Analysis of turbulent mixing in Dewakang Sill, Southern Makassar Strait

Risko1*, A S Atmadipoera2, I Jaya2 and E H Sudjono3
1 Graduate Program in Marine Science, Bogor Agricultural University, Indonesia
2 Marine Science and Technology Dept. Bogor Agricultural University, Indonesia
3 Marine Geological Institute, Bandung Indonesia

Email: risko.fisika4@gmail.com

Abstract. Dewakang Sill is located in southern Makassar Strait, conveying major path of Indonesian Throughflow (ITF), as a confluence region of different water masses, such as salty Pacific water and fresh Java Sea water. Its depth is about 680 m which blocks the ITF flow below this depth into Flores Sea. This research aimed to estimate turbulent mixing in the Dewakang Sill by applying Thorpe analysis using 24 hours "yoyo" CTD data sets, acquired from MAJAFLOX Cruise in August 2015. The results showed that stratification of water masses is dominated by Pacific water origin. Those are North Pacific Subtropical thermocline and Intermediate water masses. Mean dissipation of turbulent kinetic energy ($\epsilon$) and turbulent vertical diffusivity ($K_{\rho}$) value in the Dewakang Sill are of $O(1.08 \times 10^{-6})$ Wkg$^{-1}$, and $O(2.84 \times 10^{-4})$ m$^2$s$^{-1}$ respectively. High correlation between epsilon and internal waves oscillation suggested that internal tidal waves activities are the major forcing for turbulent mixing in the study area.

1. Introduction
Makassar Strait conveys major Pacific-Indian transfer (ITF) with mean transport volume of about 9-11 Sv ($1 \text{ Sv} = 10^8 \text{ m}^3\text{s}^{-1}$) [1, 2]. In southern part of Makassar Strait, there exists the Dewakang sill waters which is located in eastern channel of Southern Makassar Strait. This sill functions as a barrier of Indonesian Throughflow (ITF) water mass below the sill depth of about 680 m [3]. However, in the western channel, part of the ITF flows into Lombok Strait and the rest continues flowing into Flores Sea [2, 4, 5].

Theoretically, an unstable stratification of water column leads fluids in the mixing process, and they are grouped into two parts i.e. static stability and double-diffusion [6]. This shows that the static stability density changes with depth while the dynamic stability changes with velocity shear. Furthermore, this double-diffusion is related to the salinity gradient and seawater temperature. Water mass movement which is caused by the variation of turbulent flow can form a fluid mixing with a very high fluctuation, and the process of coating or stratification of the water mass is influenced by differences in ocean temperature, salinity and density. The process of water mass stratification is caused by density differences which can create the mixing. This, according to Ffield [7], causes a change in the amount of heat, salinity, and water mass momentum.

Field and Gordon [8], estimated the value of the mixing in the thermocline layer in the Indonesia seas, and the value of $1.0 \times 10^{-4}$ m$^2$s$^{-1}$ was obtained. The value of the mixing coefficient is nearly the same as the result obtained by Koch-Larrouy [9] in which she conducted the research in the waters...
of Indonesia, i.e. $1.5 \times 10^{-4}$ m$^2$s$^{-1}$. Furthermore, the same research was also conducted by Hatayama [10] using a numerical modeling that produced a vertical diffusivity value in Dewakang sill, i.e. $6.0 \times 10^{-3}$ m$^2$s$^{-1}$. An approach with an indirect measurement was also carried out by Suteja [11] in Ombai Straits and Purwandana [12] in Alor Straits, where the averaged turbulent mixing values obtained were $7.56 \times 10^{-3}$ m$^2$s$^{-1}$ and $1.0 \times 10^{-3}$ m$^2$s$^{-1}$, respectively.

Some of these approaches produced various turbulent mixing values. Therefore, it is necessary to apply another approach for calculating the turbulent mixing in Dewakang Sill that is by using Thorpe analysis method based on the vertical profile obtained from the hydrographic data (CTD). This research was carried out by field experiment in Dewakang Sill, southern Makassar Strait where the location is part of the Arlindo region, known to have strong internal tides and energetic mixing region [9, 13].

