



**HAL**  
open science

## Observation of internal tides, nonlinear internal waves and mixing in the Lombok Strait, Indonesia

Adi Purwandana, Yannis Cuypers, Pascale Bouruet-Aubertot

► **To cite this version:**

Adi Purwandana, Yannis Cuypers, Pascale Bouruet-Aubertot. Observation of internal tides, nonlinear internal waves and mixing in the Lombok Strait, Indonesia. *Continental Shelf Research*, 2021, 216, pp.104358. 10.1016/j.csr.2021.104358 . hal-03352800

**HAL Id: hal-03352800**

**<https://hal.science/hal-03352800>**

Submitted on 13 Feb 2023

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.



Distributed under a Creative Commons Attribution - NonCommercial 4.0 International License

# 1 **Observation of internal tides, nonlinear internal waves and mixing in** 2 **the Lombok Strait, Indonesia**

3 **Adi Purwandana<sup>1,2)</sup>, Yannis Cuypers<sup>1)</sup>, Pascale Bouruet-Aubertot<sup>1)</sup>**

4 <sup>1)</sup>*Laboratoire d'Océanographie et de Climatologie par Expérimentation et Approche Numérique (LOCEAN),*  
5 *Sorbonne Université, Paris, France*

6 <sup>2)</sup>*Research Center for Oceanography, Indonesian Institute of Sciences (RCO-LIPI), Jakarta, Indonesia*

7 *Corresponding author: adip001@lipi.go.id*

## 8 **Abstract**

9 Lombok Strait is one of the main exit passages of the Indonesian throughflow (ITF), connecting the  
10 interior Indonesian seas with the Indian Ocean. Satellite images have shown that this strait is also  
11 the most remarkable zone of the Indonesian seas for the generation of internal tides, which  
12 frequently evolve into internal solitary waves (ISWs). Here we present the first comprehensive  
13 high-frequency observations of the internal tide in the strait. The observations were carried out at a  
14 fixed position around 8 km from the western coast of Lombok Island at a depth of ~300 m. We  
15 confirmed the appearance of the non-linear internal tide with an amplitude of ~90 m at a fairly close  
16 distance (~30 km) from its generation site in the Nusa Penida Sills (NPS), with a dominant  $M_2$  tidal  
17 frequency. High-frequency internal waves associated with vertical velocities reaching  $20 \text{ cm s}^{-1}$   
18 were observed within the internal tidal trough. We estimated the turbulent kinetic energy dissipation  
19 rate using Thorpe and Ellison scales and found a mean dissipation rate of the order of  $\sim [10^{-6}-10^{-7}]$   
20  $\text{m}^2 \text{ s}^{-3}$ , with enhanced values in the thermocline layer, possibly driven by shear instabilities. We  
21 underlined that the use of the Ellison scale is a promising alternative for indirectly estimating the  
22 dissipation rate from high-resolution time-series data sets.

23 **Keywords:** Indonesian seas, internal solitary wave, Lombok Strait, dissipation rate, mixing

## 24 **1. Introduction**

25 The Indonesian seas connect the Pacific and Indian Oceans and represent a key passage of  
26 the Thermohaline Circulations (Gordon, 2005; Gordon and Fine, 1996). One of the most important

27 exit passages of the Indonesian throughflow (ITF) is the Lombok Strait, located in the Lesser Sunda  
28 Islands waters. This strait separates the Bali and Lombok Islands, between 115.3°-116.1° E and 8.2°-  
29 8.7° S and connects the Kangean waters (western Flores Sea) in the north, and the Indian Ocean in  
30 the south. The Lombok Strait is part of the western route of the ITF, a bypass that connects directly  
31 the ITF flowing from the Makassar Strait after passing the Dewakang waters. The flow through this  
32 strait represents 20% of the ITF (Hautala et al., 1996). In situ measurements in this strait from  
33 January 1985 to January 1996 (Murray and Arief, 1988) show an average transport of 1.7 Sv (1 Sv  
34 =  $1 \times 10^{-6} \text{ m}^3 \text{ s}^{-1}$ ) with a maximum of 3.8-4.0 Sv towards the Indian Ocean during July and August  
35 1985; and the minimum within 0-1 Sv is observed from December 1985 to January 1986. The flow  
36 in the Lombok Strait is controlled by the presence of the Nusa Penida Sill (hereinafter NPS) which  
37 reaches 350 m depth (Murray and Arief, 1988). The circulation of water masses underneath the  
38 Pacific water mass is restricted by the NPS.

39 The Lombok Strait basin is dominated by Pacific water masses, i.e. the North Pacific  
40 subtropical water (NPSW), fed by the Makassar Strait throughflow and characterized by  $\sigma_\theta = 24.5$   
41  $\text{kg m}^{-3}$  at 100-200 m depth (Hautala et al., 1996; Wyrki, 1961). This strait waters have distinctive  
42 characteristics compared to other exit passages along the Lesser Sunda Islands chain. The salinity in  
43 the Lombok Strait is influenced by the fresher water from the Java Sea and by local precipitation  
44 during Northwest Monsoon (December-May). A freshening, which can exceed 2 PSU, occurs from  
45 March to May (Sprintall et al., 2003).

46 Spatial variation in the hydrodynamics of the Lombok Strait is unique. The flow in the  
47 Lombok Strait shows a strong layered structure with a permanent southward flow in the upper layer  
48 and a reversing southward-northward flow in the lower layer (Hautala et al., 2001). The upper  
49 southward flow is maximum in March-April while the reversing northward flow is maximum in  
50 May. The southward flow is controlled by the low-frequency component of the large-scale wind  
51 forcing from the Pacific Ocean; while the lower layer reversing flow is correlated with semi-annual  
52 Kelvin wave events forced by westerly winds during the Northwest to Southeast Monsoon

53 transition period (November/December and April/May) in the equatorial Indian Ocean (Potemra et  
54 al., 2002; Sprintall et al., 2003). Since the strait opens directly onto the Indian Ocean, the northward  
55 lower layer flow generated by Kelvin waves can result in intrusions of the Indian Ocean waters  
56 (Sprintall et al., 1999; Susanto et al., 2001; Wijffels et al., 2002). The throughflow in the Lombok  
57 Strait also varies longitudinally. The stronger southward flow is observed in the western side of the  
58 strait, a phenomenon related to the higher sea level height on the Lombok Island side than that on  
59 the Bali Island side (Hautala et al., 2001; Murray and Arief, 1988).

60 The strong mixing of Pacific watermass observed in the Indonesian seas generates cold  
61 surface anomalies and fresh pycnocline anomalies, which impacts the Indian and Pacific heat and  
62 freshwater budgets (Lee et al., 2002) as well as local deep atmospheric convection and wind  
63 regimes (Koch-Larrouy et al., 2010). This mixing is mainly driven by internal tide induced  
64 turbulence as shown by previous in situ observations (Bouruet-Aubertot et al., 2018; Koch-Larrouy  
65 et al., 2015; Purwandana et al., 2020; Robertson and Field, 2005), satellite images of sea surface  
66 temperature (Ray and Susanto, 2016) and modeling studies (Hatayama, 2004; Koch-Larrouy et al.,  
67 2007; Nagai et al., 2017; Nagai and Hibiya, 2015; Nugroho et al., 2018; Schiller, 2004). Internal  
68 tides are generated in stratified waters when the tidal flow encounters a topographic feature. A  
69 vertical motion is induced, and the oscillating water mass will be subjected to the buoyancy and  
70 Coriolis restoring force. In the Indonesian seas, tidal mixing is particularly intense over rough  
71 topography, such as in the Lifamatola, Manipa, Ombai, and Lombok Straits, and the Sibutu Island  
72 chains (Purwandana et al., 2020).

73 The tide in the Lombok Strait is produced by tidal waves propagating from the Indian Ocean  
74 with a dominant  $M_2$  semi-diurnal period of 12.4 hours. The tidal flow can reach  $\pm 3.5 \text{ m s}^{-1}$  in the  
75 sill area (Murray and Arief, 1988). Numerical simulations have shown a strong conversion of  
76 semidiurnal  $M_2$  barotropic to baroclinic internal tides around the NPS (Nagai and Hibiya, 2015;  
77 Nugroho et al., 2018) which shows that the Lombok Strait is a strong mixing hot spot. The  $1/100^\circ$   
78 simulation map of the  $M_2$  internal tides energy flux and depth-integrated dissipation is reproduced

79 in **Figure 1a** evidencing the strong northward  $M_2$  internal tides fluxes as well as the region of very  
80 strong dissipation at the sill and along the eastern coast of Bali and western coast of Lombok.  
81 Analysis of satellite sea surface temperature also revealed a spring neap cycle in the Lombok Strait  
82 evidencing strong mixing associated with tides there (Ffield and Gordon, 1996; Ray and Susanto,  
83 2016).

84 Strong amplitude internal tides generally evolve non linearly into steep front and trains of  
85 short-scale large amplitude solitary waves. Such solitary waves trains can transport energy far away  
86 from the generation site and to dissipate it when they shoal at continental slopes. This process has  
87 important consequences for local biogeochemical dynamics (Bourgault et al., 2014). Two typical  
88 situations are encountered, when the flow at the sill is always subcritical, ( $Fr = U_{bar}/c < 1$ ) with  $U_{bar}$   
89 the barotropic tidal velocity and  $c$  the internal mode phase speed, the internal tides steepens  
90 progressively as it propagates away from the sill and then breaks down into solitary waves train  
91 under the effect of non-hydrostatic dispersion (Gerkema and Zimmerman, 1995). This is the first  
92 process that is believed to take place in the extensively studied, Luzon Strait region (South China  
93 Sea) (Alford et al., 2015). When the flow gets supercritical at the sill, the wave propagation is  
94 impeded a depression grows and is released when the tide slackens, the depression then rapidly  
95 breaks into a train of solitary waves. This second mechanism, known as the lee waves mechanism,  
96 has been observed in several well know Internal Solitary Waves (hereinafter ISWs) generations sites  
97 such as the Strait of Messina (Brandt et al., 1999) or the Sulu Sa (Apel et al., 1985). The strong  
98 Lombok Strait internal tides can evolve into a train of large vertical displacement (~100m) small  
99 horizontal scale (~km) ISWs (Aiki et al., 2011; Matthews et al., 2011). The model of Nagai and  
100 Hibiya (2015) also indicates that the northward propagating energy flux is stronger than that of the  
101 southward favoring ISW propagating northward (**Figure 1a**). ISW indicators are often observed  
102 from satellite images as sea surface alternating bands of slicks and rough patches propagating from  
103 NPS and became a classical illustration of ISW observation from space (see Jackson, 2007). Yet, the  
104 only in situ observation of internal solitary waves in the Lombok Strait is a short snapshot of an

105 echosounder by Susanto et al. (2005) evidencing an impressive solitary wave of ~150 m vertical  
106 displacement. Satellite and numerical-based studies in the Lombok Strait revealed that the Lombok  
107 ISWs propagate predominantly northward of the strait (Aiki et al., 2011; Matthews et al., 2011).  
108 Their studies also suggest that the ITF tends to reduce the frequency and occurrence of ISW. The  
109 strong barotropic tidal velocities at the sill also suggest that the ISW are rather generated through  
110 the lee waves mechanism.

111 Here, we present the first high-frequency observations of the non-linear internal tides and  
112 associated high-frequency waves in the Indonesian seas, i.e. in the Lombok Strait. We characterize  
113 the impact of internal tides and high-frequency waves on turbulence in the pycnocline using both  
114 the Thorpe scale method applied to CTD profiles and the Ellison scale method applied to high-  
115 frequency time series of temperature sensors.

