Transformation and mixing of North Pacific Water Mass in Sangihe-Talaud in August 2019

I Y Sani*1, A S Atmadipoera1, A Purwandana2 and F Syamsudin3

1 Department of Marine Science and Technology, Faculty of Fisheries and Marine Science, IPB University, Dramaga, Bogor 16680, West Java, Indonesia
2 Research Center for Oceanography, National Research and Innovation Agency (BRIN), Jl. Pasir Putih 1, Ancol Timur, North Jakarta 14430, Indonesia
3 Faculty of Fisheries and Marine Sciences, Padjadjaran University, Jatinangor, West Jawa, Indonesia

*E-mail: yutika_intan@apps.ipb.ac.id

Abstract. Along the pathway, ITF water is considered to be transformed due to strong diapycnal mixing. This study aims to describe the structure of ITF water and to estimate turbulent mixing. The number of 6 CTD casts and 9 repeated CTD "yoyo" measurements were obtained from the "Years of Maritime Continent" YMC cruise (a joint cruise between BPPT/IPB/UNUD-Univ.Tokyo/JAMSTEC) and onboard R.V. Baruna Jaya IV in August 2019. The CTD datasets are processed with SBE Data Processing and analyzed for water mass composition, as well as turbulent mixing with Thorpe method. The results showed that thermocline water of NPSW with S-max, and intermediate water of NPIW with S-min from North Pacific origin are dominant. Transformation of NPSW and NPIW along their pathway can be identified from decreasing S-max of NPSW and increasing S-min of NPIW. Estimates of $\epsilon$ and $K_\rho$ are $O(10^{-5})$ m$^2$s$^{-2}$ and $10^{-2}$ m$^2$s$^{-1}$, respectively. High mixing occur also in the interior layer with the $\epsilon$ and the $K_\rho$ $O(10^{-6})$ m$^2$s$^{-2}$ and $O(10^{-1})$ m$^2$s$^{-1}$, respectively. This is related to barotropic tidal activity that interacts with the bottom topography where there are many sills, causing the formation of strong baroclinic tides.

Keywords: Indonesian Throughflow, Thorpe method, transformation, turbulent mixing, water masses, western route

1. Introduction

The Indonesian Throughflow (ITF) is the lower latitude of the warm water mass from the Pacific to the Indian Ocean through Indonesian waters [1]. ITF transport plays an important role in the thermohaline circulation of the Pacific and Indian Oceans and even globally [2]. This flow of water mass is caused by the difference in pressure from the two oceans. According to [3], the southeast trade wind blows over the Pacific Ocean throughout the year which causes a buildup of water masses in the western part of the Pacific which is close to Indonesia. As a result, there is a difference in sea level between the western Pacific Ocean and the Indian Ocean in southern Indonesia. This difference in sea level then causes a pressure gradient between the two oceans and causes a flow of water mass from the Pacific Ocean to the Indian Ocean.

Sangihe-Talaud and Makassar Strait are traversed by ITF water masses via the western route. Sangihe-Talaud is an area that has a lot of sills which has a very important effect on deep sea mixing process. In the study of [4], ITF water masses are mixed by tidal energy factors and strong internal
waves. The mixing of the Pacific water mass with freshwater causes a decrease in maximum salinity and finally the outgoing ITF water mass is characterized by a fresher and cooler thermocline. Sulawesi Sea which is close to Sangihe-Talaud has a characteristic high salinity, but its maximum subsurface salinity gradually decreases along the Makassar Strait and disappears before entering the Banda Sea. Then the subsurface temperature is cooler and the thermocline depth is much shallower in the southern Makassar Strait than in the northern part.

There are two types of water masses that enter Indonesian waters through ITF, namely water masses from the North Pacific and from the South Pacific. North Pacific water masses enter through the Mindanao Strait and South Pacific water masses enter through the Halmahera Sea [5]. The dominant mass of water entering Indonesian waters is from the North Pacific, which is around 90% [6]. The North Pacific water mass is divided into two types, namely North Pacific Subtropical Water (NPSW) with maximum salinity characteristics in the thermocline layer and North Pacific Intermediate Water (NPIW) with minimum salinity characteristics in the middle layer [7]. With the mixing, the water mass will experience a change in characteristics. This study aims to identify the characteristics of the water mass and quantify the value of turbulent mixing using the Thorpe method in Sangihe-Talaud in August 2019.