Turbulent mixing that occurs in Indonesia seas is primarily caused by the internal tidal, and one way to determine the internal tidal is by measuring the CTD data for 24 hours (one tidal period) or commonly called yoyo CTD [14, 15]. The aims of this study are to analyze stratification and characteristics of the Indonesian throughflow water masses; to estimate of turbulent mixing of water masses, and to investigate physical process influencing turbulent mixing, such as surface wind stress, internal tides and sea topographic slope in Dewakang Sill, that contributes to the transformation of water masses.

2. Method

2.1. Data acquisition

This research was conducted in Dewakang Sill, South Makassar Strait (figure 1). The study on turbulent mixing analysis is part of the Makassar - Java - Flores (MAJAFLOX) cruise which was held from August 4 to August 20, 2015 by using the Research Vessel of Geomarin III of the Marine Geological Institute (P3GL), Balitbang ESDM Bandung.

![Figure 1. Bathymetry in the triangle seas of Java Sea – Makassar Strait – Flores Sea. Green dot denotes location of repeated CTD “yoyo” measurement for 24-hours in the Dewakang Sill Strait from August 10 to 11, 2015.](image)

MAJAFLOX expedition activity is part of a research collaboration between the Department of Marine Science and Technology, Faculty of Fisheries and Marine Science of IPB and P3GL-ESDM.
The data of temperature, salinity and density were obtained by using the SBE (Sea Bird Electronic) CTD 19plus Version 2 for 11 casts repetition during 24-hours. The results of the data measurement were recorded on deck unit in the form of an analog signal which is then converted by CTD probe into a digital signal by connecting it directly to a computer with a data cable. The determination of the averaged depth of each parameter is 0.5 m obtained by using the software of SBE Data Processing 7.21e. Accuracy and resolution of the temperature sensor is 0.005°C and 0.0001°C, respectively, and for conductivity sensor is 0.0005 S/m and 0.00005 S/m, respectively [16].

Winds field data were obtained from the ECMWF (European Centre for Medium-Range Forecast) through the site of (www.ecmwf.int). The data are the result of repeated analysis and meteorological data interpolation which were obtained from various centers of observations and research on the world meteorological stations. The data used in this study include the wind vectors obtained from 10 to 11 August 2015 consisting of zonal \((u)\) and meridional \((v)\) components 10 m above sea level. Tidal data in Dewakang Sill from August 10 - 11, 2015 were obtained from website of Indonesian Agency for Geospatial Information (www.tide.big.id) at the point of the study area. The data which include the observations, predictions and model of the sea from the control link center of geodesy and geometrics are taken at the point of the research station.

### 2.2. Data analysis

The water mass stratification in the water column can be seen from the Brunt-Vaisala frequency \((N^2)\) value. This is influenced by the gravitational force \((g)\), which will cause fluids with a greater density to move downward and those with the smaller density to move upward. The calculation method of vertical mixing of water masses from the inner layer of the waters is a preliminary analysis used by calculating the Brunt-Vaisala frequency using the equation \([17, 18]\)

\[
N^2 = -\frac{g}{\rho_0} \frac{d\rho}{dz}.
\]

According to Ferron [19] the density value used in the Brunt-Vaisala frequency is derived from the density data compiled in a stable condition, where \(\rho_0\) shows the average water density of the whole CTD data repetition (1.026.52 kg m\(^{-3}\)), \(d\rho\) is the density change (gradient) density toward the depth change (1 m), and \(g\) is the earth gravitational acceleration (9.79423 m s\(^{-2}\)).

![Figure 2. The schematic Thorpe displacement \((d)\) calculation. The initial density profile (blue line) are reordered to seek for the density of the static stability condition (red dotted line). Distance from \(z_a\) to the depth of \(z_b\) is stated as a Thorpe displacement \((d)\) value](image)

The estimated value dissipation of turbulent kinetic energy \((K_p)\) was conducted using the Thorpe scale analysis. Thorpe is determined from the density repetitions in the form of static stability in
accordance with the initial density or depth (Figure 2). Thorpe displacement (d) can be calculated by the equation (2).

\[ d = z_a - z_b \]  

(2)

Where \( z_a \) and \( z_b \) are the initial pressure position and pressure after the repetitions [18, 20, 21]. The \( d \) positive value indicates that the water mass will move upward to search for the static stability in which this occurs if the water mass which has a low density is under the water mass with high density. Meanwhile, the \( d \) negative value indicates that the water mass moves downward.