116 This paper is organized as follows. Section 2 describes the observation dataset and methods  
117 to derive the mixing estimates from observations. We put the detail of the method in the Appendix.  
118 Section 3 presents results and discussion gained from observations, including the analysis of the  
119 characteristics of the observed ISW and mixing estimates induced by internal tides. The results are  
120 summarized in section 4.

## 121 **2. Methodology**

### 122 **2.1 In situ Observations**

123 A field experiment was conducted during a spring tide period at the transition between  
124 Southeast Monsoon and Northwest Monsoon, on 3-5 November 2017 in the Lombok Strait waters  
125 at a fixed position,  $8.56^{\circ}$  S,  $116.02^{\circ}$  E offshore of the west coast of Lombok Island (~8 km) at a  
126 depth of 300 m during the transition from neap to spring tide (red star in **Figure 1a**). The field  
127 experiment was conducted from a small anchored boat, which precluded a deep location in the  
128 middle of the channel or too close to the sill because of the very strong currents. This location is  
129 thus appropriate for characterizing possible enhanced dissipation resulting from ISW interaction

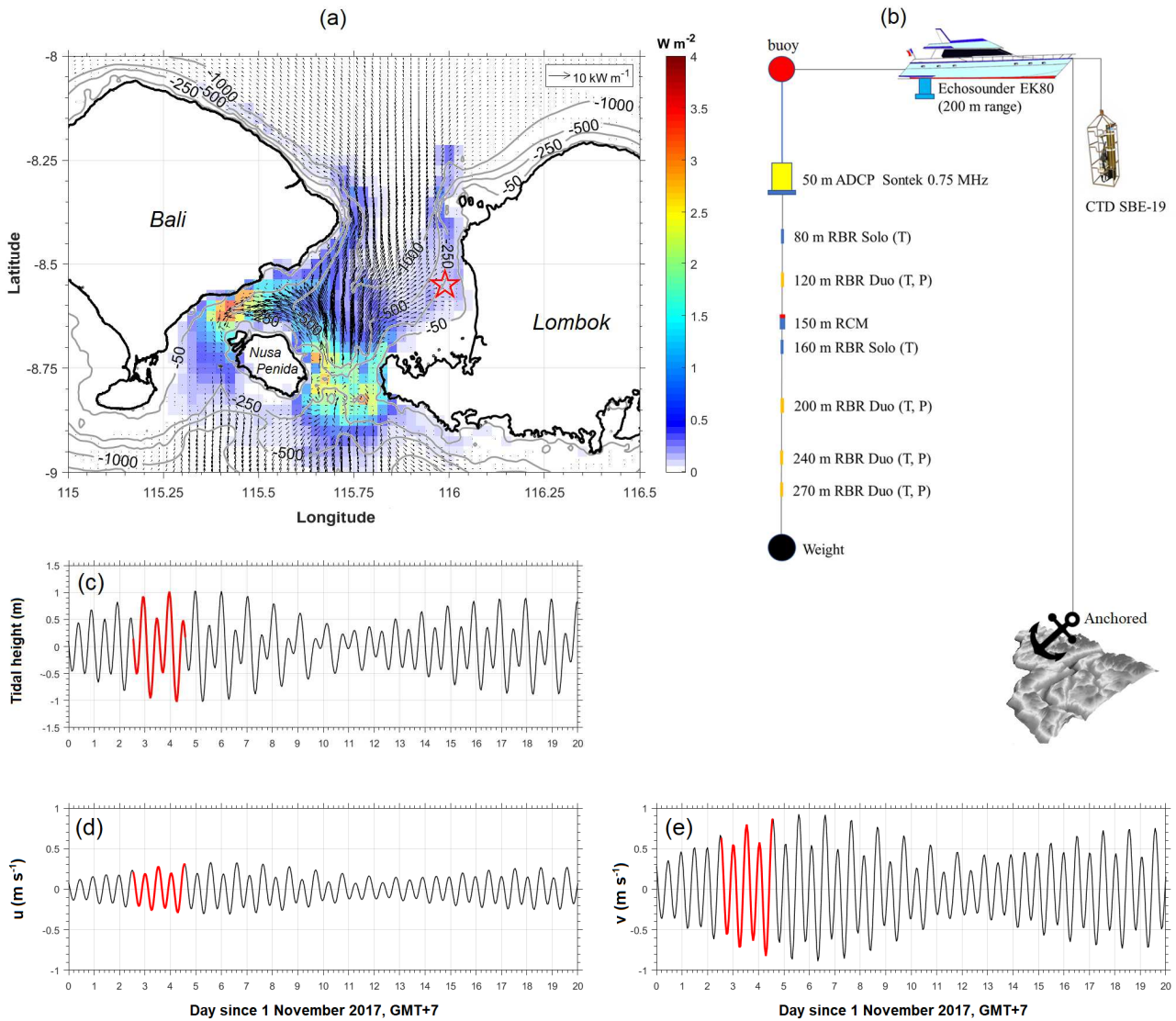
130 with the shore but does not allow to characterize the most coherent unaltered ISW structure near  
 131 their generation site. A detailed description of the measurements is shown in **Table 1**.

132 **Table 1** Available in situ measurements in the Lombok Strait, 3-5 November 2017.

Date	Hour of CTD casts (GMT+8)	Other operated instruments
November 3 <sup>rd</sup> , 2017	15.23; 18.18; 19.28; 21.24; 23.38	
November 4 <sup>th</sup> , 2017	02.12; 03.39; 04.30; 07.20; 09.27; 11.10; 13.25; 15.28; 15.46; 16.08; 18.40; 21.19	ADCP, RCM, EK80, RBRs
November 5 <sup>th</sup> , 2017	00.14; 02.07; 04.06; 04.26; 04.45; 07.18; 07.46; 08.00; 10.30; 12.22	

133 High-frequency imaging of the internal wave structure was obtained using a 200 kHz  
 134 narrow-beam echosounder Simrad EK80 that covered up to 150 m depth. A 0.75 MHz downward-  
 135 looking *Sontek* Acoustic Doppler Current Profiler (ADCP) was moored on a floater connected to the  
 136 anchored boat. The instrument was lowered at a depth around 50 m and measured the currents down  
 137 to ~100 m depth with 5 m bin resolution and 1 minute time averaging. An *Anderaa Seaguard* RCM  
 138 was also deployed at the depth of ~150 m to get the current variability in the deeper layer, sampling  
 139 with 1-minute temporal resolution. Temperature field variability was measured at high frequency (1  
 140 Hz) using 4 RBR duet sensors (temperature and pressure) and 2 RBR solo sensors (temperature  
 141 only) that were attached at several depths between 80 m and 270 m. Note that the internal tide  
 142 motion-induced large heaving of the mooring instruments, hence the portion of the water column  
 143 sampled was not constant. Moreover, the mooring was anchored to the boat but could oscillate  
 144 horizontally with an amplitude of ~100 m. The instantaneous absolute velocities can be biased by  
 145 this motion, and for this reason, we consider only in the analysis the horizontal shear and the time  
 146 mean velocity where high-frequency oscillations of the ADCP position around the boat are  
 147 smoothed out. Temperature and salinity profiles were collected using a *Sea-Bird Electronics* (SBE)  
 148 CTD 19 Plus from the surface to depths of ~200 m around every two hours and 2-3 yoyo casts were  
 149 repeated when a sequence of internal waves was observed in the echosounder. The profiles were de-

150 spiked and averaged into 0.5 m using the SBE data processing modules. Data above 5 m were  
 151 removed due to contamination of the CTD probe soaking effect and ship's wake. Predicted tidal  
 152 height, zonal and meridional current components gained from Oregon State University tidal  
 153 inversion software (OTIS) (Egbert and Erofeeva, 2002), including the observation time in the  
 154 Lombok Strait and field observation design is shown in **Figure 1c-e**.



155

156 **Figure 1** (a) A  $0.025^\circ$  grid averaged of the depth-integrated semidiurnal  $M_2$  driven turbulent kinetic  
 157 energy (TKE) dissipation rate (colored plot,  $W m^{-2}$ ) and depth-integrated  $M_2$  baroclinic energy flux  
 158 ( $kW m^{-1}$ ) inferred from a 3D hydrostatic model by Nagai and Hibiya (2015) (personal  
 159 communication) in the Lombok Strait. The bathymetry is provided by the Gebco\_08 grid. The red  
 160 star is the observation site. (b) The measurements scheme, which involved ADCP, CTD,



161 echosounder and RBR sensors, (c, d, e) is the predicted tidal height, zonal current ( $u$ ), and  
162 meridional current ( $v$ ), respectively on 1-20 November 2017 above the Nusa Penida sills. Red lines  
163 indicate the timing of field observation on 3-5 November 2017.

## 164 **2.2 Mixing estimates**

165 Turbulence dissipation rates were inferred using both an improved Thorpe scale ( $L_{Th}$ )  
166 method (Purwandana et al., 2020) and Ellison scale ( $L_E$ ) method from the CTD casts and high  
167 temporal resolution temperature sensors at a fixed depth measured by moored instruments,  
168 respectively. Both Thorpe and Ellison methods are indirect estimates to infer the dissipation rates  
169 based on the detection of the largest turbulent eddies signature. Both methods rely on the  
170 assumption of a statistical linear relationship between the largest overturns size and the Ozmidov  
171 scale (Mater et al., 2015). The Ozmidov scale is a theoretical dynamical scale directly related to the  
172 turbulent energetics through  $\varepsilon$  as  $L_o = \varepsilon^{(1/2)} N^{(-3/2)}$ . A turbulent overturn will have a signature in term of  
173 density inversion along  $z$  for a given time (Thorpe method) and high-frequency density oscillation  
174 at a given depth (Ellison scale method). For both methods, we have used temperature as a proxy for  
175 density, as justified in the Appendix. For the Thorpe method we obtain an estimation of the largest  
176 overturns size from the standard deviation of density inversions (built from temperature), while for  
177 the Ellison scale method, the largest overturn size is obtained from the standard deviation of  
178 turbulent time fluctuations of temperature divided by the mean temperature gradient.