2. Materials and Methods
Data collection activities were carried out in the Sangihe-Talaud and the Makassar Strait in August 2019 with the Baruna Jaya IV vessel at 6 snapshot measurement stations and CTD-yoyo for 9 repetitions for 24 hours at one CTD station in Sangihe-Talaud. The purpose of the CTD-yoyo measurement is to observe the variation of mixing during a tidal period and snapshot station is to see the transformation of water mass characteristics. The study area can be seen in Figure 1.

![Figure 1. Study area of Sangihe-Talaud and Makassar Strait.](image)

2.1 Procedure
CTD data analysis went through 2 processing stages, namely through the pre-processing stage and the processing stage. Pre-processing is the conversion of CTD data with SBE Data Processing and manual correction with Ms. Excel, which then goes through the processing stage with Ocean Data View and Matlab software. The flow chart can be seen in Figure 2.
Figure 2. Flow chart of CTD data processing.

In addition, manual correction is also necessary because it is possible for errors to occur after the previous process. After the pre-processing stage, the converted and corrected data is then processed at the processing stage using Ocean Data View and Matlab software. Processing with Ocean Data View software produces vertical and transverse plots of temperature, salinity, density and Brunt-Vaisala frequency vertical plots for stratification and stability analysis of water masses as well as making T-S diagrams for water mass density analysis. The processing using Matlab software produces Thorpe scale values, turbulent kinetic energy dissipation, and Eddy vertical diffusivity for the analysis of the strength of water mass mixing in the study area.

2.2 Data Analysis

Determination of the water mass layer using a vertical temperature gradient referring to [8] of 0.05 °C/m as the upper and lower limits of the thermocline layer. Stability is indicated by the Brunt Vaisala frequency which can be calculated by the equation [9]:

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

The Thorpe scale is very useful for describing the vertical range of turbulent mixing. The first step to determine the water mass mixing is to determine the Thorpe displacement value \( d_T \). This is done for identification in order to know the existence of overturn. The \( d_T \) value is obtained by rearranging the density profile obtained from the CTD data into the form of static stability, the density is arranged with the position of the low density water mass being above the high density water mass. The displacement distance to the depth of the static stability condition is the value of \( d_T \). Suppose a vertical density profile with \( n \) samples and a density \( n \) observed at depth. If the sample at depth is moved to depth to form a stable condition, then \( d_T \) can be calculated by the equation [10]:

\[ d_T = z_a - z_b \]  \hspace{1cm} (2)

\( z_a \) = sample depth in real condition (m)
\( z_b \) = sample depth in stable condition (m)

The vertical resolution, density resolution [11] to detect reversal are not lower than:

\[ L_z = 5 \, \partial Z \]  \hspace{1cm} (3)
\[ L_\rho \approx 2 \frac{g \delta \rho}{N^2 \rho_0} \] (4)

\[ \partial Z = \text{vertical resolution of CTD data (m)} \]

\[ \delta \rho = \text{CTD ability to detect density difference (~10^{-3})} \]

[11] characterize the tightness of this relationship with the parameter \( \xi = \max(\xi_T, \xi_S) \); where \( \xi_T \) and \( \xi_S \) is the linear correlation of density-temperature and density-salinity, respectively. The value near zero indicates a tight T–S relationship and large values signify a loose relationship, and it was suggested an upper bound of 0.5 for valid overturns. This threshold is quite sensitive, and may potentially discard many detected overturns [12] hence subject to adaptation. Following [13], we choose a slightly higher threshold of 0.66. [14] propose an overturn ratio based on the number of positive and negative displacements to reject false overturns, with a threshold 0.2.

The Thorpe scale value (\( L_T \)) was then calculated using the equation [10]:

\[ L_T = \left( \frac{1}{n} \sum_{i=1}^{n} d_i^2 \right)^{1/2} \] (5)

n = number of samples within a turbulent patch

\( d_i \) = Thorpe displacement in depth i (m)

The amount of kinetic energy that undergoes the dissipation process can be calculated using the equation [10, 13, 15]:

\[ \varepsilon_{\text{Th}} = 0.64 L_T^2 N^3 \] (6)

\[ \varepsilon_{\text{GM–background}} = \frac{\varepsilon_0 N^2}{N_0^2} \] (7)

\[ \varepsilon_{\text{background}} = \max \left( 1 \times 10^{-10}, \varepsilon_{\text{GM–background}} \right) \] (8)

\( \varepsilon_{\text{Th}} \) = dissipation rate when overturn is observed (m\(^2\)s\(^{-2}\))