After calculating the \( d \) value, the minimum thickness displacement of CTD vertical resolution was estimated. The reason for this is that the \( d \) value becomes the real displacement value and does not come from the CTD noise. The principle of this estimation is based on the fact that CTD has a limited ability to detect inversion of water mass. This refers to the Nyquist sampling theory, where if a inversion occurring is two times lower than the vertical resolution, the inversion cannot be measured. Determination of a stronger reversal can be done if there are adequate samples in which based on the regulation, the number of samples must be at least five samples [22] or 7-8 samples [23].

Furthermore, the determination of the threshold of the reversal value of the data that have been replicated was carried out by using the GK method [23]. Based on this method, a minimum value (5 m) will be ignored and excluded for further calculations. Interval of the vertical depth (\( \partial z \)) of CTD data is 1 meter to support the reversal of water mass threshold i.e. \( L_z = 5\partial z \). The determination of the turbulent kinetic energy dissipation is used to describe the amount of kinetic energy that is lost or its form is changed in the water. The calculation of the amount of kinetic energy experiencing a dissipation process based on the Ozmidov scale (1965) in Part [17] is

\[ \varepsilon = L_o^2 N_i^3 \]  

(3)

where \( L_o \) and \( \varepsilon \) denote the Ozmidov length scale and dissipation of turbulent kinetic energy. Furthermore, the determination of Thorpe scale was obtained by using an equation (4)

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

(4)

where \( d_i \) is the Thorpe displacement at the depth of \( i \) and \( n \) is the number of samples [17, 20, 21, 24]. Each \( L_T \) value was obtained from the average of \( n \) samples in the desired depth. The average \( L_T \) value of this research was obtained by dividing the depth of the waters into three layers including the mixed layer, thermocline layer, and homogeneous layer of the inner part. The depth of each layer varies depending on the vertical profile of water masses.

Thorpe scale was used in determining of the value of Ozmidov length scale at each layer and the calculation was carried by using an equation [20] (5)

\[ L_o = 0.8 \times L_T \]  

(5)

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

\[ K_{\rho i} = \frac{\Gamma \varepsilon_i}{N_i^2} \]  

(6)

where \( \Gamma \) states the mixing efficiency (0.2) [17, 24, 25].

The effects of wind stress on the vertical turbulent mixing in the mixed layer ware considered by calculating the wind stress at sea level with equation (7)

\[ \tau_x = \rho_{air} C_D (U_{10})^2 \]  

(7)

where \( \rho_{air} \) (the density of air) = 1.3 kgm\(^{-3}\); \( C_D = 1.5 \times 10 \) ms\(^{-1}\); \( U \) = wind speed [26]. The relationship between the vertical turbulence and windstress was obtained by calculating their linear correlation coefficient.

3. Results and discussion

3.1. Water mass static stability

Static stability of the water masses is one of the basic variables in oceanography which define the increasing density with descending depth. Instability in the water column can lead to water mass
mixing that can be identified by calculating the Brunt Vaisala frequency values ($N^2$). The calculation results of Brunt Vaisala frequency from the CTD yoyo deployment showed a relatively large value of $N^2$ was found in the thermocline layer ($6.39 \times 10^{-5} - 1.58 \times 10^{-4}$) cycl/s at depth of 100-200 m. Figure 3 shows a vertical profile of $N^2$ in Dewakang Sill waters. The similar profile to $N^2$ in this region was obtained by Park [17] and Suteja [11] in their research sites of Kerguelen Plateau and the Ombai Strait respectively. The high value of $N^2$ in the thermocline layer occurred because of the existence of the pycnocline layer that is a layer where the density gradient increases significantly with depth (pressure) [26]. The higher value of $N^2$ in a particular ocean layer, the larger the static stability of the layer. On the contrary, if the value of $N^2$ becomes more negative, the water column becomes more unstable, or in other words, it is in a state of static instability. In reference to all of the $N^2$ repetitions, it can be seen that most of them are relatively negative. This suggests that a highly instable water column condition is presumably linked to a high current condition as well as an interaction with the sill in the region of Dewakang Sill which triggers a turbulence to occur.