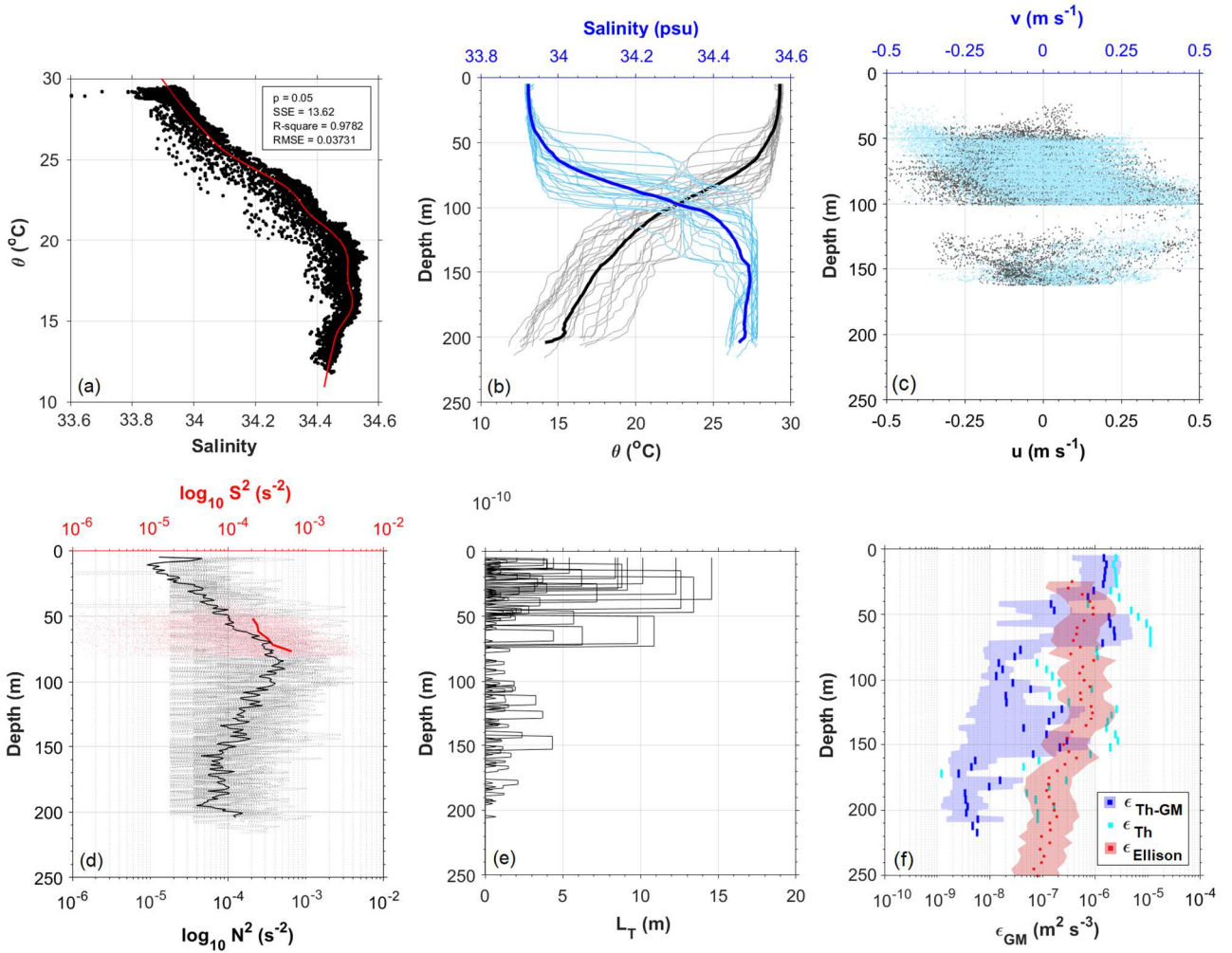
179 The Ellison method estimation of the dissipation rate was found comparable to the  
180 microstructure measurements direct estimates according to a study by (Moum, 1996); yet as stated  
181 by Cimadoribus et al. (2014), the Ellison scale may overestimate/underestimate the real turbulence  
182 dissipation rate during the initial stage/late stage of the turbulence decay. The Thorpe method uses a  
183 single temperature/density profile and inspects turbulent events represented by overturning events  
184 where the denser parcel lies above the less dense parcel. The Ellison method inspects turbulence  
185 events from high-resolution time-series measurements and inspects any fluctuations which are  
186 expected to be related to turbulence once this signal is decomposed into some possible sources:

187 surface waves, internal waves and turbulent events. . In our case, the screening for the temperature  
188 is crucial to omit any spurious fluctuations related to the high-frequency surface waves and internal  
189 waves (see Appendix). Our dissipation rate estimate by employing the time series data sets using the  
190 Ellison scale method is similar to a study by Cimatoribus et al. (2014). The detailed processing for  
191 Thorpe and Ellison scale method is provided in the Appendix.

## 192 **3. Results and Discussion**

### 193 **3.1 Hydrography**

194 Our observations were conducted during a transition period from the Southeast Monsoon to  
195 the Northwest Monsoon. The stratification derived from the CTD profiles shows a thick pycnocline  
196 extending between 50 and 150 m with a maximum stratification near 85 m depth (**Figure 2d**). The  
197 time-mean currents (over the four  $M_2$  tidal periods recorded) show a strongly sheared structure  
198 with strong background westward-eastward zonal current of  $\sim 0.3 \text{ m s}^{-1}$  and northward-southward  
199 meridional current of  $\sim 0.5 \text{ m s}^{-1}$  (**Figure 2c**). We suggest that the southward currents are related to  
200 the ITF current, while the northward currents are presumably influenced by semi-annual Kelvin  
201 wave events, which appear during the Southeast Monsoon transition period (November/December  
202 and April/May) (Sprintall et al., 2003, 1999; Susanto et al., 2001; Wijffels et al., 2002). The time-  
203 mean squared shear measured by the ADCP between 55 and 100 m shows strong values exceeding  
204 the  $N^2$  values and leading to near-critical  $Ri$  values.



205

206 **Figure 2** (a) Temperature-salinity relationship inferred using spline fit with the goodness values are  
 207 indicated, (b) potential temperature and salinity profiles (grey and light blue lines, respectively)  
 208 with its mean values (black and blue lines, respectively); (c) zonal,  $u$  and meridional,  $v$  (grey and  
 209 light blue plots, respectively); (d) the squared buoyancy and squared shear profiles (black and red  
 210 plots, respectively) with its mean values (lines), (e) the Thorpe length; (f) the 10 m mean dissipation  
 211 rate with background Garrett-Munk (GM) dissipation rate applied (blue), without background GM  
 212 dissipation (cyan) and that of the 10 m running mean dissipation values inferred using Ellison scale  
 213 (red). The shading plot for the GM-Thorpe and Ellison estimates in (f) indicates the 95% confidence  
 214 intervals values for each bin.

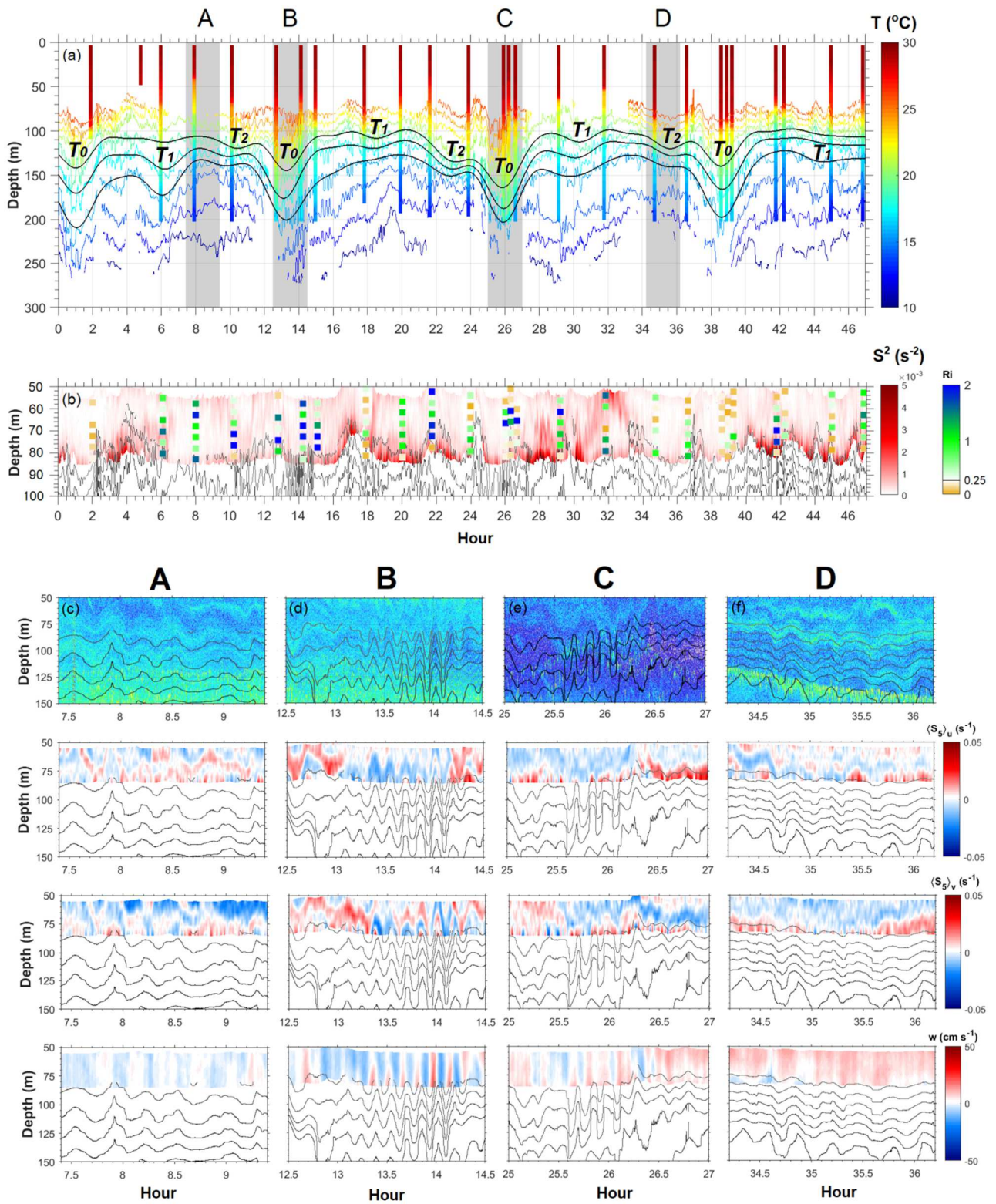
215 The time series of temperature recorded from the mooring sensors show a clear non-linear  
 216 semi-diurnal signal characterized by semi-diurnal periodic large troughs (**Figure 3a**). The peak to  
 217 peak amplitude reaches  $\sim 90$  m for the 20°C isotherm located in the pycnocline, which is larger than

218 internal tides observed in the Makassar Strait and Buru Strait (Purwandana et al., 2020); Seram Sea,  
219 and Flores Sea (Ffield and Robertson, 2008); and comparable to the largest observed internal tides  
220 in the North Halmahera Sea and Ombai Strait, which were associated to very large turbulent  
221 dissipation rates (Bouruet-Aubertot et al., 2018). The low pass filter (period >3 hours) of 18°, 20°,  
222 22° C isotherms time series show that each M<sub>2</sub> period is composed of one large trough ( $T_0$ ) of ~70  
223 m and two smaller troughs ( $T_1$ ,  $T_2$ ) of ~20 m, evidencing higher harmonics (M<sub>4</sub>, M<sub>6</sub>) in the signal.

### 224 3.2 ISWs characteristics

225 The barotropic velocities predicted by the OTIS model (Egbert and Erofeeva, 2002) over the  
226 NPS,  $U_{bar}$ , reach ~0.9 m s<sup>-1</sup> during the period we sampled, while the mode-1 phase speed,  $c_1$  is  
227 typically of 1 m s<sup>-1</sup>. This gives the Froude number,  $Fr = U_{bar}/c_1$ , close to supercritical hence it is  
228 expected that ISW are generated by the lee waves mechanism (Maxworthy, 1979). Numerical  
229 experiments study (Aiki et al., 2011) and satellite observation images (Astawa Karang et al., 2012)  
230 show that an ISW packet can already be formed at about 10-20 km from NPS which is consistent  
231 with a rapid formation by a lee wave mechanism. In our observations, high-frequency waves are  
232 observed almost continuously, but the largest amplitude wave-packet is observed in the large trough  
233 of the second tidal period (**Figure 3** event B). A significant wave packet is also observed during the  
234 third period (**Figure 3** event C) but no clear wave packet is observed within the large trough of the  
235 first and fourth periods. The waves generated during these events have a strong vertical coherency, a  
236 maximum amplitude of ~40 m, a short period of 9-12 minutes, and are associated with strong  
237 vertical velocities reaching 20 cm s<sup>-1</sup>. The wave packet A has also a clear signature in the echo-  
238 sounder, consistent with the isotherms displacements. The echo-sounder signal for packet B was  
239 unfortunately too noisy to be exploited. The high frequency waves induce a clear heaving of a  
240 background strongly layered shear. Since the high frequency waves are highly coherent vertically,  
241 they do not likely contribute to this small-scale (~20 m vertical scale) shear structure. This  
242 background shear structure may result from the interaction of internal tide, ITF, Kelvin waves and  
243 possibly near inertial waves, but our measurements period is too short to characterize them.

