



HAL
open science

Deep circulation driven by strong vertical mixing in the Timor Basin

Yannis Cuypers, Stéphane Pous, Janet Sprintall, Agus Atmadipoera, Gurvan
Madec, Robert Molcard

► **To cite this version:**

Yannis Cuypers, Stéphane Pous, Janet Sprintall, Agus Atmadipoera, Gurvan Madec, et al.. Deep circulation driven by strong vertical mixing in the Timor Basin. *Ocean Dynamics*, 2017, 67 (2), pp.191-209. 10.1007/s10236-016-1019-y . hal-01430803

HAL Id: hal-01430803

<https://hal.sorbonne-universite.fr/hal-01430803v1>

Submitted on 10 Jan 2017

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.

Deep circulation driven by strong vertical mixing in the Timor basin

Y. Cuypers¹, S. Pous¹, J. Sprintall², A. Atmadipoera³, G. Madec¹ and R. Molcard¹.

1. Sorbonne Universités (UPMC, Univ Paris 06)-CNRS-IRD-MNHN, LOCEAN Laboratory, 4 place Jussieu, F-75005 Paris, France

2 Scripps Institution of Oceanography, La Jolla, California

3 Department of Marine Sciences and Technology, Bogor Agricultural University, Indonesia

Abstract

The importance of deep mixing in driving the deep part of the overturning circulation has been a long debated question at the global scale. Our observations provide an illustration of this process at the Timor basin scale of ~ 1000 km. Long-term averaged moored velocity data at the Timor western sill suggest that a deep circulation is present in the Timor Basin. An inflow transport of ~0.15 Sv is observed between 1600 m and the bottom at 1890 m. Since the basin is closed on its eastern side below 1250 m depth, a return flow must be generated above 1600 m with a ~0.15 Sv outflow. The vertical turbulent diffusivity is inferred from a heat and transport balance at the basin scale and from Thorpe scale analysis. Basin averaged vertical diffusivity is as large as $1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$. Observations are compared with regional low resolution numerical simulations, and the deep observed circulation is only recovered when a strong vertical diffusivity resulting from the parameterization of internal tidal mixing is considered. Furthermore, the deep vertical mixing appears to be strongly dependent on the choice of the internal tide mixing parameterization and also on the prescribed value of the mixing efficiency.

Corresponding author: Yannis Cuypers Yannis.Cuypers@locean-ipsl.upmc.fr

Laboratoire d'Océanographie et du Climat: Expérimentations et approches numériques.

Unité Mixte de Recherche 7159 CNRS / IRD / Université Pierre et Marie Curie/MNHN.

Institut Pierre Simon Laplace.

Boîte 100 - 4, place Jussieu 75252 PARIS Cedex 05.

1. Introduction

The necessity for the deep ocean to mix with the upper layer to close the meridional overturning circulation requires, according to numerical model and bulk budget estimations, an average abyssal vertical diapycnal diffusivity K_z of $10^{-4} \text{ m}^2 \text{ s}^{-1}$ [Munk and Wunsch, 1998]. However experimental measurements from microstructure instruments [Osborn, 1980] and tracers [Ledwell et al., 2000] show that open ocean diapycnal diffusivity is an

31 order of magnitude smaller ($10^{-5}\text{m}^2\text{ s}^{-1}$). Fine scale parameterizations of internal wave dissipation, believed to be
32 the main source of mixing in the ocean interior, provide a similarly low value of $K_z=10^{-5}\text{m}^2\text{ s}^{-1}$ far from boundaries
33 [Polzin et al., 1995; Kunze et al., 2006]. It is therefore suspected that some regions might locally concentrate very
34 high vertical diapycnal diffusivity in order to close the balance. Viable candidates are topographically complex
35 boundary areas such as passages across mid ocean ridges and fracture zones [Polzin et al., 1997; Saint Laurent et
36 al., 2002; Thurnherr, 2006; MacKinnon et al., 2008], continental shelves [Rudnick et al., 2003], plateaus away
37 from boundaries such as the Kerguelen plateau [Meyer et al., 2015] or the Yermak plateau [Fer et al., 2015], and
38 semi-enclosed seas [Heywood et al., 2002] where internal tides and/or strong overflows can induce diapycnal
39 diffusivity several orders of magnitude higher than in the open ocean (K_z can reach 10^{-4} to $10^{-2}\text{m}^2\text{ s}^{-1}$). Recent
40 compilations of sparse direct microstructures measurements or fine scale parameterization suggest that the global
41 average mixing rate reaches the required value of $K_z\sim 10^{-4}\text{m}^2\text{ s}^{-1}$ [Waterhouse et al., 2014], however this estimate
42 is based on an upper bound for the mixing efficiency Γ of 0.2 [Osborn, 1980] and may overestimate the mixing
43 [Lavergne et al., 2015].