![Figure 3. Brunt Vaisala frequency ($N^2$) profiles from 11 CTD repeated measurement in the Dewakang Sill.](image-url)
3.2. Analysis of vertical turbulent mixing

3.2.1. Dissipation of turbulent kinetic energy

The estimated average value dissipation of turbulent kinetic energy ($\varepsilon$) obtained from Dewakang Sill waters at all depths is $1.08 \times 10^{-6}$ W kg$^{-1}$. This result is similar to that obtained by Suteja [11] in Ombai Strait i.e. $10^{-6}$ W kg$^{-1}$, however, it is larger than the outcome obtained by Purwandana [12] in Alor Strait i.e. around $O(1.08 \times 10^{-5})$ W kg$^{-1}$ the Southeast monsoon. The average value dissipation of turbulent kinetic energy ($\varepsilon$) at the average of every 50 m depth in Dewakang Sill is presented in Table 1. It can be seen that the bottom layer has the smallest value in comparison with the layers of turbulent mixing and thermocline. The low value of $\varepsilon$ in deeper layer is suspected to result in smaller kinetic energy from the turbulent flow, which will break into another smaller form (dissipation) to transfer energy to another media.

**Table 1.** Tabulation of average values dissipation of turbulent kinetic energy ($\varepsilon$) on every 50 meter-depth in Dewakang Sill as resulted from 11 CTD profiles

| Depth (m) | Dissipation of Turbulent Kinetic Energy ($\varepsilon$) (W kg$^{-1}$) | Standard Deviation |
|-----------|-------------------------------------------------|--------------------|
| 25.8      | $9.6 \times 10^{-7}$                            | $\pm 13.58 \times 10^{-7}$ |
| 76        | $30.89 \times 10^{-7}$                          | $\pm 67.74 \times 10^{-7}$ |
| 125.2     | $28.74 \times 10^{-7}$                          | $\pm 31.82 \times 10^{-7}$ |
| 175.7     | $22.51 \times 10^{-7}$                          | $\pm 28.11 \times 10^{-7}$ |
| 225.5     | $8.21 \times 10^{-7}$                           | $\pm 9.48 \times 10^{-7}$ |
| 273.8     | $4.95 \times 10^{-7}$                           | $\pm 11.08 \times 10^{-7}$ |
| 322.9     | $4.27 \times 10^{-7}$                           | $\pm 4.67 \times 10^{-7}$ |
| 378.4     | $1.68 \times 10^{-7}$                           | $\pm 2.34 \times 10^{-7}$ |
| 427.4     | $2.02 \times 10^{-7}$                           | $\pm 3.01 \times 10^{-7}$ |
| 476.6     | $3.63 \times 10^{-7}$                           | $\pm 7.59 \times 10^{-7}$ |
| 522.3     | $1.34 \times 10^{-7}$                           | $\pm 1.88 \times 10^{-7}$ |
| 573.7     | $0.87 \times 10^{-7}$                           | $\pm 0.98 \times 10^{-7}$ |

Dissipation of turbulent kinetic energy values, which are relatively large, are found in the thermocline layer i.e. $(6.835 \times 10^{-7} - 4.37 \times 10^{-6})$ W kg$^{-1}$. This indicates that the thermocline layer in Dewakang Sill waters is the layer where the kinetic energy experiences the highest breakage presumably happens due to the sill functioning as the barrier which causes the mixing process. The level of the values is presumably related to the internal wave activity and as a result of tidal phases on the location of the data collection, which then, will affect the range of values of displacement and dissipation of turbulent kinetic energy.