244            Distinct high-frequency wave packets are also generated on the front face of  $T_2$  (as shown  
245 by the close up A and D) and are likely generated by the higher tidal harmonics composing the  
246 internal tide. They are characterized by a longer wave period of ~20 minutes and smaller maximum  
247 amplitude of ~20 m. They are associated with weaker vertical velocities of  $\sim 5 \text{ cm s}^{-1}$ . They also  
248 induce clear heaving of the small-scale shear structure.



249

250 **Figure 3** (a) Isotherm contours inferred from RBR sensors and temperature profiles from CTD  
 251 casts. The black line is low pass filtered ( $>3h$ ) for isotherms 18, 20, and 22° C. (b) The squared  
 252 shear with isotherms and Richardson number,  $Ri$  superimposed. (c, d, e, f) The enlarged view of  
 253 shaded areas A, B, C, and D in (a) for echosounder image (in arbitrary unit) in the first upper panel,  
 254 zonal shear,  $(S_5)_u$ , meridional shear  $(S_5)_v$ , and vertical current  $w$ . Enlarged images for annotated areas

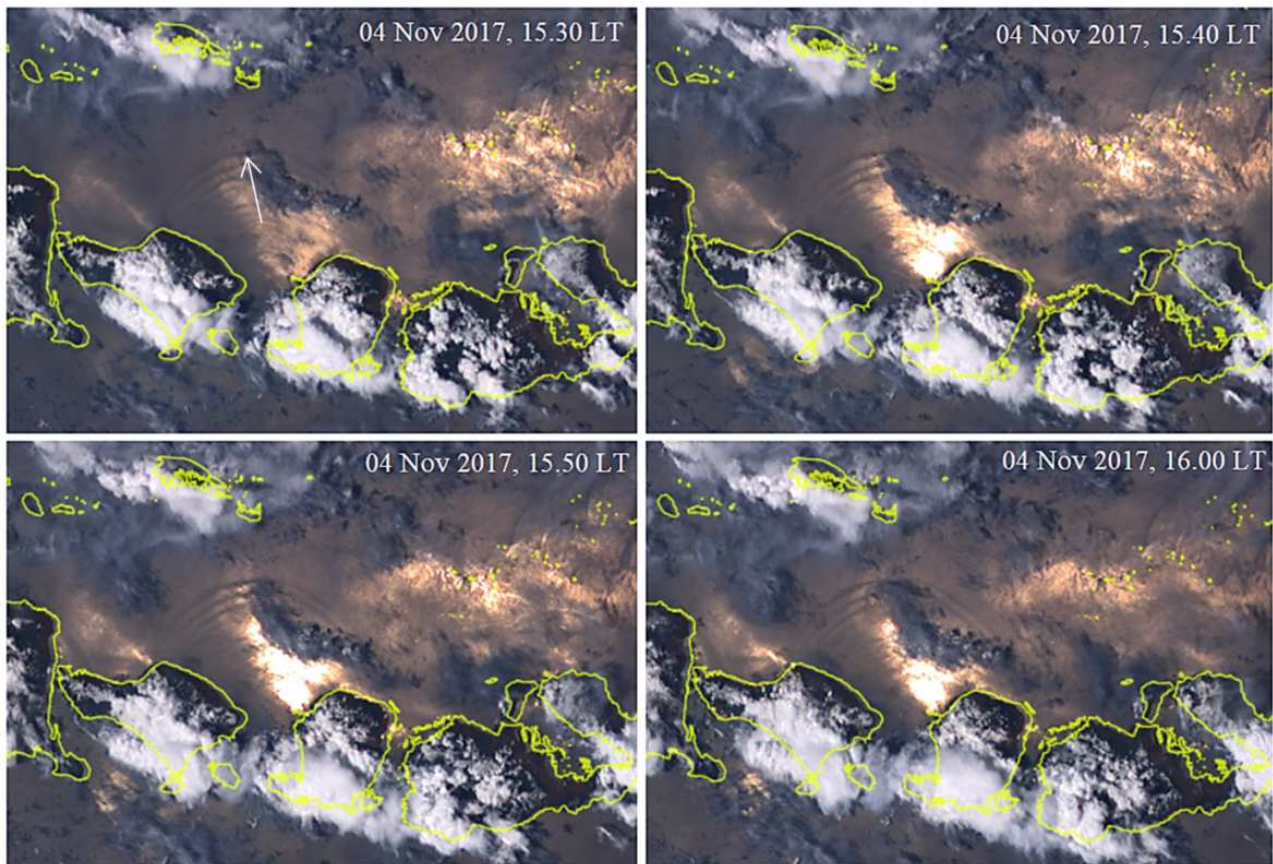
255 A, B, C, and D denote the appearance of shortperiod internal waves (B, C) and longer period  
256 internal waves (A, D).  $T_0$  and  $(T_1, T_2)$  are the large and the small troughs, an indication of  $M_2$  and  
257 higher harmonic signals ( $M_4, M_6$ ), respectively.

258         The phasing of the largest high-frequency waves (events B, C) with the large troughs is  
259 consistent with a lee wave mechanism, yet none of the high-frequency waves observed is rank-  
260 ordered and they do not show a typical *sech*<sup>2</sup> solitary waves profiles, as is generally expected from  
261 the lee wave mechanism (Apel et al., 1985; Maxworthy, 1979). This may have been the result of an  
262 interaction with the sloping topography at our observation site. **Figure 1a** shows that the modeled  
263 internal tide fluxes are indeed across the slope at our location and that the dissipation is increasing  
264 onshore suggesting strong interaction of the internal tides with the slope. In addition, the depth  
265 variation and friction effects can alter the nonlinear terms driving the solitary wave packet  
266 evolution, as was shown from classical extended KdV framework in the Northwest shelf of  
267 Australia by Holloway et al. (2002). Simulations and observations of non-rank-ordered wave  
268 packets were also reported at the Stellwagen Bank in the Massachusetts Bay in conditions of  
269 moderate to strong forcing while rank-ordered packets were recovered for weaker (neap tide)  
270 forcing (Scotti et al, 2007). Numerical and observational studies in the Gibraltar Strait by Vlasenko  
271 et al. (2009) have also evidenced frequent non-rank-ordered packets that were attributed to multiple  
272 reflections from lateral boundaries along with friction and wave breaking. They also observed that  
273 if the non-rank-ordered wave packet pursues its propagation over a flat bottom, the high-frequency  
274 waves re-order.

275         It is interesting to notice that satellite observations from the Himawari-8 satellite made on  
276 November, 4<sup>th</sup> between 15.30 and 16.00 LT (local time) show the propagation of a train of ISW with  
277 a clear pattern at the northern exit of the strait (**Figure 4**). These images were captured about 12  
278 hours after we observed the main high-frequency wave packets B. The leading ISW is about 80 km  
279 from our in-situ observation location, which suggests a phase speed of  $\sim 1.9 \text{ m s}^{-1}$ , consistent with  
280 previous estimates (Susanto et al., 2005). The decreasing satellite signal intensity from the leading



281 wave to the trailing waves suggests that the wave packet is rank-ordered when it exits Lombok  
282 Strait and that the non-rank order signal we observe is a local process.



283

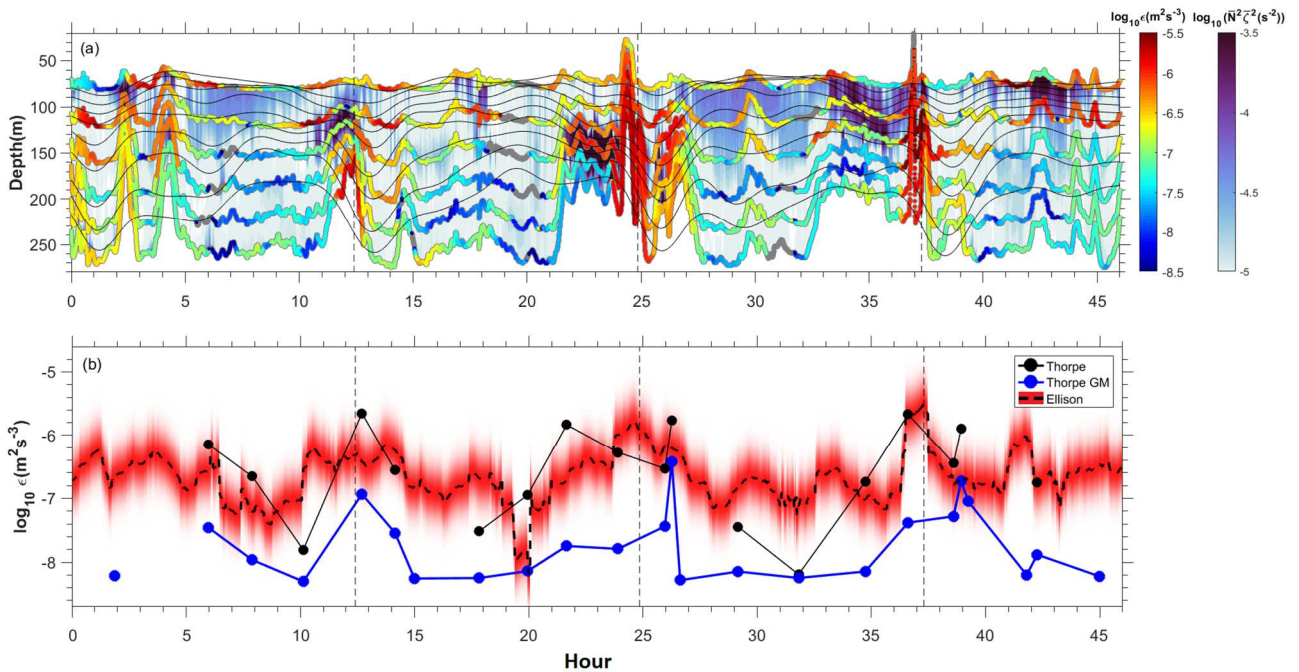
284 **Figure 4** The observed ISWs signature (white arrow) by satellite images Himawari-8 on November  
285 4<sup>th</sup>, 2017 at: (a) 15:30 LT (GMT+8), (b) 15:40 LT, (c) 15:50 LT and (d) 16:00 LT.

286 The fact that the ISW train is weaker on the third large trough and absent on the fourth large  
287 trough is surprising considering that the barotropic forcing is increasing. A similar strong variation  
288 of the number of high-frequency waves from one tidal cycle to another was also reported at the  
289 Stellwagen Bank (Scotti et al., 2007). According to Scotti et al. (2007), changes in stratification and  
290 forcing alone could not explain this variability. The dissipation and 3D effects should be considered  
291 to explain this chaotic behavior. As previously stated, these effects are likely important considering  
292 that a large part of the internal tides flux is directed onshore at our location and that the dissipation  
293 rate increases toward the shore.



294 **3.3 Dissipation estimates**

295 Our ADCP measurements within the thermocline layer as shown in **Figure 2d** show a wide  
 296 range of variation of the shear. The mean value of the squared shear profile is higher than the mean  
 297 of the squared buoyancy profile while the instantaneous profile shows frequent supercritical  
 298 conditions ( $Ri=N^2/S^2 < 0.25$ ), which is remarkable considering that we are located in the strongly  
 299 stratified thermocline. Elevated dissipation rates are observed in the temporal evolution of  $\epsilon_{E\text{-patch}}$   
 300 (**Figure 5b**), where the most persistent elevated dissipation rate occurred in the upper layer >120 m  
 301 depth.