44 Large scale models used in climatic simulations usually do not include explicit internal tides, or their resolution is
45 only sufficient to resolve the first few lower modes [Simmons et al., 2004; Muller et al., 2012]. Parameterizations
46 of internal tidal mixing commonly used in large scale models are only based on a few in situ observations [e.g. St.
47 Laurent et al., 2002], and important questions regarding the vertical distribution of this mixing in the water column
48 still remain [Melet et al., 2013].

49 In the Indonesian seas region, large vertical mixing might be expected because of abrupt topography that
50 separates several deep semi-enclosed basins where all the baroclinic tidal energy may remain trapped locally. In
51 numerical simulations large diapycnal diffusivities were mainly attributed to internal tides breaking and the
52 addition of an internal tidal mixing parameterization [Koch-Larrouy et al., 2007] greatly improved the water
53 mass transformation in the region. These hypotheses were confirmed by recent high resolution simulations
54 including explicit tidal forcing [Kartadikaria et al., 2011; Nagai and Hibiya, 2015]. Microstructure measurements
55 performed in the interior of the large Banda Sea show diapycnal diffusion similar to open ocean values [Alford
56 et al., 1999], although these measurements were performed at a location far from topographic influence and
57 during the low-wind intermonsoon period [Field and Robertson, 2008]. Indeed recent microstructure
58 measurements performed during the INDOMIX campaign in the complex topographic region of the Halmahera
59 Sea and Ombai Strait show much larger vertical turbulent diffusion reaching $10^{-2}\text{m}^2\text{ s}^{-1}$ [Bouruet Aubertot et al.,
60 2012; Koch Larrouy et al., 2015].

61 We focus here on the impact of vertical diffusivity in driving a deep circulation and upwelling in the Timor Basin
62 (Figure 1). Because of its relatively simple geometry and boundary conditions, the Timor Basin appears to be an
63 ideal natural laboratory to study these processes that are expected to play a crucial role at the global scale. Moreover
64 important mixing is expected in the area since it is subjected to relatively large barotropic tidal currents (20 cm s^{-1})
65 at the western Timor sill where the Timor Basin opens to the Indian Ocean, and also over the Australian
66 continental shelf [Fieux et al., 1994; Molcard et al., 1996; Sprintall et al., 2004, 2009]. These tidal currents generate
67 strong internal tides with vertically integrated energy fluxes reaching several kW m^{-1} [Katsumata et al., 2010]. Our
68 objectives are to:

- 69 1) Assess the impact of turbulent mixing on deep circulation in the Timor Basin;
- 70 2) Estimate vertical turbulent diffusivity from a heat budget at the Timor Basin scale ($\sim 1000 \text{ km}$);
- 71 3) Test large scale model internal tide mixing parameterization in the simple configuration of the Timor Basin.

72 We use both observations and numerical simulations to address these questions. Section 2 describes the moored
73 velocity and temperature time series at Timor Sill used to estimate the transport, and the hydrographic stations
74 used to estimate the mean heat transport and its uncertainty. The numerical simulations, their tidal dissipation
75 parameterization and underlying assumptions are presented and discussed in Section 3. Section 4 details the
76 circulation in the deep Timor Basin from the observations and the numerical simulations. The importance of
77 rotational effects and whether the time mean overflow at the sill is hydraulically controlled is addressed in Section
78 5. An estimate of the mixing from the simulations and observed heat transport budget and Thorpe scales are given
79 in Section 6 with the accompanying error assessment. We conclude with a Discussion in Section 7 and comment
80 on the broader significance of our main findings to the importance of mixing in basins of $O(1000\text{km})$.

82 **2. The measurements**

83
84 The Timor Basin is a 1000 km long and narrow northeast oriented trench bounded by the $\sim 1000 \text{ m}$ isobaths of the
85 Timor and Leti islands southern shelves and the North Australian continental shelf, and the 1890 m sill at the
86 western end and the 1250 m sill at the eastern end (Figure 1). The basin is closed under 1890 m into the Indian
87 Ocean by the western sill, and closed on the eastern sill at 1250 m between Leti Islands and East Timor (Figure
88 1a).

