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

3 **Adi Purwandana^{1,2)}, Yannis Cuypers¹⁾, Pascale Bouruet-Aubertot¹⁾**

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

6 ²⁾*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 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 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° S, 116.02° 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

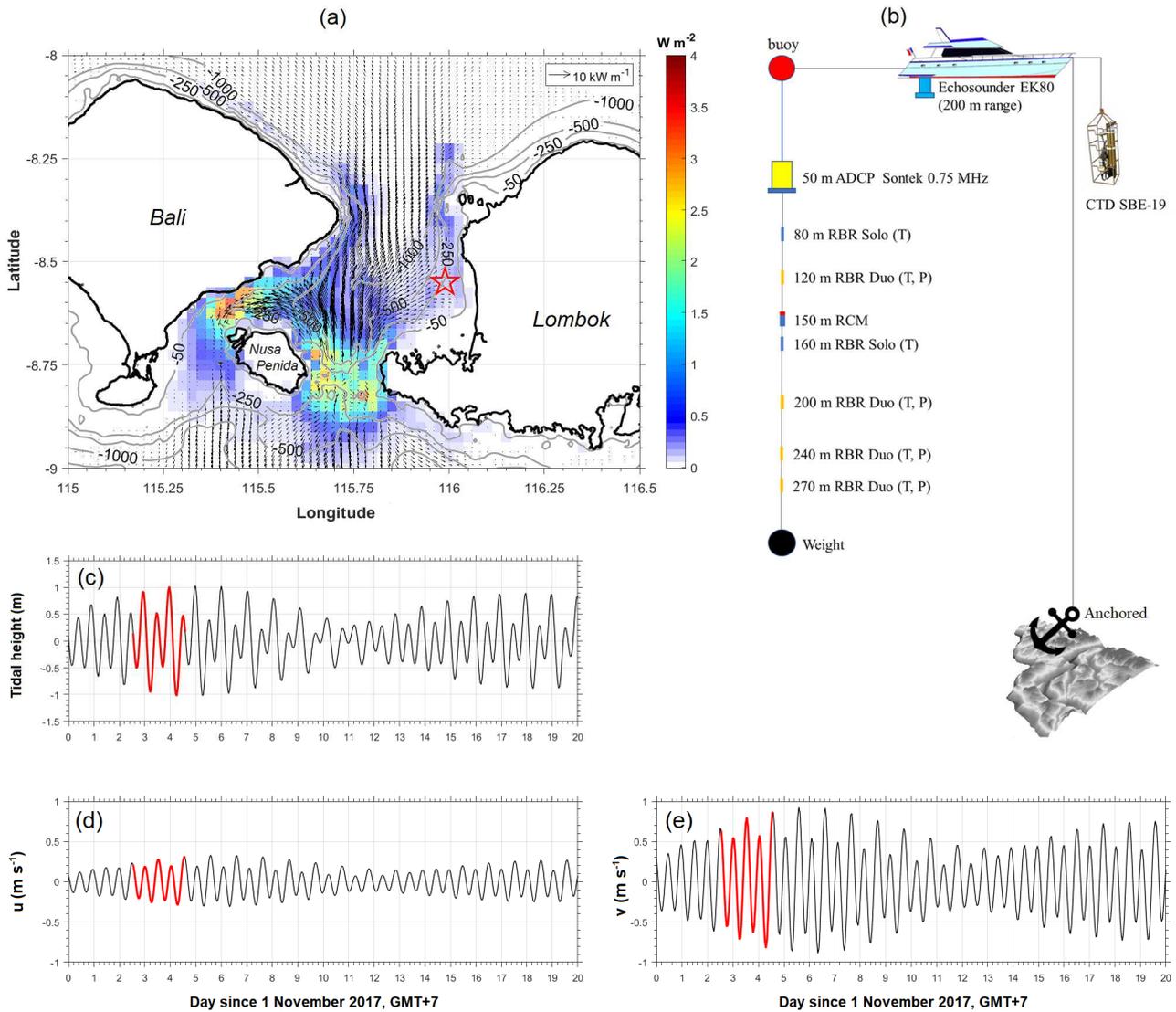
130 with the shore but does not allow to characterize the most coherent unaltered ISW structure near
 131 their generation site. A detailed description of the measurements is shown in **Table 1**.

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 rd , 2017	15.23; 18.18; 19.28; 21.24; 23.38	
November 4 th , 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 th , 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° 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 ε 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.