302  
 303 **Figure 5** (a) Spatiotemporal evolution of the dissipation rate estimated using Ellison scale ( $\epsilon_E$ ) at  
 304 several mooring depths (scaled in  $\log_{10}$ , in colored contours; note that the depth of the dissipation  
 305 estimate follows the sensors displacements), the strain (scaled in  $\log_{10}$ , in blue-scaled contours), and  
 306 the low pass filtered (>3h) isotherms (black contours); (b) mean over the depth covered by the  
 307 mooring of the dissipation rates inferred using Thorpe scale patch and that considering the GM-  
 308 background dissipation rate (black dotted line and blue dotted line, respectively), and using Ellison  
 309 scale (dashed line for the mean and shaded red area for the confidence interval assuming the

310 Gaussian distribution shown in Figure A3b for the error made on the estimate of the local  
311 temperature gradient. Note that the shading intensity decreases with the probability.

312 The time and depth mean values of the dissipation rate result in an upper bound patch  
313 estimate of  $\varepsilon_{\text{Th-patch}}=2.19\times 10^{-6} \text{ m}^2\text{s}^{-3}$  and its standard deviation  $4.56\times 10^{-6} \text{ m}^2\text{s}^{-3}$ ; the lower bound GM  
314 background estimate of  $\varepsilon_{\text{Th-GM}}=2.88\times 10^{-7} \text{ m}^2\text{s}^{-3}$  and its standard deviation  $7.61\times 10^{-7} \text{ m}^2\text{s}^{-3}$ ;  
315  $\varepsilon_{\text{E}}=3.28\times 10^{-7} \text{ m}^2\text{s}^{-3}$  and its standard deviation  $3.85\times 10^{-7} \text{ m}^2\text{s}^{-3}$ . The time-mean depth profile of the  
316 dissipation rate varies in the range  $[10^{-8}\text{-}10^{-6}] \text{ m}^2\text{s}^{-3}$  with smaller values for the Thorpe method at the  
317 pycnocline oscillating in the range 75-120 m (see **Figure 2f**), which may result from a bias in the  
318 detection of overturns for the strongest stratification. These values characterize strong dissipation  
319 rates in the pycnocline, which can be compared to the highest thermocline dissipation rate found in  
320 the Halmahera Sea during the INDOMIX Program (Bouruet-Aubertot et al., 2018). The discrepancy  
321 between the Thorpe and Ellison methods is larger when we consider the lower bound inferred from  
322 the GM background dissipation rate estimate that results from the much less frequent detection of  
323 overturns in the CTD profiles than that of in the temperature-thermistor time series profile. The time  
324 evolution of the depth-mean dissipation rate of  $\varepsilon_{\text{Th}}$  and  $\varepsilon_{\text{E}}$  is shown in **Figure 5b**. A clear semi-  
325 diurnal pattern emerges with dissipation peaks exceeding  $1\times 10^{-6} \text{ m}^2\text{s}^{-3}$  and reaching  $1\times 10^{-5} \text{ m}^2\text{s}^{-3}$   
326 observed within the large internal tide troughs (marked by vertical dashed lines in **Figure 5b**).  
327 Similar observations have been made in a different context by Xie et al. (2013) in the central Bay of  
328 Biscay where maximum dissipation is encountered in narrow troughs of the nonlinear internal tides.  
329 It is indeed expected that the front of the internal tide will be associated with a sharp increase of the  
330 velocity, velocity shear, and strain which can favor shear and/or convective instabilities. In our  
331 observations, the tidal front is associated with an initial squeezing of the isopycnals at the tidal edge  
332 followed by stretching during the tidal abrupt trough. A strong strain is notably generated during the  
333 squeezing of the isopycnal at the leading edge of the wave just preceding the abrupt fall (we have  
334 added a computation of the large-scale strain in the background of **Figure 5a**). This strain is  
335 associated with strong dissipation rates. In this study, the background scaled-strain is derived from

336 the temperature sensors as  $\bar{N}^2 \bar{\zeta}^2$  with  $\zeta = (N^2 - \bar{N}^2)/\bar{N}^2$ ,  $N^2$  is the time varying stratification  
337 computed from the temperature derived potential density and  $\bar{N}^2$  is its time average. Similar  
338 periodic strain/dissipation increase has been described by Klymak et al. (2008) in the vicinity of the  
339 deep Kaena ridge where strong internal tides are generated. They attributed this periodic dissipation  
340 to direct breaking resulting from convective instabilities. Unfortunately, our ADCP sampling is of  
341 limited range so that we can not determine the  $Ri$  number below  $\sim 90$  m depth to characterize the  
342 possible onset of shear instabilities or local  $Fr$  number for convective instabilities. In any case, the  
343 phase lock of the dissipation with the tides suggests a rapid “direct” breaking mechanism as  
344 opposed to a weak wave-wave interaction mechanism as parameterized by the Gregg-Henyey-  
345 Polzin (hereinafter GHP) parameterization (Polzin et al., 2014). This is in agreement with the  
346 conclusions of Bouruet-Aubertot et al. (2018) and Klymak et al. (2008) that the GHP  
347 parameterization may not be valid in the near field of the internal tides. The time series of the  
348 dissipation rate from the Ellison estimate also shows a decreasing trend of the dissipation rate with  
349 depth, with the higher values observed for the two shallower temperature sensors (80 and 120 m  
350 average depth) surrounding the pycnocline peak where frequent supercritical conditions are  
351 observed from shear measurements.

#### 352 **4. Summary**

353 We provide the first high-frequency in-situ observations of strong semi-diurnal internal tides  
354 ( $\sim 90$  m amplitude) in the Lombok Strait at a short distance of about 30 km from the generation  
355 source at the Nusa Penida Sill. Frequent signatures of high frequency (10-20 minutes period), high  
356 vertical velocity internal wave trains are observed. The largest high-frequency wave packets (with  
357 waves reaching up to 40 m) are observed within the  $M_2$  periodic large internal tide trough, but  
358 weaker high-frequency wave packets seem to be associated with higher tidal harmonics. Although  
359 satellite observations during the same period suggest the presence of a well-organized rank-ordered  
360 ISW packet north of the Lombok Strait, the high-frequency wave packets we observed are not rank-  
361 ordered. We attribute this effect to the possible interaction of the high-frequency waves with the

362 sloping topography. Another striking feature is the disappearance of ISW within the fourth observed  
363 internal tidal trough while the barotropic forcing is increasing. We have no clear explanation for this  
364 observation yet, it is possible that the strong mixing generated at the sill alters the stratification  
365 structure and the generation of ISW. It is also possible that stronger nonlinearities, dissipation, 3D  
366 effects, and interactions with the Lombok slope disrupt the wave packet structure.

367 Two turbulence length scales, the Thorpe and Ellison scales, were used to estimate the  
368 turbulent dissipation rate from intermittent CTD measurements and high temporal resolution  
369 temperature sensors. The dissipation rate ( $\epsilon_{\text{Th-GM}}$ ,  $\epsilon_{\text{E-GM}}$ ) in the Lombok Strait thermocline is high of  
370 the order of  $\sim[10^{-6}\text{-}10^{-7}] \text{ m}^2\text{s}^{-3}$ , comparable to that observed in other turbulent hot-spot locations  
371 such as the north and south Halmahera Straits and the Ombai Strait (Bouruet-Aubertot et al., 2018).  
372 The shear measurements made over a limited layer in the upper thermocline show that frequent  
373 supercritical conditions are reached. The shear vertical structure is characterized by a small scale of  
374  $\sim 20$  m with high-frequency variation associated to the high-frequency waves heaving. We also  
375 found a clear phasing between the  $M_2$  internal tidal trough and the elevated dissipation rate  
376 estimated using both turbulent scales. We suggest investigating the dependence of dissipation rate  
377 on the scaled strain in the future to understand the dynamical processes involved (small scale shear  
378 layers, high-frequency waves, etc.). This points out the requirement for a specific parameterization  
379 of these high-frequency processes.

380 Our observation site is interesting to characterize the enhanced dissipation induced when the  
381 ISWs interact with the shore. Future observations study should focus on the NPS where the 3D  
382 hydrostatic model by Nagai and Hibiya (2015) (**Figure 1a**) shows the highest dissipation. A transect  
383 along the center of the Lombok Strait should also provide some insight into the transformation of  
384 the internal tides into ISW while avoiding the complex slope/solitary waves interaction process.

### 385 **Acknowledgment**

386 This observation was funded by the Indonesian Endowment Fund Scholarship Program  
387 (LPDP), Indonesian Ministry of Finance; the program of CNRS LEFE INSU “*Ondes solitaires et*

388 *turbulence dans le detroit de Lombok*” as well as the PHC Nusantara Project, a joint program  
389 between Campus France and Indonesian Ministry of Research and Technology (KEMENRISTEK);  
390 and Research Center for Oceanography-Indonesian Institute of Sciences (RCO-LIPI). We are also  
391 grateful to Kang Hendra Munandar and all staff of Marine Bio-industry Research Station RCO-LIPI  
392 Lombok for preparing the accommodations during the field research, and Pak Djatmiko Irianto for  
393 fully helping the data acquisitions. The publication of this research is supported by the funding  
394 scheme under the Deputy of Earth Sciences of the Indonesian Institute of Sciences (LIPI).

## 395 **Appendix : Thorpe and Ellison scale processing**

### 396 *Thorpe scale analysis*

397 Our first estimate of the dissipation rate is based on the classical Thorpe method, performed  
398 by examining the overturns in the density profiles. The Thorpe scale is defined as the RMS value of  
399 the displacements that one needs to apply to re-order an unstable density profile in a stable way  
400 (Ferron et al., 1998; Kunze et al., 2006; Park et al., 2014; Thorpe, 1977; Yang et al., 2014). Due to a  
401 large noise level observed from the unpumped CTD in the salinity profiles, we prefer to use the  
402 temperature-based estimates in this study, as in Mater et al. (2015). To this end, we need to assume  
403 that the potential temperature is a good proxy for the potential density. Here, it is a reasonable  
404 assumption since 1) considering the temperature and salinity range variation observed, the  
405 temperature-induced potential density variation is one order of magnitude larger ( $4 \text{ kg m}^{-3}$ ) than the  
406 salinity induced variation ( $0.4 \text{ kg m}^{-3}$ ), 2) the salinity and temperature variation both agree with to a  
407 potential density increase with depth down to about 150 m and 3) although the salinity slightly  
408 decreases below 150 m depth, its range of variation below this depth is very weak ( $\sim 0.03 \text{ PSU}$ )  
409 compared to the temperature variation ( $\sim 3^\circ\text{C}$ ) resulting in a factor of  $\sim 30$  between the temperature  
410 and salinity induced potential density variation. **Figure 5a** shows the temperature-salinity  
411 relationship inferred using the spline fit method with the goodness values are indicated. For the  
412 temperature, we set the noise level as 10 times the temperature sensor resolution of the SBE 19  
413 *Seacat Profiler*,  $\delta_T = 0.0001 \times 10 \text{ }^\circ\text{C}$ . Spurious noise-induced inversions were then eliminated using  
414 the intermediate profile method, firstly proposed by Ferron et al. (1998) and later modified by  
415 Gargett and Garner (2008). A method proposed by Gargett and Garner (2008) is also applied to  
416 reject a suspected false overturn, using the overturn ratio,  $R_o = \min(L^+/L, L^-/L)$ ; where  $L^+$  and  $L^-$  are  
417 respectively the vertical distances occupied by positive and negative Thorpe displacements within  
418 an overturn of vertical extent  $L$ . A single perfect overturn should contain equal lengths of positive  
419 and negative displacements, hence  $R_o = 0.5$ , and an imperfect overturn would have  $R_o$  less than 0.5.  
420 Mostly, false overturns are associated with values of  $R_o < 0.2$  (Gargett and Garner, 2008). Following