89 Four moorings (Figure 1a) equipped with current meters and temperature sensors were anchored across the western
90 sill in Timor Passage between [122.68°E , 10.82°S] and [123.06°E , $12^\circ 06' \text{S}$] in 2004–2006 as part of the program

91 INSTANT [Sprintall et al., 2004]. Mooring velocity instrumentation configuration was fairly similar on all
92 moorings (see Figure 3 of Sprintall et al., 2009), with an upward-looking ADCP deployed so as to resolve the
93 surface to thermocline flow, and single-point current meters positioned at depth to resolve the sub-thermocline to
94 intermediate depth flow. Temperature and current velocity analyzed in the present study were recorded from single
95 point Aanderaa current meters/temperature sensors RCM07 or RCM08 with manufacturer given accuracy of 0.01
96 m/s, and 0.1°C, as well as additional temperature sensors Seabird SBE-37 and SBE-39 (manufacturer given initial
97 accuracy of 0.002°C) (Fig. 1b). Complete details of mooring instrumentation, deployment depths, data coverage
98 and the quality control for treatment of compass errors, time drifts, and fouling can be found in the online data
99 report [Cowley et al., 2008]. Average full-depth transport at the Timor strait was estimated from these moorings
100 as ~ 7.5 Sv [Sprintall et al., 2009]. Current and temperature profiles above the sill are considered here during the
101 second phase of INSTANT (from June 2005 to December 2006), when complete temperature profiles were
102 available. We have interpolated velocity and temperature data across the passage to estimate the transports. Several
103 interpolation schemes were used in the calculation of transport, and the range of transports obtained is considered
104 to represent the uncertainty error [e.g. see Sprintall et al., 2009 for further details]. In the following we use the
105 Along Strait Velocity (ASV) which is the Cartesian velocity rotated in the orientation of the along strait direction
106 at the western sill location (246°T). We focus on the deep circulation (below 1250 m) and the related vertical
107 mixing. For this depth range, the Timor Basin is semi-enclosed, and flow exchanges are limited to the Indian
108 Ocean through the deep western Timor sill [Gordon et al., 2003]. In addition to these mooring data we use
109 temperature and salinity profiles available from the World Ocean Database 2013 [hereafter WOD13, Boyer et al.,
110 2013] to characterize water masses within the basin. This dataset includes 131 high vertical resolution CTD and
111 38 recalibrated low resolution Ocean Station Data [see Boyer et al., 2013 for additional information on data]
112 conducted in the Timor Basin and reaching depths greater than 1250 m. In addition, we also use four CTD stations
113 from the collaborative Indonesian, Timorese and Australian "Arafura Timor Sea (ATSEA)" cruise conducted in
114 May 2010 onboard RV Baruna Jaya 8 (<http://atsea-program.org>) and kindly provided by A. Atmadipoera. These
115 data were collected from SBE-911 CTD casts with 24 Hz sampling rate. Data processing and analysis of CTD data
116 followed SBE Seasoft data processing (<http://www.seabird.com/software/sbe-data-processing>) with bin average
117 over one meter. The locations of all these stations are shown in Figure 1a. These data are used to estimate mean
118 temperature over time and iso-depths surfaces.

120 3. Numerical simulations

121 Numerical simulations have been made following the framework of Koch-Larrouy et al. [2007; 2008a; 2008b].
122 The model configuration is a sub-domain of the global model ORCA025 described by Barnier et al. [2006] using
123 the primitive equation ocean general circulation model NEMO [Madec, 2008]. This model configuration uses the
124 ORCA global tri-polar grid [Madec and Imbard, 1996] at $1\frac{1}{4}^\circ$ resolution and extends from 26.75°E to 142.25°E in
125 longitude and from 33.2°S to 30.3°N in latitude. The vertical grid has 46 levels, with a resolution ranging from 5
126 m at the surface to 250 m at the bottom. It uses a partial step representation of the bottom topography and a
127 momentum advection scheme which both yield significant improvements [Penduff et al., 2007; Le Sommer et al.,
128 2009]. Bathymetry is a smooth combination of ETOPO2 and GEBCO over shelves, with manual corrections within
129 the Lombok, Ombai and Timor straits to better represent the main pathways of the Indonesian Throughflow. The
130 numerical western Timor sill is set to 1880 m whereas the numerical eastern Timor sill reaches 1350 m. The model
131 is forced by daily climatological atmospheric forcing computed for the period 1990-2001 from the interannual
132 hybrid DRAKKAR Forcing Set 4 (DFS4) extensively described in Brodeau et al. [2010]. The western boundary
133 is closed by the African continent. The eastern, northern and southern boundaries are radiative open boundaries
134 [Treguier et al., 2001] constrained with a 150 day time-scale relaxation to a combination of (i) 5-day-average
135 climatological velocities computed over 1990-2001 from the interannual global $1/4^\circ$ simulation DRAKKAR
136 ORCA025-B83 [Dussin et al., 2009] and (ii) monthly climatological temperature and salinity from Levitus et al.
137 [1998] as biases were observed in the global simulation temperature and salinity properties. In doing so, we ensure
138 that the water masses entering the model have realistic properties [Koch-Larrouy et al., 2008a]. Vertical mixing is
139 modeled with a prognostic turbulent kinetic energy scheme, with background vertical diffusion and viscosity of
140 $10^{-5} \text{ m}^2 \text{ s}^{-1}$ and $10^{-4} \text{ m}^2 \text{ s}^{-1}$, respectively [Blanke and Delecluse, 1993; Madec, 2008]. In case of static instability,
141 vertical viscosity and diffusivity are raised to $10 \text{ m}^2 \text{ s}^{-1}$. The bottom friction is parameterized using a quadratic
142 stress law with a non dimensional drag coefficient of 10^{-3} .

