±3.47 x 10^{-3}    |
| 625       | 7.56 x 10^{-4}                                       | ±2.96 x 10^{-4}    |
| 675       | 2.34 x 10^{-3}                                       | ±6.50 x 10^{-4}    |
| 725       | 1.00 x 10^{-3}                                       | ±4.80 x 10^{-4}    |

Figure 6 showed fluctuations in tidal predictions in Selayar waters which showed tidal semi-diurnal mixed tide. Tidal components have major frequency values and minor frequency values that can be described in an elliptical rotary graph. The value of tidal components depicted in elliptical shapes can be used to interpret tidal propagation or propagation. Figure 7 showed a M2 tidal component (semi-diurnal lunar principal) representing the moon's gravity with a circular orbit parallel to the earth's equator and K1 (solar diurnal lunar) is the declination of the solar and solar system [22]. Color shows the amplitude value in M2 components ranging from 0.2-0.6 m and in K1 components 0.2-0.3 m. The ellipse interprets the propagation or propagation of barotropic tidal energy. In the M2 component, there is a trace of the Makassar Strait turning towards the Selayar waters with diminishing speeds and on this Selayar Waters in the West-East direction. In component K1 the ellipse direction is almost the same as the M2 component, but the speed is very small. Barotropic tidal energy will be reduced due to basic friction and converted into internal currents when barotropic flow passes through different topography such as ridges and propagates far into the open ocean.
Figure 6. Tidal Fluctuations in Selayar Waters 24 hour period on August 13 2018 when CTD data collection (below)

Figure 7. Ellips tidal component M2 (left) and K1 (right) of Selayar Sea

Selayar Sea are dominated by M2 internal tides, this is in accordance [22] which states that Indonesian sea are generally dominated by M2 internal tides (semidiurnal) with a period of 12.42 hours. Tidal M2’s presence enters Indonesian waters from the Pacific Ocean and Indian Ocean with waves stronger than the Indian Ocean. M2 pairs enter from the Indian Ocean through the Timor Sea and some pass through the Lombok Strait and then enter the waters south of the Makassar Strait. At the southern end of the Makassar Strait there is a reduction in tidal amplitude caused by tides coming through the Lombok Strait that meet the tides of the Pacific Ocean that enter through the Makassar Strait [23].

4. Conclusion
Selayar Sea have dissipation value of turbulent kinetic energy ($\varepsilon$) and the turbulent vertical diffusivity ($K_z$) which is high and the value is decreases with increasing depth. The highest dissipation value of $\varepsilon$ $K_z$ were found in the surface layers with respective values of $1.97 \times 10^{-5}$ Wkg$^{-1}$ and $5.78 \times 10^{-2}$ m$^2$s$^{-1}$. This sea is categorized as an area that has strong turbulent mixing value. Wind stress is most likely affect the mixing in the surface layer, since high correlation is found between dissipation of turbulent kinetic energy ($\varepsilon$) and wind stress (corr = 0.66). Whereas in the thermocline layer and deep layer the vertical mixing can caused by the topography of the water.
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