\( \varepsilon_0 \) = background dissipation rate (6.7\times10^{-10} m\(^2\)s\(^{-2}\))

\( N_0 \) = reference buoyancy frequency (3 cycl/h)

\( \varepsilon_{\text{GM–background}} \) = background dissipation rate GM (m\(^3\)s\(^{-1}\))

\( \varepsilon_{\text{background}} \) = final background dissipation rate (m\(^3\)s\(^{-1}\))

The vertical turbulence value is calculated by calculating the eddy diffusivity (\( K_\rho \)) with the equation [13, 16]:

\[ K_\rho \approx K_{\rho \text{Th–GM}} = \frac{\gamma \varepsilon_{\text{Th–GM}}}{N^2} \] (9)

\( \gamma \) = mixing efficiency (0.2) [17]

\( K_{\rho \text{Th–GM}} \) = vertical eddy diffusivity (m\(^2\)s\(^{-1}\))

The calculation process to obtain these values is carried out by several processes which are presented in the plot in Figure 3. Part (a) shows the vertical profile of the potential anomaly density which is then calculated \( d_T \) based on static stability conditions (low density is above high density). Then in parts (c) and (d) each is a constant of the relationship between density with temperature and salinity where then both are calculated the resultant in part (e) to select noise. The results of \( d_T \) (f) are re-selection based on the overturn ratio (g).
3. Results and Discussion

The vertical profile of temperature, salinity, and density of Sangihe-Talaud is shown in Figure 4. Based on the determination of the temperature gradient ≥0.05 °C/m, the depth of the upper limit of the thermocline layer bordering the mixed layer and the lower limit bordering the deep layer can be determined. The temperature and salinity range in the mixed layer in Sangihe-Talaud is 26.86 – 28.52 °C and 34.36 – 34.54 psu, respectively. Based on previous research, the surface temperature value in Sangihe Talaud ranges from 30 °C [18] and the surface salinity value in Sangihe-Talaud has a range of 33.6 – 34.4 psu [19]. The lower temperature and salinity range than before can be caused by the weak El-Nino phenomenon that occurred in the dry season of 2019, where the surface temperature of the waters tends to be colder than usual. The thermocline layer lies below the mixed layer and is the layer where the temperature drops drastically with depth. The upper and lower limits of the thermocline layer in the Sangihe-Talaud range from 54 – 76 m and 246 – 311 m, respectively, with a temperature change of 27.41 – 9.83 °C. This thermocline layer is slightly increased compared to the results of research by [18], where the range of the thermocline layer is around 75 – 350 m.

The mixed and thermocline layer (the boundary of the two layers has a temperature gradient ≥0.05 °C) on the CTD-yoyo have an average thickness of 58.11 ± 8.54 m and 220.44 ± 27.60 m, respectively. The thickness of the mixed layer is the largest in repetition 5 (03-05) and the smallest is
in repetition 1 (03-01). These differences in the thickness of mixed layer indicate the presence of internal tides (baroclinic), which is the layer will become thinner when the wave crest passes through the water. When the crest of the internal tides passes through the water column, the mixed layer will be compressed and will become thinner, but the mixed layer will become thicker when passed by the internal tides trough. The thinning of the mixed layer is also accompanied by an increase in the upper limit of the thermocline layer. Changes in the thickness and range of temperature and salinity values in the mixed layer and the thermocline can be caused by physical factors, one of which is current. According to [20], several factors also affect the stratification of the water mass, including current generation energy, as well as upwelling and downwelling.

Figure 5. Water mass stratification in Sangihe-Talaud (CTD-yoyo).