421 Dillon (1982), Ferron et al. (1998) and Moum (1996), we assume that  $L_T$  has a linear relationship  
 422 with the Ozmidov scale  $L_O = 0.8 L_T$ , where  $L_O = (\varepsilon N^{-3})^{1/2}$ , so that we get finally:

$$\varepsilon_{Th} = 0.64 L_T^2 N^3 \quad (1)$$

423 where  $N$  is the mean buoyancy frequency of the 2.5 m sliding mean in the overturn patch region,  
 424 defined as

$$N^2 = \frac{g}{\rho_0} \frac{d\rho}{dz} \quad (2)$$

425 The Thorpe scale method provides dissipation rate estimates when density overturns or  
 426 turbulent fluctuations overcome the noise level or spurious process contamination (surface waves,  
 427 mooring line motion). As a result, the statistics of dissipation rate are biased toward high values  
 428 when no assumption is made for the main portion of the water column where no density overturns  
 429 or turbulent fluctuations are detected. For this section, we assume a lower bound background  
 430 dissipation rate from a fine-scale parameterization for internal waves. Briefly, the TKE dissipation  
 431 rate is estimated as (Purwandana et al., 2020),

$$\varepsilon_{Th-GM} = \begin{cases} 0.64 L_T^2 N^3, & \text{when there is overturn} \\ \max \left( 1 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}, \varepsilon_0 \left( \frac{N^2}{N_0^2} \right) \right), & \text{when no overturn observed} \end{cases} \quad (5)$$

432  $1 \times 10^{-10}$  is the typical minimum dissipation rate observed by microstructure measurements in a quiet  
 433 region, the Banda Sea (Alford et al., 1999; Bouruet-Aubertot et al., 2018; Koch-Larrouy et al.,  
 434 2015);  $\varepsilon_0 = 7 \times 10^{-10} \text{ m}^2 \text{ s}^{-3}$  and  $N_0 = 3 \text{ cph}$  is the canonical Garret and Munk (hereinafter GM)  
 435 dissipation rate and buoyancy frequency reference, respectively;  $\varepsilon_0(N^2/N_0^2)$  is the GM-background  
 436 dissipation rate given by the fine-scale parameterization for internal waves (Gregg, 1989) yet  
 437 assuming that the local shear variance  $S_{10}^2$  is equal to the canonical shear variance  $S_{GM}^2$  due to lack  
 438 of shear measurements over the whole water column. This improved dissipation rate estimate that

439 takes into account the background dissipation rate is termed  $\epsilon_{\text{Th-GM}}$ , and provides continuous  
440 profiles of dissipation rates as opposed to the standard Thorpe method,  $\epsilon_{\text{Th}}$ .

#### 441 *Ellison scale analysis*

442 The high-resolution temporal resolution of the temperature time series allows us to use an  
443 alternative proxy of the dissipation rate that is based on the Ellison scale. It scales the vertical  
444 distance traveled by fluctuating density/temperature from its vertical gradient of the mean  
445 density/temperature (Moum, 1996). In our case, the Ellison scale is estimated from horizontal  
446 profiles of temperature time series hence omitting the influence of vertical oscillation due to internal  
447 and surface waves is important to diminish spurious turbulence.

448 First, as was done for the Thorpe scale, we assume that the potential temperature is a good  
449 proxy for the potential density and we define the Ellison scale ( $L_E$ ) as (Cimatoribus et al., 2014)

$$L_E = \overline{(\theta'^2)}^{1/2} / \left( \frac{d\bar{\theta}}{dz} \right) \quad (3)$$

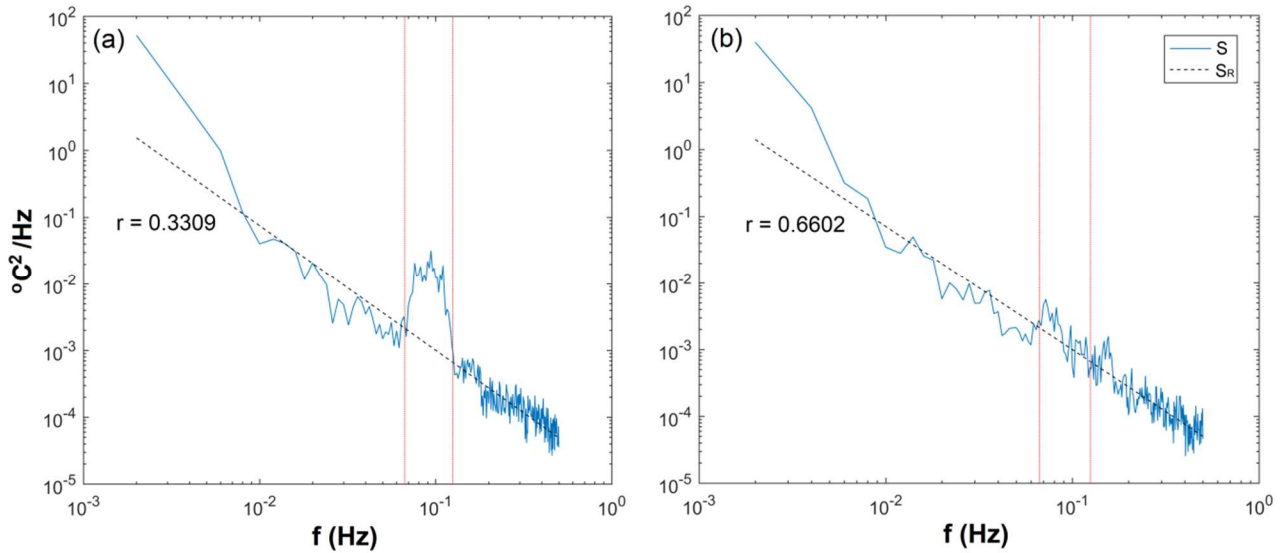
450 with  $\theta'$  is the potential temperature fluctuation around the (time) mean value. Note that the  
451 temperature time series is filtered to consider only the fluctuation expected to characterize turbulent  
452 events as described in detail as follows.

453 As in Cimatoribus et al. (2014), we define  $\theta'$  as the potential temperature turbulent  
454 fluctuation around the (time) mean value. In a well-controlled experiment, Cimatoribus et al. (2014)  
455 have compared the Thorpe and Ellison scales and demonstrated that they fall within a factor of two.  
456 If we want  $L_E$  to be representative of turbulent overturns length scale,  $\theta'$  must be representative of  
457 turbulent fluctuations, hence we need to define the mean value carefully to avoid internal wave  
458 contamination in the temperature fluctuations (Cimatoribus et al., 2014; Itsweire et al., 1986). This  
459 was done through a high pass wavelet filtering in Cimatoribus et al. (2014) with a cut-off frequency  
460  $f_c \cong 1/600$  Hz, here we chose to simply apply a high pass 8<sup>th</sup> order Butterworth filter with a much  
461 higher cut-off frequency  $f_c = 1/60$  Hz. We chose this value to smooth as much as possible



462 temperature fronts associated with the non-linear high-frequency internal waves packet while  
 463 preserving high-frequency turbulent fluctuations. The large ratio of 10 between the cut-off  
 464 frequency of Cimatoribus et al. (2014) and our study roughly matches the ratio of the buoyancy  
 465 frequency  $N$  in the two experiments, i.e.  $N \cong 1 \times 10^{-4} \text{ rad s}^{-1}$  in the low stratified environment of  
 466 Cimatoribus et al. (2014) and  $N \cong 2 \times 10^{-3} \text{ rad s}^{-1}$  in our strongly stratified environment. Our cut-off  
 467 value is also close to the lower bound defining the turbulent inertial-convective sub-range chosen by  
 468 Zhang and Moum (2010) on an equatorial mooring where the stratification is comparable to the  
 469 Lombok Strait stratification. Zhang and Moum (2010) outlined the contamination of the  
 470 temperature fluctuations by strong surface waves pumping of their mooring. They eliminated this  
 471 contamination using its signature on the cross-correlation between the accelerometers and the  
 472 temperature signal. During our experiment, the conditions were almost flat, yet a close examination  
 473 of the temperature signal and spectra revealed the occasional signature of contamination by surface  
 474 waves in the form of oscillation at a period of  $\sim 0.1$  Hz and an associated spectral peak. Lacking  
 475 accelerometers, we derive an empirical criterion to eliminate these portions of the signal. We  
 476 compute the spectral density  $S(f)$  of the de-trended  $\theta$  over 2000 s windows. The valuable frequency  
 477 range related to the turbulent fluctuations is set  $> 1/60$  Hz and we define the frequency range  
 478 possibly related to the contamination by the surface wave within a frequency of  $[1/15 - 1/8]$  Hz. A  
 479 reference fitted spectrum  $S_R(f)$  is then built by a power-law fitting of the temperature spectrum  $S(f)$   
 480 omitting points in the possibly contaminated range  $[1/15 - 1/8]$  Hz. Finally, the ratio  $r =$   
 481  $\int_{1/12}^{1/6} S(f)df / \int_{1/12}^{1/6} S_R(f)df$  of the variance of the signal in the surface waves frequency band  
 482 between the fitted and observed spectrum is computed.

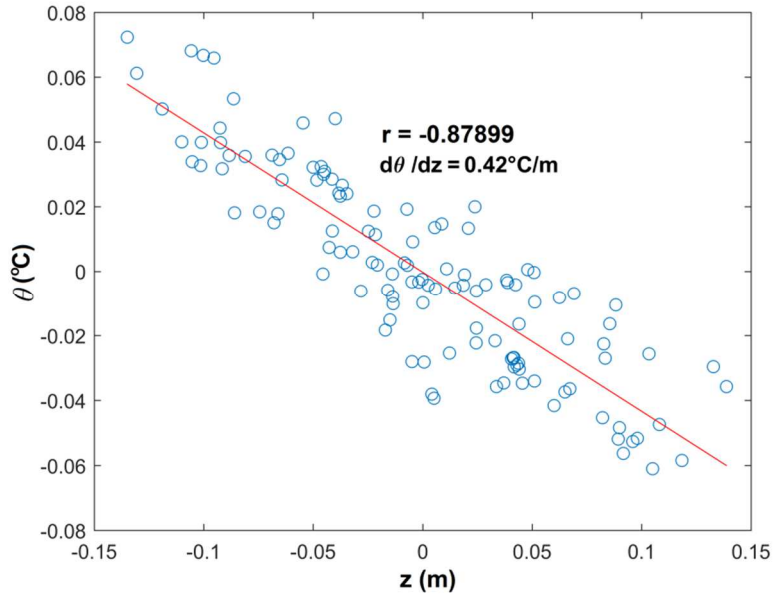
483 By visual inspection of the spectra  $S(f)$  and corresponding time series, we determined that  
 484 surface wave contamination was not detectable for  $r > 0.6$ , with this criteria, 20% of the data were  
 485 rejected. Figure A1 shows an example of the rejected spectrum and the accepted spectrum.



486

487 **Figure A1** Spectrum of temperature fluctuations showing (a) contaminated by surface waves ( $r =$   
 488 0.3309) and (b) uncontaminated by surface waves ( $r = 0.6602$ ). The black dashed line represents the  
 489 power-law fit  $S_R(f)$  (see text for details). The red dashed lines mark the  $[1/15, 1/8]$  Hz which is the  
 490 frequency band of possible surface waves.