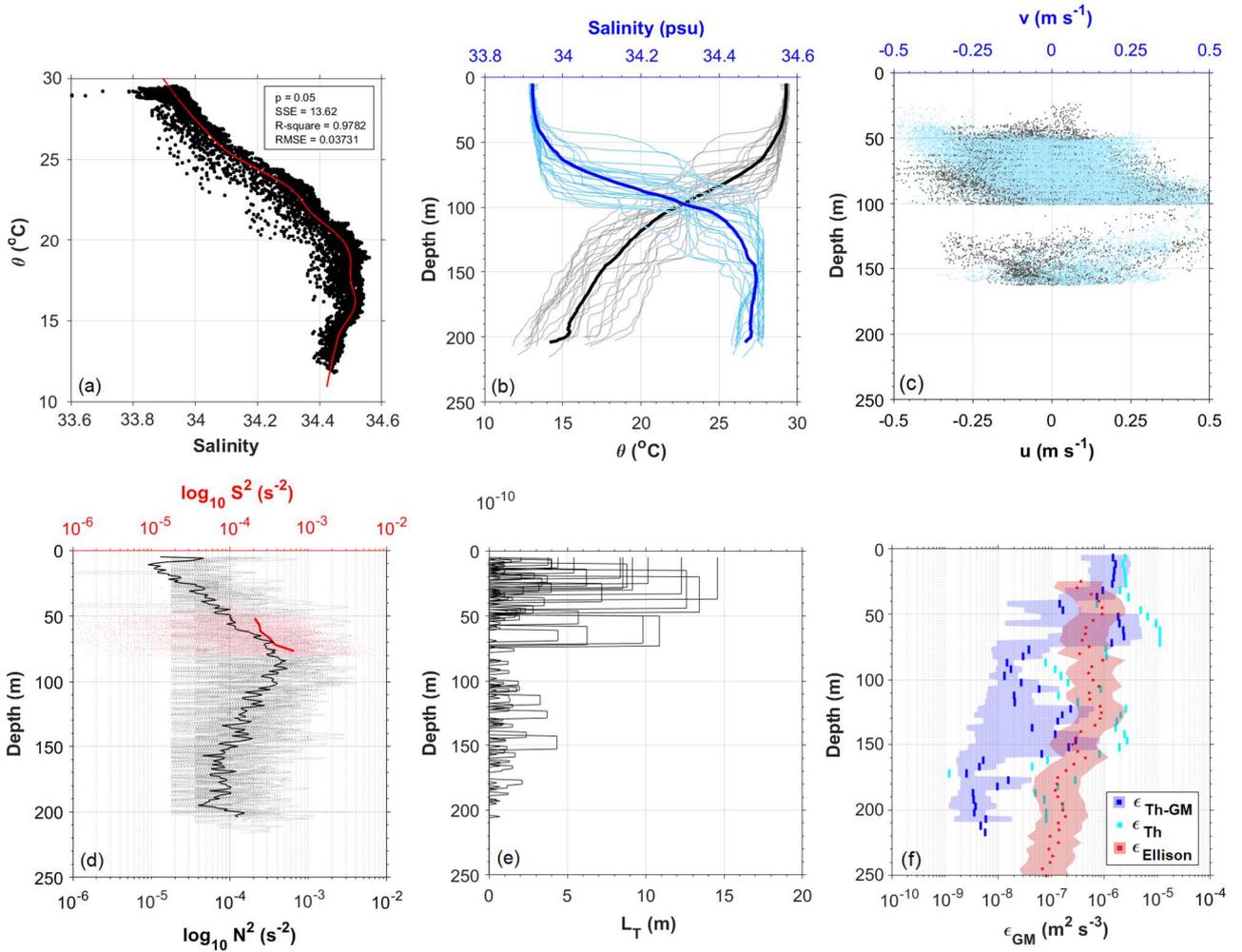
179 The Ellison method estimation of the dissipation rate was found comparable to the
180 microstructure measurements direct estimates according to a study by (Moum, 1996); yet as stated
181 by Cimadoribus et al. (2014), the Ellison scale may overestimate/underestimate the real turbulence
182 dissipation 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.