3.2.2. Turbulent vertical diffusivity

In general, the average value of turbulent vertical diffusivity ($K_p$) for the entire CTD measurements is $2.8 \times 10^{-4}$ m$^2$ s$^{-1}$. This is considered to be larger than that from the research conducted by Koch-Larrouy [9] to obtain the average value of $1.5 \times 10^{-4}$ m$^2$ s$^{-1}$ for the Indonesian waters and nearly the same as the research result obtained by Purwandana [12] in Alor Strait i.e. $O(10^{-4} - 10^{-3})$ m$^2$ s$^{-1}$

More intensively powerful turbulent mixing is found in the mixed layer surface which is presumably caused by relatively high wind stress during the field experiment (the Southeast monsoon), as shown in Figure 4. This is marked by the considerably high correlation value between the dissipation of turbulent energy and wind stress (corr = 0.67). Therefore, it has more energy for the vertical turbulent to occur in the mixed layer. This reflects the major contribution of atmospheric cooling and air
pressure to push the turbulent mixing into the surface layer. Furthermore, stronger vertical turbulent mixing is also found in the thermocline layer and the layers near the bottom waters (Table 2). Turbulent vertical diffusivity \((K_p)\) value in the thermocline layer is higher than that obtained by Ffield and Gordon [8] and Hatayama [10] at Dewakang Sill i.e. \(1.0 \times 10^{-4}\) m\(^2\)s\(^{-1}\) and \(6.0 \times 10^{-4}\) m\(^2\)s\(^{-1}\), respectively. In this study, the range of \(K_p\) value is higher than the results obtained by Matsuno [27] that is equal to \(O(10^{-6} - 10^{-4})\) m\(^2\)s\(^{-1}\) in the East China Sea at a depth of 200 m. The correlation value between dissipation of turbulent kinetic energy \((\varepsilon)\) and internal wave \((A_{int} \times N^2)\) is also reasonably high \((corr=0.68)\). Subsequently, it is suggested that at this layer there are reasonably high influence of internal tidal waves. This condition indicates that Dewakang Sill is a region with reasonably high turbulent mixing. These results are consistent with Koch-Larrouy [9] that Dewakang Sill is an energetic turbulent mixing region.

![Figure 4](image-url)  
**Figure 4.** Snapshot of surface wind vector 10 m above sea level; (a) August 10, 2015 (18:00) (b) August 11, 2105 (00:00); (c) August 11, 2015 (06:00); (d) August 11, 2015 (12:00); (e) August 11, 2015 (18:00) during the data collection of "yoyo" CTD in Dewakang Sill and its adjacent waters
Figure 5. Correlation between the dissipation of turbulent kinetic energy and wind stress in the mixed surface layer (3-80 m) (a); (b) between the dissipation of turbulent kinetic energy and internal waves in the thermocline layer (80-250 m).

Table 2. Tabulation of averaged turbulent vertical diffusivity ($K_p$) on every 50 meter-depth in Dewakang Sill as resulted from 11 CTD profiles.

| Depth (m) | Vertical Eddy Diffusivity ($K_p$) (m$^2$/s) | Standard Deviation |
|-----------|-------------------------------------------|--------------------|
| 25.8      | $4.74 \times 10^{-3}$                      | $\pm 5.99 \times 10^{-3}$ |
| 76        | $4.91 \times 10^{-3}$                      | $\pm 8.63 \times 10^{-3}$ |
| 125.2     | $3.60 \times 10^{-3}$                      | $\pm 4.40 \times 10^{-3}$ |
| 175.7     | $2.96 \times 10^{-3}$                      | $\pm 2.83 \times 10^{-3}$ |
| 225.5     | $2.68 \times 10^{-3}$                      | $\pm 2.94 \times 10^{-3}$ |
| 273.8     | $2.00 \times 10^{-3}$                      | $\pm 2.88 \times 10^{-3}$ |
| 322.9     | $2.64 \times 10^{-3}$                      | $\pm 2.98 \times 10^{-3}$ |
| 378.4     | $1.36 \times 10^{-3}$                      | $\pm 1.56 \times 10^{-3}$ |
| 427.4     | $1.99 \times 10^{-3}$                      | $\pm 2.01 \times 10^{-3}$ |
| 476.6     | $2.91 \times 10^{-3}$                      | $\pm 5.49 \times 10^{-3}$ |
| 522.3     | $2.07 \times 10^{-3}$                      | $\pm 2.39 \times 10^{-3}$ |
| 573.7     | $2.20 \times 10^{-3}$                      | $\pm 3.38 \times 10^{-3}$ |