491 An important source of error is the estimation of  $\overline{d\theta/dz}$ . Indeed, we estimate the mean  
 492 temperature gradient from the temperature difference between successive temperature sensors.  
 493 Since these sensors are separated by a large distance of 40 m on average, this estimate may not  
 494 represent the local gradient. Following the method of Moum et al. (2011), an estimation of this local  
 495 gradient can be obtained when surface waves dominate the signal over their frequency band  $B_W =$   
 496  $[1/15, 1/8]$  Hz. Indeed, in this case, most of the temperature fluctuations generated in the range  $B_W$   
 497 results from vertical oscillations of the temperature sensors induced by the surface wave pumping of  
 498 the mooring, as this is the case spectrum shown in **Figure A1a**. The local temperature gradient can  
 499 then be obtained by filtering the temperature and pressure/depth fluctuations in the range  $B_W$  and  
 500 fitting  $\theta$  against  $z$  to get the local slope  $\overline{d\theta/dz}$ . This was done for successive sequences of 120 s. An  
 501 example of such a fit is shown in **Figure A2**.

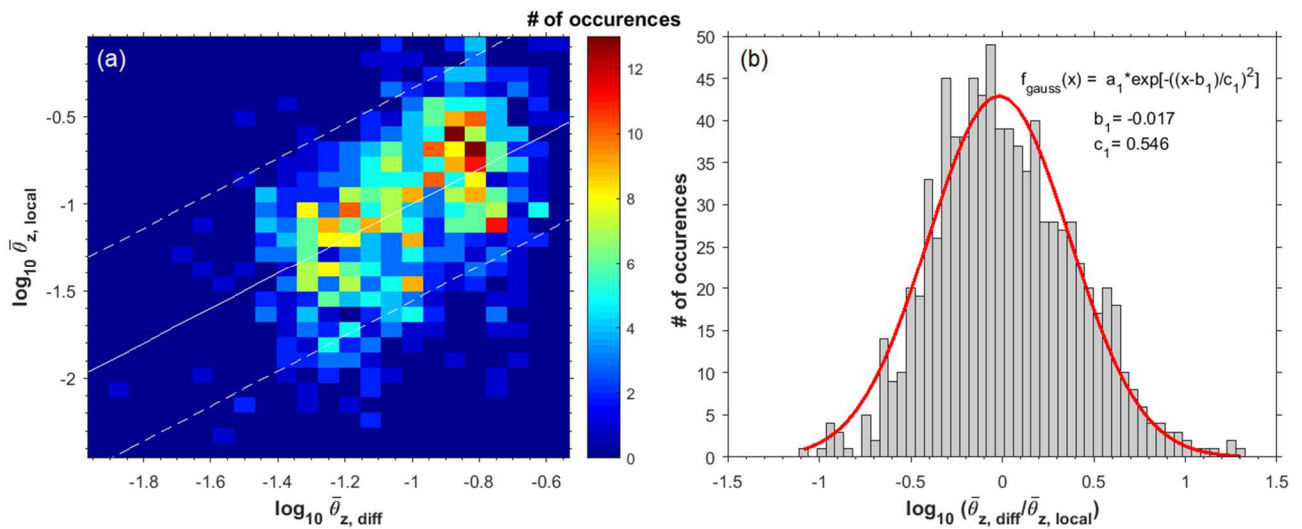


502

503 **Figure A2** Fluctuations of temperature in the band  $Bw = [1/15, 1/8]$  Hz as a function of the sensor  
 504 depth for a time sequence where  $Bw$  is dominated by surface waves.

505 This local estimate  $\bar{\theta}_{z,local}$  can only be computed when surface waves dominate the  
 506 frequency band  $Bw$ , that is  $\sim 20\%$  of the time, but it can be used to estimate the dispersion between  
 507 the local temperature gradient and the one based on the temperature difference between two sensors  
 508  $\bar{\theta}_{z,diff}$ . The comparison is shown in Figure A3. We can see that although there is an important  
 509 dispersion yet there is no systematic bias between the two estimates. We find that 90% of  $\bar{\theta}_{z,local}$   
 510 values are within  $[0.28, 4.6] \bar{\theta}_{z,diff}$  (marked by white dashed lines in Figure A3a). The dispersion  
 511 for the  $\log_{10}$  of the ratio  $x = \log_{10}(\bar{\theta}_{z,diff}/\bar{\theta}_{z,local})$  is close to Gaussian with a close to zero mean  
 512 value of  $x = -0.017$  and a standard deviation  $\sigma_x = 0.55$  (Figure A3b). In the following, we use this  
 513 Gaussian fit to compute the confidence interval on  $\bar{\theta}_{z,diff}$  and the resulting error on the  
 514 computation of the Ellison scale and derived dissipation rate. Then, the Ellison scale,  $L_E$  is  
 515 calculated using Eq. (3), and eventually, we determine the turbulent dissipation rate as:

$$\varepsilon_E = L_E^2 N^3 \quad (4)$$



516

517 **Figure A3** (a) Number of occurrences plot in bins of  $\log_{10} \bar{\theta}_{z,diff}$  and  $\log_{10} \bar{\theta}_{z,local}$ , (b) distribution  
 518 of  $x = \log_{10}(\bar{\theta}_{z,diff}/\bar{\theta}_{z,local})$  and its Gaussian fit. The white dashed lines in (a) represent 90% of  
 519  $\bar{\theta}_{z,local}$  values are within  $[0.28, 4.6] \bar{\theta}_{z,diff}$ .

520 **References**

521 Aiki, H., Matthews, J.P., Lamb, K.G., 2011. Modeling and energetics of tidally generated wave  
 522 trains in the Lombok Strait: Impact of the Indonesian Throughflow. *J. Geophys. Res. Ocean.*  
 523 116. doi:10.1029/2010JC006589

524 Alford, M.H., Gregg, M.C., Ilyas, M., 1999. Diapycnal mixing in the Banda Sea: Results of the first  
 525 microstructure measurements in the Indonesian throughflow. *Geophys. Res. Lett.* 26, 2741–  
 526 2744. doi:10.1029/1999GL002337

527 Alford, M.H., Peacock, T., MacKinnon, J.A., Nash, J.D., Buijsman, M.C., Centurioni, L.R., Chao,  
 528 S.-Y., Chang, M.-H., Farmer, D.M., Fringer, O.B., Fu, K.-H., Gallacher, P.C., Graber, H.C.,  
 529 Helfrich, K.R., Jachec, S.M., Jackson, C.R., Klymak, J.M., Ko, D.S., Jan, S., Johnston, T.M.S.,  
 530 Legg, S., Lee, I.-H., Lien, R.-C., Mercier, M.J., Moum, J.N., Musgrave, R., Park, J.-H.,  
 531 Pickering, A.I., Pinkel, R., Rainville, L., Ramp, S.R., Rudnick, D.L., Sarkar, S., Scotti, A.,  
 532 Simmons, H.L., St Laurent, L.C., Venayagamoorthy, S.K., Wang, Y.-H., Wang, J., Yang, Y.J.,  
 533 Paluszkiwicz, T., (David) Tang, T.-Y., 2015. The formation and fate of internal waves in the

- 534 South China Sea. *Nature* 521, 65–69. doi:10.1038/nature14399
- 535 Apel, J.R., Holbrook, J.R., Liu, A.K., Tsai, J.J., 1985. The Sulu Sea Internal Soliton Experiment. *J.*  
536 *Phys. Oceanogr.* doi:10.1175/1520-0485(1985)015<1625:TSSISE>2.0.CO;2
- 537 Astawa Karang, I.W.G., Nishio, F., Mitnik, L., Osawa, T., 2012. Spatial-Temporal Distribution and  
538 Characteristics of Internal Waves in the Lombok Strait Area Studied by Alos-Palsar Images.  
539 *Earth Sci. Res.* 1, 11–22. doi:10.5539/esr.v1n2p11
- 540 Bourgault, D., Morsilli, M., Richards, C., Neumeier, U., Kelley, D.E., 2014. Sediment resuspension  
541 and nepheloid layers induced by long internal solitary waves shoaling orthogonally on uniform  
542 slopes. *Cont. Shelf Res.* 72, 21–33. doi:10.1016/j.csr.2013.10.019
- 543 Bouruet-Aubertot, P., Cuypers, Y., Ferron, B., Dausse, D., Ménage, O., Atmadipoera, A., Jaya, I.,  
544 Olivier, M., Atmadipoera, A., Jaya, I., 2018. Contrasted turbulence intensities in the Indonesian  
545 Throughflow: a challenge for parameterizing energy dissipation rate. *Ocean Dyn.* 68, 1–75.  
546 doi:10.1007/s10236-018-1159-3
- 547 Brandt, P., Romeiser, R., Rubino, A., 1999. On the determination of characteristics of the interior  
548 ocean dynamics from radar signatures of internal solitary waves. *J. Geophys. Res. Ocean.* 104,  
549 30039–30045. doi:10.1029/1999JC900092
- 550 Cimadoribus, A.A., Van Haren, H., Gostiaux, L., 2014. Comparison of Ellison and Thorpe scales  
551 from Eulerian ocean temperature observations. *J. Geophys. Res. Ocean.* 119, 7047–7065.  
552 doi:10.1002/2014JC010132
- 553 Dillon, T.M., 1982. Vertical overturns: A comparison of Thorpe and Ozmidov length scales. *J.*  
554 *Geophys. Res.* 87, 9601. doi:10.1029/JC087iC12p09601
- 555 Egbert, G.D., Erofeeva, S.Y., 2002. Efficient Inverse Modeling of Barotropic Ocean Tides. *J.*  
556 *Atmos. Ocean. Technol.* 19, 183–204. doi:10.1175/1520-  
557 0426(2002)019<0183:EIMOBO>2.0.CO;2

- 558 Ferron, B., Mercier, H., Speer, K., Gargett, A., Polzin, K., 1998. Mixing in the Romanche Fracture  
559 Zone. *J. Phys. Oceanogr.* 28, 1929–1945. doi:10.1175/1520-  
560 0485(1998)028<1929:MITRFZ>2.0.CO;2
- 561 Ffield, A., Gordon, A.L., 1996. Tidal mixing signatures in the Indonesian seas. *J. Phys. Oceanogr.*  
562 doi:10.1175/1520-0485(1996)026<1924:TMSITI>2.0.CO;2
- 563 Ffield, A., Robertson, R., 2008. Temperature finestructure in the Indonesian seas. *J. Geophys. Res.*  
564 *Ocean.* 113, 1–19. doi:10.1029/2006JC003864
- 565 Gargett, A., Garner, T., 2008. Determining Thorpe scales from ship-lowered CTD density profiles.  
566 *J. Atmos. Ocean. Technol.* 25, 1657–1670. doi:10.1175/2008JTECHO541.1
- 567 Gerkema, T., Zimmerman, J.T.F., 1995. Generation of Nonlinear Internal Tides and Solitary Waves.  
568 *J. Phys. Oceanogr.* 25, 1081–1094. doi:10.1175/1520-  
569 0485(1995)025<1081:GONITA>2.0.CO;2
- 570 Gordon, A.L., 2005. Oceanography of the Indonesian Seas and their throughflow. *Oceanography*  
571 18, 14–27. doi:10.5670/oceanog.2005.18
- 572 Gordon, A.L., Fine, R.A., 1996. Pathways of water between the Pacific and Indian oceans in the  
573 Indonesian seas. *Nature* 379, 146–149. doi:10.1038/379146a0
- 574 Gregg, M.C., 1989. Scaling turbulent dissipation in the thermocline. *J. Geophys. Res. Ocean.* 94,  
575 9686–9698. doi:10.1029/JC094iC07p09686
- 576 Hatayama, T., 2004. Transformation of the Indonesian throughflow water by vertical mixing and its  
577 relation to tidally generated internal waves. *J. Oceanogr.* 60, 569–585.  
578 doi:10.1023/B:JOCE.0000038350.32155.cb
- 579 Hautala, S.L., Reid, J.L., Bray, N., 1996. The distribution and mixing of Pacific water masses in the  
580 Indonesian Seas. *J. Geophys. Res.* 101, 12,375–12,389. doi:10.1029/96JC00037
- 581 Hautala, S.L., Sprintall, J., Potemra, J.T., Chong, J.C., Pandoe, W., Bray, N., Ilahude, a. G., 2001.