The transverse depth-time profile for salinity and density parameters of Sangihe Talaud is shown in Figure 6. Based on the salinity profile, it can be clearly seen the depth with maximum and minimum salinity. The maximum salinity ($S_{\text{max}}$) is in the upper layer of the thermocline and the minimum salinity ($S_{\text{min}}$) is in the intermediate layer. The transverse plot of the density data for 24 hours indicated the presence of internal tides propagation. Internal tides detected in Sangihe-Talaud tend to form a semidiurnal pattern. This can be seen through the density profile where there are 2 peaks at isopycnal 22.5 kg/m$^3$ and 2 valleys at isopycnal 25.5 kg/m$^3$. The peak of the tide tends to be on repetitions 3 and 6 while the low tide tends to be on repetitions 1 and 5. The change in the isopycnal line according to the pattern of the internal tides indicates that the effect of internal tides is very strong on the vertical structure of the water column [21]. At high tide, the isopycnal line of the water mass in the lower layer will be pushed (induced) to a lower pressure or shallower depth, but on the contrary at low tide the isopycnal line the water mass will be pushed to a deeper layer. An example of the ebb effect (valley) is most clearly seen in repetitions 1 and 5 where the effect of the internal tides reaches near the bottom of the water, i.e. about 1175 m depth on repetition 1 and about 1025 m depth on repetition 5. The change in the isopycnal line according to the pattern of the internal tides indicates that the effect of the internal tides is very strong on the vertical structure of the water column. [22] explain that internal tides activity in a water column will greatly affect the vertical profile (temperature, salinity, and density) of the waters. This is in line with what is found in the vertical profile of temperature, salinity, and density of the water column where when the internal tides reaches the bottom, the vertical profile of temperature and salinity becomes irregular (Figure 4) and causes instability in the water column in the deep layer (Figure 8).
The water mass of ITF passing through the waters of Sangihe Talaud and Makassar Strait can be seen through the T-S diagram in Figure 7. According to [6], 90% of the dominant water mass entering Indonesian waters comes from the North Pacific. Based on the T-S diagram analysis, the structure of the North Pacific water mass in Sangihe Talaud and the Makassar Strait is dominated by NPSW and NPIW water masses. According to [8], NPSW is characterized by a maximum salinity of 34.8 – 35.2 psu and a temperature of 20 – 24 °C, while the NPIW is characterized by a minimum salinity of 34.1 – 34.5 psu and a temperature of 7 – 11°C. The water mass structure of NPSW was found in Sangihe Talaud in the upper thermocline layer, namely at a depth of 68 – 176 m, and in Makassar Strait was found at a depth of 70 – 126 m. Furthermore, the mass of NPIW water was found in Sangihe Talaud in the middle layer at a depth of 235 – 453 m and in the Makassar Strait was found at a depth of 201 – 445 m. The decreasing maximum salinity value and the increasing minimum salinity from Sangihe Talaud to the Makassar Strait indicate that there is a mass transformation of ITF water due to the mixing process.
Table 1. Water masses characteristic in Sangihe-Talaud and Makassar Strait.

| Parameters                              | Sangihe-Talaud | Makassar Strait |
|-----------------------------------------|----------------|-----------------|
|                                         | NPSW | NPIW | NPSW | NPIW |
| Potential Temperature (°C)              | 20 – 24 | 7 – 11 | 17.5 – 22.0 | 8 – 13 |
| Salinity (PSU)                          | 34.8 – 35.2 | 35.0 – 35.6 | 34.5 – 34.7 | 34.4 – 34.6 |
| Potential Density Anomaly (kg/m³)       | 24 | 24 | 26.5 | 26.5 |

Frekuensi Brunt Vaisala (N) menunjukkan kestabilan lapisan massa air yang dipengaruhi oleh stratifikasi air sehinga menimbulkan dinamika turbulen[23]. A fluid is said to be stable when there is a large gradient in temperature and salinity in the water. If the gradient tends to be small, then the energy required for the fluid to move is not large, so the fluid is neutral. If there is no stratification, the fluid becomes unstable so that the vertical fluid displacement is further away from the initial position. The frequency value close to zero (N≈0) does not represent loss of stratification in total, but the stratification weakened due to mixing [24]. The N profile is affected by the vertical profile of temperature and potential density (σ₀). The results of the vertical N profile in each water layer are presented in Figure 8. The N values in the mixed, thermocline, and deep/homogeneous layers in Sangihe-Talaud ranged from -8.24 – 14.23 cycl/h, -3.99 – 24.84 cycl/h, and -8.43 – 12.23 cycl/h, respectively.

![Figure 8](image-url)
The thermocline layer has the highest $N$ value compared to the mixed layer and deep layer. This shows that the thermocline layer is the most stable wherein [25]. It is explained that in the thermocline layer the density value increases sharply with depth so that the density gradient gets bigger and causes an increase in the static stability of the waters. Meanwhile, the low $N$ in the mixed layer, especially in the deep layer where many negative values were found, indicated that the water conditions tended to be statically unstable.

In the upper thermocline layer in repetition 5 (50–200 m) and deep layer in repetition 6 (800-1000 m), it is seen that the water mass tends to be more unstable than the other repetitions. This is thought to be due to a very strong internal tides activity in repetition 5 to the bottom of the waters of repetition 6 (there is a time lag) resulting in a mixing process.