Furthermore, the deeper layers are most likely related to the internal tidal influence and caused by the influence of the sill in this research area. Moum [28] reported that the sill is able to cause changes in the average flow velocity, modify the vertical gradient velocity, strengthen density gradient, and reduce the flow. In addition, Naulita and Kitade [29] also argued that the slope of the seabed topography corresponding to the tilt direction of propagation of internal tides can cause breaking internal tides, then, resulting mixing in the water column.
4. Conclusion

Dewakang Sill is categorized as an area that has a strong turbulent mixing value. Thus, the highest values dissipation of turbulent kinetic energy ($\varepsilon$) and the turbulent vertical diffusivity were found in the mixed and thermocline layers with respective values of $1.08 \times 10^{-6}$ Wkg$^{-1}$ and $2.84 \times 10^{-4}$ m$^2$ s$^{-2}$. The surface layer is most likely affected by wind stress, since high correlation is found between dissipation of turbulent kinetic energy ($\varepsilon$) and wind stress (corr = 0.67). Meanwhile, on the thermocline layer, dissipation of turbulent kinetic energy ($\varepsilon$) is highly correlated with internal tidal wave oscillation (corr=0.68). Those suggest that the strong Southeast monsoon wind stress and internal tidal waves activities are the major forcing for the energetic turbulent mixing within the mixed layer and thermocline layer in Dewakang Sill.

Acknowledgement

The MAJAFLOX cruise is a part of the collaborative research between Department of Marine Science and Technology FPIK-IPB and Marine Geological Institute (P3GL) Bandung (Principal Investigator: A. Atmadipoera and R. Zuraida). A. Atmadipoera was supported by research grant from Kemenristek Dikti (PUPTN 714/IT3.11/PN/2016). We are very grateful to the crews of R.V. Geomarin III for their professional assistance during the field experiment. The first author would like to thank the LPDP Scholarship (Indonesia Endowment Fund for Education) for the master degree scholarship.

References

[1] Gordon A L, Orsi A H, Muench R, Huber B A, Zambianchi E and Visbeck M 2009 Western Ross Sea continental slope gravity currents Deep-Sea Res. II. 56 796-817
[2] Susanto R D, Ffield A, Gordon A L and Adi T R 2012 Variability of Indonesian throughflow within Makassar Strait, 2004–2009. Lamont-Doherty Earth Observatory, Columbia University, Palisades, New York, USA. J of Geophys Res. 117, doi :10.1029/2012JC008096
[3] Gordon A L 2001 Inter-ocean exchange, in Ocean Circulation and Climate: Observing and Modeling the Global Ocean. Siedler G, Church J, and Gould J (Ed) Int. Geophys. 77 303–314 (New York: Elsevier)
[4] Gordon A L, Giulivi C F and Ilahude 2003 A. Deep Topographic Barriers Within the Indonesian Seas Deep-Sea Res. 50 2205-2228
[5] Atmadipoera A, Mocard R, Madec G, Wijffels S, Sprintall J, Koch-Larrouy A, Jaya I, Supangat A 2009 Characteristics and variability of the Indonesian Throughflow water at the outflow straits Deep-Sea Res 1. 56 1942-1954
[6] Stewart R H 2003 Introduction to Phys Oceanogr. Texas A and M University. Departement of Oceanography
[7] Ffield A 1994 Tidal Mixing in the Indonesia Seas. Paper presented at Internasional Scientific symposium at the IOC-WESTPAC, IOC Bali, Indonesia
[8] Ffield A and Gordon A L 1992 Vertical mixing in the Indonesian thermocline J Phys Oceanogr. 22 184-195
[9] Koch-Larrouy A, Madec G, Bouruet-Aubertot P, Gerkema T, Bessi`eres L and Molcard R 2007 On the transformation of Pacific Water into Indonesian Throughflow water by internal tidal mixing Geophys Res Lett. 34 1-6
[10] Hatayama T 2004 Transformation of the Indonesian Throughflow Water by Vertical Mixing and Its Relation to Tidally Generated Internal Waves. Japan Agency for Marine-Earth Science and Technology, Yokohama Institute for Earth Sciences, Showa-machi, Kanazawa-ku, Yokohama, Kanagawa 236-0001, Japan J of Oceanogr. 60 569-585
[11] Suteja Y 2015 Percampuran Turbulen Di Selat Ombai J of Trop Mar Scien and Tech. 7 71-82.
[12] Purwandana A 2015 Distribusi Percampuran Turbulen di Perairan Selat Alor J Mar Scien 19 43-54
[13] Atmadipoera A, Kusmanto E, Purwandana A, Nurjaya I 2015 Observation Of Coastal Front And Circulation In The Northeastern Java Sea, Indonesia J of Trop Mar Scien and Tech 7 91-108