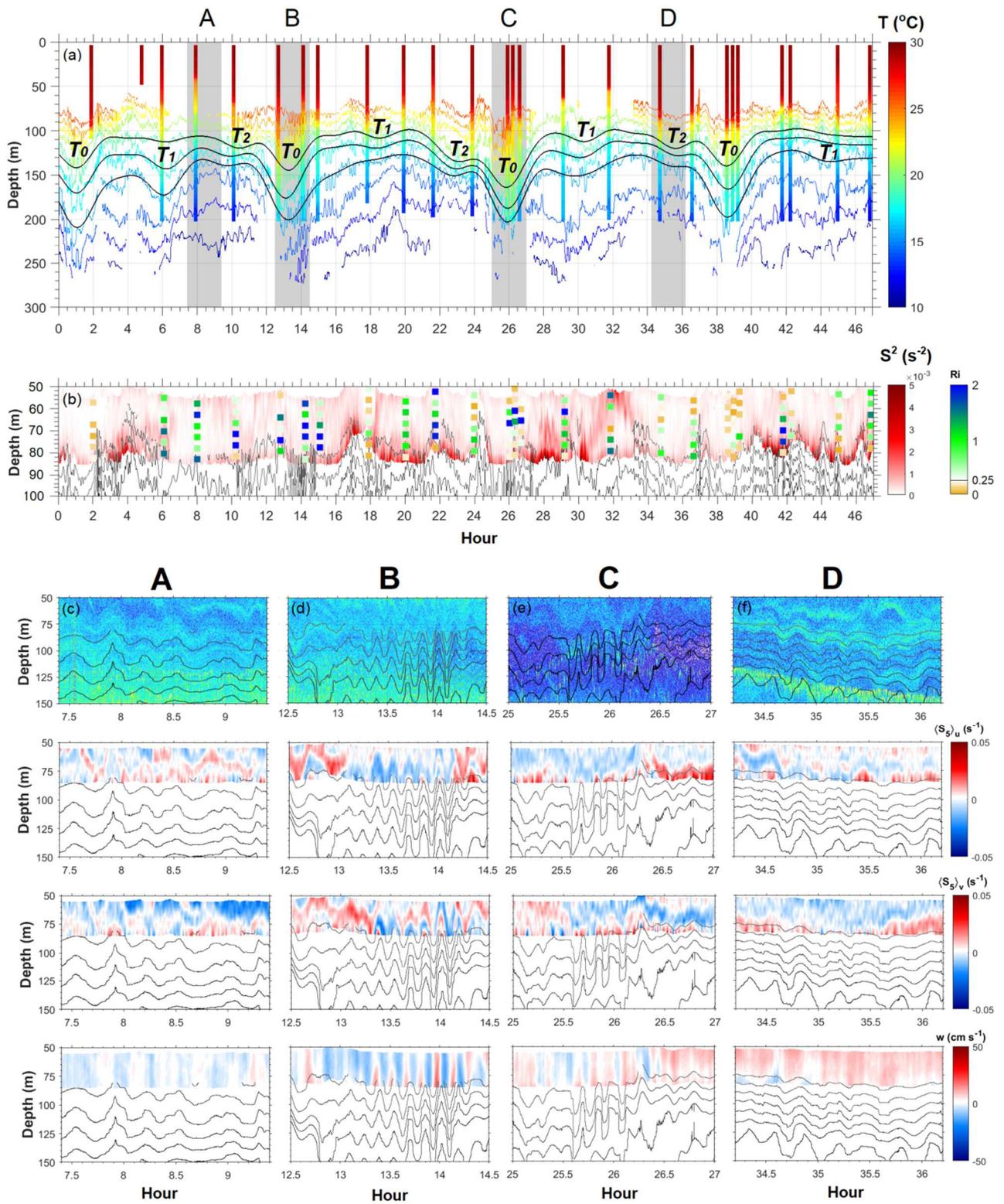
215 The time series of temperature recorded from the mooring sensors show a clear non-linear
 216 semi-diurnal signal characterized by semi-diurnal periodic large troughs (**Figure 3a**). The peak to
 217 peak amplitude reaches ~ 90 m for the 20°C isotherm located in the pycnocline, which is larger than

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 M₂ 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 (M₄, M₆) 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⁻¹ during the period we sampled, while the mode-1 phase speed, c_1 is
227 typically of 1 m s⁻¹. 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⁻¹. 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