The strength of turbulent mixing was assessed by estimating the value of the kinetic energy dissipation rate ($\epsilon$) and the vertical eddy diffusivity ($K_\rho$). Epsilon ($\epsilon$) indicates the value of the dissipated energy released when it occurs turbulence and $K_\rho$ indicate the diffusivity of water mass for transfers energy when there is oscillation. The mixing value in Sangihe-Talaud varies in each layer which can be seen from figure 9, which shows the 25 bin averaged mean value. The interior layer has higher $\epsilon$ and $K_\rho$ values than the mixed layer and the thermocline, reaching $5.54 \times 10^{-6}$ m$^2$ s$^{-2}$ and $1.92 \times 10^{-1}$ m$^2$ s$^{-1}$, respectively. The high value in this layer is strongly suspected to be influenced by barotropic activity that interacts with the ocean topography and the effect of having a sill at the bottom of the water. The existence of sill in Sangihe-Talaud area allows the process of dissipation (energy release), high dissipation in the threshold area has an important effect on the process of mixing water masses where the topography interacts strongly with barotropic tidal currents. Rough topographical features such as sills, when fed by barotropic tides will experience local mixing and some will become baroclinic energy [26].

![Figure 9](image-url).

Figure 9. Comparison of 25 bin averaged mean from CTD-yoyo (a) dissipation rates, (b) vertical eddy diffusivity.

The high displacement near the bottom of the water that occurs in several repetitions indicates strong mixing as found in repetition 2 (925 – 1025 m) and in repetition 5 (950 – 1025 m). From the $K_\rho$
values in all repetitions, it can be seen that the $K_p$ value in repetition 2 and 5 was higher than the other repetition while the lowest value was obtained in repetition 1 and 7. The time lag of the tidal effect in the first valley causes a receding effect on the internal tides of repetition 5, which clearly affects the deep layer of repetition 2. Otherwise, it could also be due to strong mixing near the bottom due to sill (Figure 10). The displacement, Thorpe scale, and the kinetic energy of turbulent eddy dissipation, which later all of these values will affect the high and low value of $K_p$ in each layer of the water column. It is strongly suspected that turbulent mixing also affects the transformation of water masses along ITF's western trajectory, from Sangihe-Talaud to the Makassar Strait. This can be seen from the changes in the maximum and minimum salinity values based on the vertical profile (Figure 4), with the dissipation value ($\epsilon$) reaching $O(10^{-5})$ m$^2$ s$^{-2}$ and the vertical diffusivity value ($K_p$) reaching $O(10^{-2})$ m$^2$ s$^{-1}$. This is in accordance with [27], where the water masses of NPIW and NPSW experience changes in the maximum and minimum salinity values up to 0.3 psu.

![Figure 10. 2D and 3D bathymetry appearances around the data collection point indicating the presence of rough topography such as sill.](image)

### 4. Conclusion

Sangihe-Talaud and the Makassar Strait are dominated by ITF water masses from the North Pacific, namely North Pacific Subtropical Water (NPSW) and North Pacific Intermediate Water (NPIW). The values of $S_{\text{max}}$ and $S_{\text{min}}$ changed from Sangihe-Talaud to the Makassar Strait, which was evidenced by the high value of turbulent mixing in the layer flowing by the water mass. Turbulent mixing occurs most strongly in the interior layer, which tends to be homogeneous. The existence of tidal activity (baroclinic) is highly suspected as the cause of the high value of turbulent mixing in the deep sea, and especially near the bottom close to the topography in the form of sills around 1175 m depth with the maximum value is $O(10^{-1})$ m s$^{-1}$.

### Acknowledgments

Thanks to Prof. Hibiya and Dr. Nagai from Tokyo University, to the Captain and crew of Baruna Jaya IV, and to the YMC Cruise research team, who have played a major role in the data collection process. This project is also funded by the research scheme of Program Riset Unggulan COREMAP CTI 2021-2022 (17/A/DK/2021).