[14] Field A and Gordon A L. 1996 Tidal mixing signature in the Indonesian Seas. J Phys Oceanogr. 26 1924-1936

[15] Koch-Larrouy A, Atmadioera A, Van Beek P, Madec G, Aucan J, Lyard F, Grelet J and Souhaut M 2015 Estimates of tidal mixing in the Indonesian archipelago from multidisciplinary INDOMIX in-situ data Deep-Sea Part I. doi : 10.106/j.dsr.2015.09.007

[16] Backes J 2011 SBE 19plus CTD Operating and Repair Manual (Washington USA: Bellevue)

[17] Park Y H, Fuda J L, Durand I and A C N Garabato 2008 Internal tides and vertical mixing over the Kerguelen Plateau Deep-Sea Res Pt II. 55 582-593

[18] Thompson A F, Gille S T, Mackinnon J A, Sprintall J 2007 Spatial and temporal patterns of small-scale mixing in Drake Passage J Phys Oceanogr. 37 572-592

[19] Ferron B, Mercier H, Speer K, Gargett A, Polizin K 1998 Mixing in the Romanche Fracture Zone J Phys Oceanogr. 28 1929-1945.

[20] Dillon T M 1982 Vertical overturns: a comparation of Thorpe and Ozmidov length scale J Geophys Res. 87 9601-9613

[21] Finnigan T D, Luther DS and Lukas R 2002 Observation of enhanced diapycnal mixing near the Hawaiian Ridge J Phys Oceanogr. 32 2988-3002

[22] Koch S E, Desjardins M and Kocin P J 1983 An interactive Barnes objective map analysis scheme for use with satellite and conventional data J Climate Appl Meteor. 22 1487-1503

[23] Galbraith P S and Kelley E 1996 Identifying overturn in CTD profiles J Atmos Ocean Tech. 13 688-702

[24] Cisewski B, Strass V H and Prandke H 2005 Upper-ocean vertical mixing in the Antarctic polar front zone Deep-Sea Res II. 52 1087-1108

[25] Osborn T R 1980 Estimates of the local rate of vertical diffusion from dissipation measurements. J. Phys. Oceanogr. 10 83–89

[26] Pond S and Pickard GL 1983 Introductory Dynamical Oceanography Ed ke-2 (Oxford: Pergamon Press)

[27] Matsuno T, Shimizu M, Morii Y, Nishida H and Takaki Y 2005 Measurement of the Turbulent Energy Dissipation Rate around the Shelf Break in the East China Sea. J of Ocean. 61 1029-1037

[28] Moum J N and Smyth W D 2001 Upper Ocean Mixing Processes (Oregon (US): Academic Press)

[29] Naulita Y and Kitade Y 2011 Observed Turbulence Properties over the Continental Shelf and Slope off Jogashima, Sagami Bay Société franco-japonaise d’océanographie La mer. 49 1 – 15