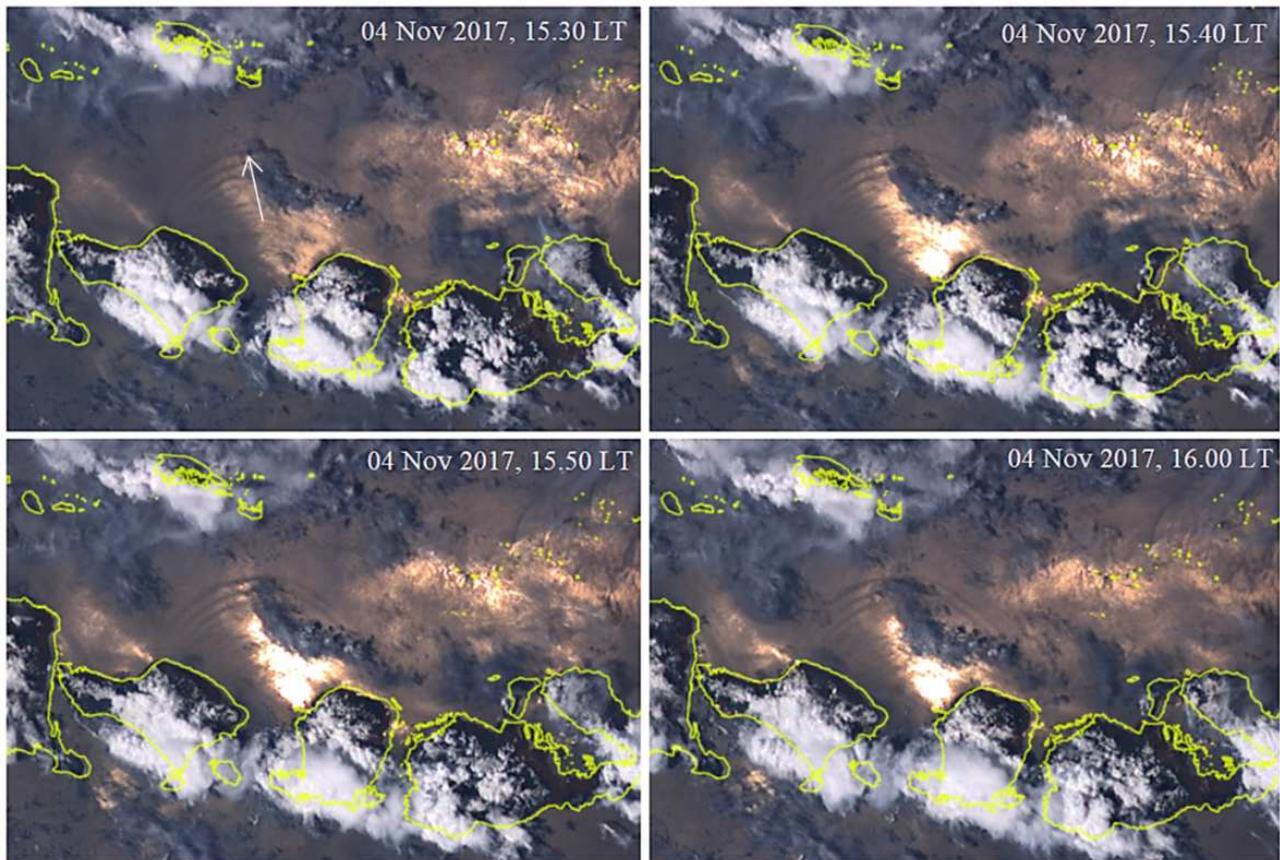
250 **Figure 3** (a) Isotherm contours inferred from RBR sensors and temperature profiles from CTD
 251 casts. The black line is low pass filtered ($>3h$) for isotherms 18, 20, and 22° C. (b) The squared
 252 shear with isotherms and Richardson number, Ri superimposed. (c, d, e, f) The enlarged view 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*² 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, 4th 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.