### References

[1] Gordon A L 2005 Oceanography of the Indonesian seas and their throughflow *J. Oceanogr.* **18**(4) 14-27
[2] Gordon A L 1986 Interocean exchange of thermocline water J. Geophys. Res. Atmos. 91(C4) 5037-46
[3] Hasanudin M 1998 Arus lintas Indonesia Jurnal Oseana 23(2) 1-9
[4] Fan W, Jian Z, Chu Z, Dang H, Wang Y, Bassinot F, Han X and Bian Y 2018 Variability of the Indonesian in the Makassar Strait over the last 30 ka Sci. Rep. 8 1-8
[5] Feng M, Zhang N, Liu Q and Wijffels S 2018 The Indonesian throughflow, its variability and centennial chang Geosci. Lett. 5(3) 1-10
[6] Atmadipoera A, Molcard R, Madec G, Wijffels S, Sprintall J, Koch-Larrouy A, Jaya I and Supangat 2009 Characteristics and variability of the Indonesian throughflow water at outflow strait Deep Sea Res I. 56 1942-54
[7] Horhoruw S M, Atmadipoera A S, Purba M and Purwandana A 2015 Current structure and spatial variation of Indonesian Throughflow in Makassar Strait under Ewin 2013 IJMS 20(2) 87-100
[8] Wyrtki K 1961 Scientific Results of Marine Investigations of the South China Sea and the Gulf of Thailand vol 2 Naga Rep. (California: University of California)
[9] Stewart R H 2002 Introduction to Physical Oceanography (Texas: A&M University)
[10] Dillon T M 1982 Vertical overturns: a comparation of Thorpe and Ozmidov length scale J. Geophys. Res. 87 9601-13
[11] Galbraith P S and Kelley E 1996 Identifying overturn in CTD profiles J. Atmos. Ocean. Tech. 13 688-702
[12] Stansfield K, Garrett C, Dewey R 2001 The probability distribution of the Thorpe displacement within overturns in Juan de Fuca Strait J. Phys. Oceanogr. 31 3421-34.
[13] Purwandana A, Cuypers Y, Bouruet-Aubertot P, Nagai T, Hibiya T and Atmadipoera A S 2020 Spatial structure of turbulent mixing inferred from historical CTD datasets in the Indonesian seas Prog.Oceanogr. 184 1-20
[14] Gargett A and Garner T 2008 Determining Thorpe scales from ship-lowered CTD density profiles J. Atmos. Oceanic. Technol. 25 1657-70
[15] Garrett C and Munk W 1975 Space-time scales of internal waves: a progress report J. Geophys. Res. 80 291–7
[16] Park Y H, Fuda J L, Durand I and Garabato A C N 2008 Internal tides and vertical mixing over the Kerguelen Plateau Deep-Sea Res II. 55 582-93
[17] Osborn T R 1980 Estimates of the local rate of vertical diffusion from dissipation measurement J. Phys. Oceanogr. 10 83-9
[18] Radjawane I M and Hadipoerwanto P P 2014 Water masses characteristics at the Sanghie Talaud entry passage of Indonesian Throughflow using index Satal data 2010 JIITKT 6(2) 525-36
[19] Ismail M F A and Taofiqurohman A 2020 The spatial distribution of temperature, salinity and density in Sangihe Talaud waters North Sulawesi Jurnal Kelautan Tropis 23(2) 191-8 (in Bahasa Indonesia)
[20] Tomczak M 2000 Education: an introduction to online physical oceanography J Oceanogr. 13(2) 104-5
[21] Suteja Y 2011 Turbulent Mixing Caused by Internal Tide and Their Implication on Nutrient in Ombai Strait Master Thesis IPB University, Bogor
[22] Prasad K V S R and Rajasekhar M 2006 Observations of oceanic internal waves in Bay of Bengal using Synthetic Aperture Radar Proc.SEASAR. (Frascati: unknown)
[23] Purwandana A 2012 Transformation and Mixing of Water Mass in Alor Strait in July 2011 Master Thesis IPB University, Bogor
[24] Roseli N H, Akhir M F, Husain M L, Tangang F and Ali A 2015 Water mass characteristics and stratification at the shallow Sunda shelf of Southern South China Sea OJMS. 5 455-67
[25] Pond S and Pickard G L 1983 Introductory Dynamical Oceanography 2nd edition (Toronto: Pergamon Press)
[26] Hermansyah H, Nugroho D, Atmadipoera A S, Pratono T, Jaya I and Syamsudin F 2018 Tidal barotropic and baroclinic kinetic energy dissipation in the Sulawesi Sea *JITKT* 10(2) 365-80 (in Bahasa Indonesia)

[27] Naulita Y 2016 Study on turbulent mixing processes in Labani Channel, the Makassar Strait *JITKT* 8(1) 345-55 (in Bahasa Indonesia)