582 Velocity structure and transport of the Indonesian Throughflow in the major straits restricting  
583 flow into the Indian Ocean. *J. Geophys. Res.* 106, 19527. doi:10.1029/2000JC000577

584 Holloway, P., Pelinovsky, E., Talipova, T., 2002. Internal Tide Transformation and Oceanic Internal  
585 Solitary Waves BT - Environmental Stratified Flows, in: Grimshaw, R. (Ed.), . Springer US,  
586 Boston, MA, pp. 29–60. doi:10.1007/0-306-48024-7\_2

587 Itsweire, E.C., Helland, K.N., Van Atta, C.W., 1986. The evolution of grid-generated turbulence in a  
588 stably stratified fluid. *J. Fluid Mech.* 162, 299–338. doi:10.1017/S0022112086002069

589 Jackson, C., 2007. Internal wave detection using the Moderate Resolution Imaging  
590 Spectroradiometer (MODIS). *J. Geophys. Res.* 112, C11012. doi:10.1029/2007JC004220

591 Klymak, J.M., Pinkel, R., Rainville, L., 2008. Direct Breaking of the Internal Tide near Topography:  
592 Kaena Ridge, Hawaii. *J. Phys. Oceanogr.* 38, 380–399. doi:10.1175/2007JPO3728.1

593 Koch-Larrouy, A., Atmadipoera, A., van Beek, P., Madec, G., Aucan, J., Lyard, F., Grelet, J.,  
594 Souhaut, M., 2015. Estimates of tidal mixing in the Indonesian archipelago from  
595 multidisciplinary INDOMIX in-situ data. *Deep. Res. Part I Oceanogr. Res. Pap.* 106, 136–153.  
596 doi:10.1016/j.dsr.2015.09.007

597 Koch-Larrouy, A., Lengaigne, M., Terray, P., Madec, G., Masson, S., 2010. Tidal mixing in the  
598 Indonesian seas and its effect on the tropical climate system. *Clim. Dyn.* 34, 891–904.  
599 doi:10.1007/s00382-009-0642-4

600 Koch-Larrouy, A., Madec, G., Bouruet-Aubertot, P., Gerkema, T., Bessières, L., Molcard, R., 2007.  
601 On the transformation of Pacific Water into Indonesian Throughflow Water by internal tidal  
602 mixing. *Geophys. Res. Lett.* 34. doi:10.1029/2006GL028405

603 Kunze, E., Firing, E., Hummon, J.M., Chereskin, T.K., Thurnherr, A.M., 2006. Global Abyssal  
604 Mixing Inferred from Lowered ADCP Shear and CTD Strain Profiles. *J. Phys. Oceanogr.* 36,  
605 1553–1576. doi:10.1175/JPO2926.1

- 606 Lee, T., Fukumori, I., Menemenlis, D., Xing, Z.F., Fu, L.-L., 2002. Effects of the Indonesian  
607 Throughflow on the Pacific and Indian oceans. *J. Phys. Oceanogr.* 32, 1404–1429.  
608 doi:10.1175/1520-0485(2002)032<1404:Eotito>2.0.Co;2
- 609 Mater, B.D., Venayagamoorthy, S.K., St. Laurent, L., Moum, J.N., 2015. Biases in Thorpe-Scale  
610 Estimates of Turbulence Dissipation. Part I: Assessments from Large-Scale OvertURNS in  
611 Oceanographic Data. *J. Phys. Oceanogr.* 45, 2497–2521. doi:10.1175/JPO-D-14-0128.1
- 612 Matthews, J.P., Aiki, H., Masuda, S., Awaji, T., Ishikawa, Y., 2011. Monsoon regulation of Lombok  
613 Strait internal waves. *J. Geophys. Res. Ocean.* 116, 1–14. doi:10.1029/2010JC006403
- 614 Maxworthy, T., 1979. A note on the internal solitary waves produced by tidal flow over a three-  
615 dimensional ridge. *J. Geophys. Res.* 84. doi:10.1029/JC084iC01p00338
- 616 Moum, J.N., 1996. Energy-containing scales of turbulence in the ocean thermocline. *J. Geophys.*  
617 *Res. Ocean.* 101, 14095–14109. doi:10.1029/96JC00507
- 618 Moum, J.N., Nash, J.D., Smyth, W.D., 2011. Narrowband oscillations in the upper equatorial  
619 Ocean. Part I: Interpretation as shear instabilities. *J. Phys. Oceanogr.* 41, 397–411.  
620 doi:10.1175/2010JPO4450.1
- 621 Murray, S.P., Arief, D., 1988. Throughflow into the Indian Ocean through the Lombok Strait,  
622 January 1985–January 1986. *Nature* 333, 444–447. doi:10.1038/333444a0
- 623 Nagai, T., Hibiya, T., 2015. Internal tides and associated vertical mixing in the Indonesian  
624 Archipelago. *J. Geophys. Res. C Ocean.* 3373–3390. doi:10.1002/2014JC010592
- 625 Nagai, T., Hibiya, T., Bouruet-Aubertot, P., 2017. Nonhydrostatic Simulations of Tide-Induced  
626 Mixing in the Halmahera Sea: A Possible Role in the Transformation of the Indonesian  
627 Throughflow Waters. *J. Geophys. Res. Ocean.* 122, 8933–8943. doi:10.1002/2017JC013381
- 628 Nugroho, D., Koch-larrouy, A., Gaspar, P., Lyard, F., Re, G., Tranchant, B., 2018. Modelling  
629 explicit tides in the Indonesian seas : An important process for surface sea water properties.



630 Mar. Pollut. Bull. doi:10.1016/j.marpolbul.2017.06.033

631 Park, Y.H., Lee, J.H., Durand, I., Hong, C.S., 2014. Validation of Thorpe-scale-derived vertical  
632 diffusivities against microstructure measurements in the Kerguelen region. *Biogeosciences* 11,  
633 6927–6937. doi:10.5194/bg-11-6927-2014

634 Polzin, K.L., Naveira Garabato, A.C., Huussen, T.N., Sloyan, B.M., Waterman, S., 2014. Finescale  
635 parameterizations of turbulent dissipation. *J. Geophys. Res. Ocean.* 119, 1383–1419.  
636 doi:10.1002/2013JC008979

637 Potemra, J.T., Hautala, S., Sprintall, J., Pandoe, W., 2002. Interaction between the Indonesian Seas  
638 and the Indian Ocean in Observations and Numerical Models \*. *J. Phys. Oceanogr.* 32, 1838–  
639 1854. doi:10.1175/1520-0485(2002)032<1838:IBTISA>2.0.CO;2

640 Purwandana, A., Cuypers, Y., Bouruet-Aubertot, P., Nagai, T., Hibiya, T., Atmadipoera, A.S.A.S.,  
641 2020. Spatial structure of turbulent mixing inferred from historical CTD datasets in the  
642 Indonesian seas. *Prog. Oceanogr.* 184, 102312. doi:10.1016/j.pocean.2020.102312

643 Ray, R.D., Susanto, R.D., 2016. Tidal mixing signatures in the Indonesian seas from high-resolution  
644 sea surface temperature data. *Geophys. Res. Lett.* 43, 8115–8123. doi:10.1002/2016GL069485

645 Robertson, R., Field, A., 2005. M2 Baroclinic Tides in the Indonesian Seas. *Oceanography* 18, 62–  
646 73. doi:10.5670/oceanog.2005.06

647 Schiller, A., 2004. Effects of explicit tidal forcing in an OGCM on the water-mass structure and  
648 circulation in the Indonesian throughflow region. *Ocean Model.* 6, 31–49. doi:10.1016/S1463-  
649 5003(02)00057-4

650 Scotti, A., Beardsley, R.C., Butman, B., 2007. Generation and propagation of nonlinear internal  
651 waves in Massachusetts Bay 112, 1–19. doi:10.1029/2007JC004313

652 Sprintall, J., Chong, J., Syamsudin, F., Morawitz, W., Hautala, S., Bray, N., Wijffels, S., 1999.  
653 Dynamics of the South Java Current in the Indo-Australian Basin. *Geophys. Res. Lett.* 26,

- 654 2493–2496. doi:10.1029/1999GL002320
- 655 Sprintall, J., Potemra, J.T., Hautala, S.L., Bray, N.A., Pandoe, W.W., 2003. Temperature and salinity  
656 variability in the exit passages of the Indonesian Throughflow. *Deep. Res. Part II Top. Stud.*  
657 *Oceanogr.* 50, 2183–2204. doi:10.1016/S0967-0645(03)00052-3
- 658 Susanto, R.D., Gordon, L., Zheng, Q., 2001. Upwelling along the coasts of Java and Sumatra and its  
659 relation to ENSO. *Geophys. Res. Lett.* 28, 1599–1602. doi:10.1029/2000GL011844
- 660 Susanto, R.D., Mitnik, L., Zheng, Q., 2005. Ocean Internal Waves Observed in the Lombok Strait.  
661 *Oceanography* 18, 80–87. doi:10.5670/oceanog.2005.08
- 662 Thorpe, S.A., 1977. Turbulence and Mixing in a Scottish Loch. *Philos. Trans. R. Soc. A Math. Phys.*  
663 *Eng. Sci.* 286, 125–181. doi:10.1098/rsta.1977.0112
- 664 Vlasenko, V., Sanchez Garrido, J.C., Stashchuk, N., Garcia Lafuente, J., Losada, M., 2009. Three-  
665 Dimensional Evolution of Large-Amplitude Internal Waves in the Strait of Gibraltar. *J. Phys.*  
666 *Oceanogr.* 39, 2230–2246. doi:10.1175/2009jpo4007.1
- 667 Wijffels, S., Sprintall, J., Fieux, M., Bray, N., 2002. The JADE and WOCE I10 / IR6 Throughflow  
668 sections in the southeast Indian Ocean. Part 1: water mass distribution and variability. *Deep*  
669 *Sea Res. Part II Top. Stud. Oceanogr.* 49, 1341–1362. doi:10.1016/S0967-0645(01)00155-2
- 670 Wyrтки, K., 1961. Physical oceanography of the Southeast Asian waters., *Scientific Results of*  
671 *Marine Investigations of the South China Sea and the Gulf of Thailand.*  
672 doi:10.1017/S0025315400054370
- 673 Xie, X.H., Cuypers, Y., Bouruet-Aubertot, P., Ferron, B., Pichon, A., Lourenço, A., Cortes, N.,  
674 2013. Large-amplitude internal tides, solitary waves, and turbulence in the central Bay of  
675 Biscay. *Geophys. Res. Lett.* 40, 2748–2754. doi:10.1002/grl.50533
- 676 Yang, Q., Zhao, W., Li, M., Tian, J., 2014. Spatial Structure of Turbulent Mixing in the  
677 Northwestern Pacific Ocean. *J. Phys. Oceanogr.* 44, 2235–2247. doi:10.1175/JPO-D-13-0148.1

678 Zhang, Y., Moum, J.N., 2010. Inertial-Convective subrange estimates of thermal variance  
679 dissipation rate from moored temperature measurements. *J. Atmos. Ocean. Technol.* 27, 1950–  
680 1959. doi:10.1175/2010JTECHO746.1