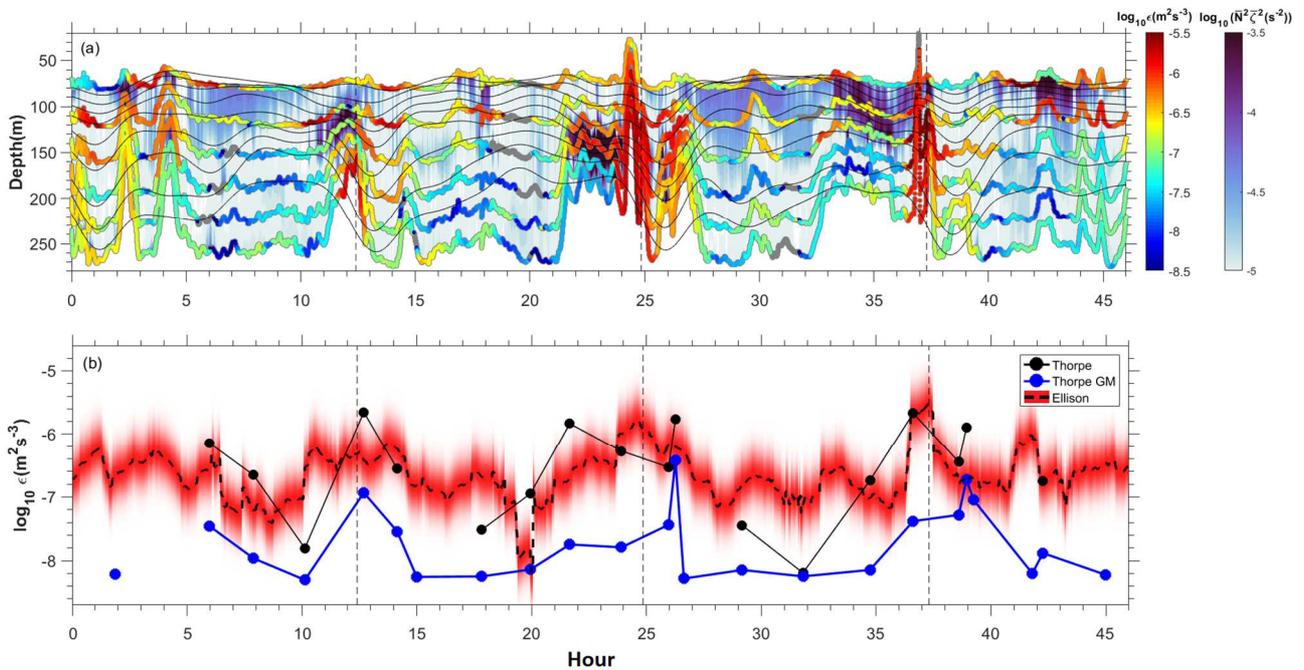
283

284 **Figure 4** The observed ISWs signature (white arrow) by satellite images Himawari-8 on November
285 4th, 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 (ϵ_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 ε_{Th} and ε_{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 ~ 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 (~ 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 ~ 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.

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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 kg m^{-3}) than the
406 salinity induced variation (0.4 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 ~ 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*, $\delta_T = 0.0001 \times 10^\circ\text{C}$. 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 1×10^{-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, ϵ_{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)

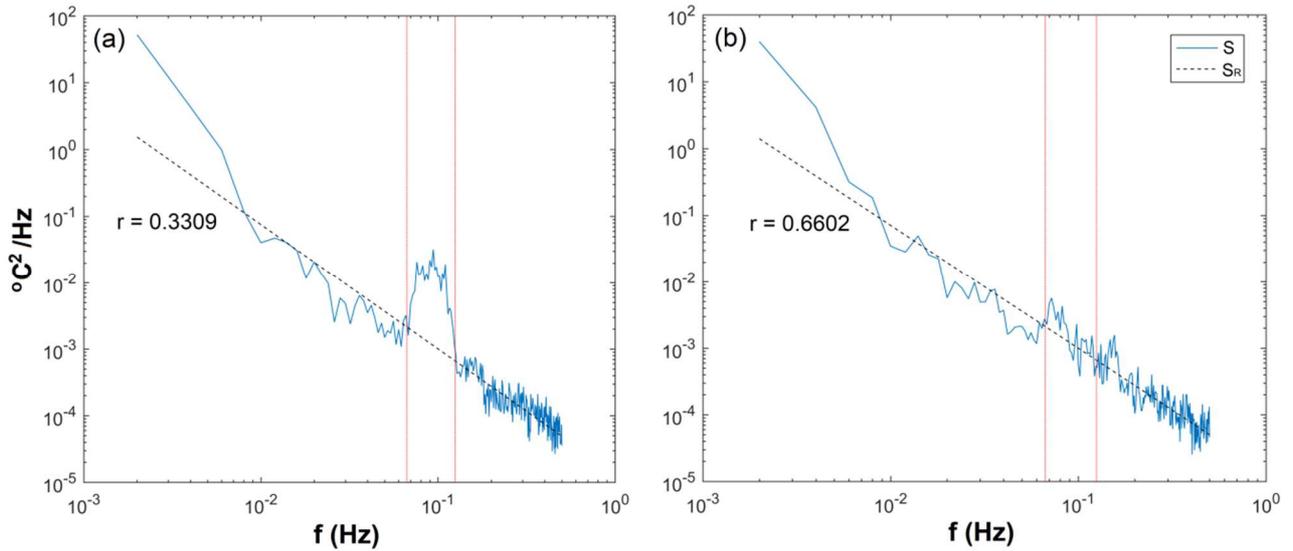
$$L_E = \overline{(\theta'^2)^{1/2}} / \left(\frac{d\bar{\theta}}{dz} \right) \quad (3)$$

450 with θ' 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 θ' 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, θ' 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 8th 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 ~ 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 θ 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.

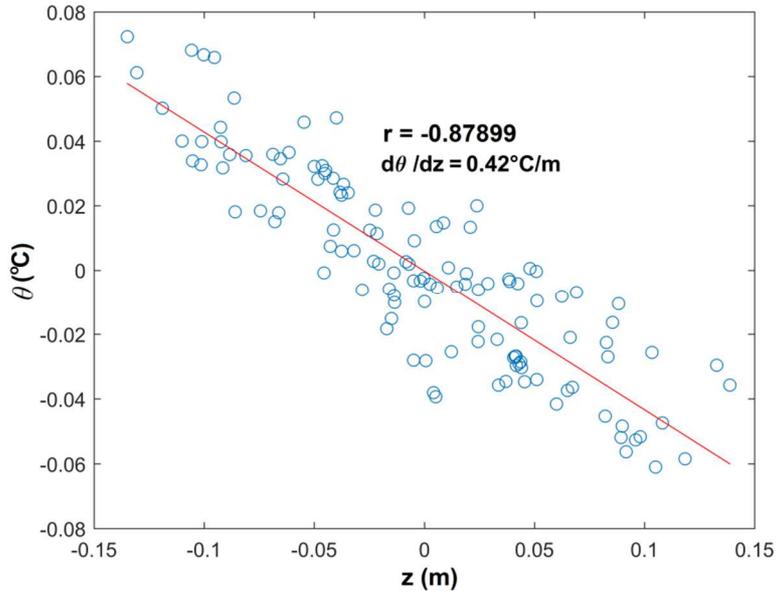
483 By visual inspection of the spectra $S(f)$ and corresponding time series, we determined that
 484 surface wave contamination was not detectable for $r > 0.6$, with this criteria, 20% of the data were
 485 rejected. Figure A1 shows an example of the rejected spectrum and the accepted spectrum.



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 θ 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**.

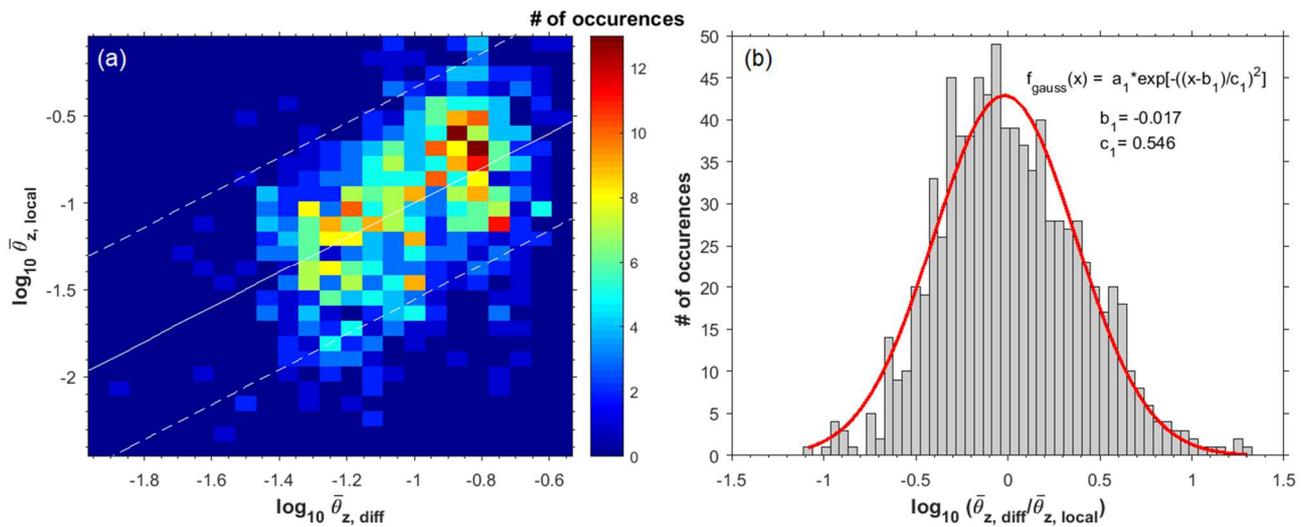


502

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 $[0.28, 4.6] \bar{\theta}_{z,diff}$.

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