143 Internal tides mixing parameterization

144 Two parameterizations of the enhanced vertical mixing due to internal tidal dissipation are tested: the St. Laurent
145 et al. [2002] parameterization (SL02 hereafter) and the Koch-Larrouy et al. [2007] parameterization (KL07
146 hereafter) which have the same general form but strongly differs in the prescribed vertical structure of the
147 dissipation rate. Both parameterization are designed for large scale models which do not resolve explicitly internal
148 tides and have been widely used in global or regional models, see for instance [Simmons et al 2004 (b), Jayne
149 2009] for the SL02 parameterization and [Koch Larrouy et al 2010, Melet et al 2011] for the KL07

150 parameterization. In both parameterization the horizontal advection and radiation of internal tide energy is
 151 neglected leading to a vertical turbulent kinetic energy balance. Consequently, the horizontal (x,y) dependence of
 152 the mixing in the parameterizations results only from the variation of the local power input into internal tides per
 153 bottom floor unit surface $E(x,y)$ [W m^{-2}]. The total internal tide energy dissipation through viscous friction and
 154 buoyancy flux B (expressed here in W m^{-3}) can therefore be written in the form of a separable function of horizontal
 155 and vertical coordinates of the form:

$$156 \quad + B = QE(x, y)F(z) \quad (1)$$

157 Where $F(z)$ is the vertical structure of the total dissipation rate normalized to ensure energy conservation
 158 $\int F(z)dz = 1$. When the total dissipation is integrated over the water column we have:

$$159 \quad \int (+ B)dz = QE(x, y) \quad (2)$$

160 Which shows that Q ($0 \leq Q \leq 1$) represents the fraction of the power input to internal tides dissipated locally. The
 161 fraction $(1-Q)$ of energy not dissipated locally is simply not taken into account in the parameterization and is
 162 supposed to contribute to the mean turbulent dissipation background of the model.

163 The buoyancy flux can be expressed as a diffusive turbulent flux $B = \rho_0 K_z N^2$, while the ratio between the dissipation
 164 rate by viscous friction and the buoyancy flux ($B/$) is known as the mixing efficiency and is noted Γ . According
 165 to Osborn [1980], Γ has an upper bound of 0.2, a value used in both KL07 and SL02 parameterization. Substituting
 166 by $\rho_0 K_z N^2 / \Gamma$ in Eq. 1 we can derive the following expression for the parameterization of K_z

$$167 \quad K_z = R_{if} Q \frac{E(x, y)F(z)}{\rho_0 N^2} \left[\text{m}^2 \text{s}^{-1} \right] \quad (3)$$

168 Where $R_{if} = \Gamma / (\Gamma + 1) = B / (+B)$ is known as the flux Richardson number. The parameterization can be rewritten in the
 169 equivalent form

$$170 \quad K_z = \Gamma q \frac{E(x, y)F(z)}{\rho_0 N^2} \left[\text{m}^2 \text{s}^{-1} \right] \quad (4)$$

171 It is this last form (Eq. 4) that has been used in previous studies [Saint Laurent et al., 2002; Koch Larrouy et al.,
172 2007; Melet et al., 2011; Melet et al., 2014]. However it is worth noting that in this expression $q=Q/(\Gamma+1)$
173 represents the fraction of internal tide dissipated by viscous friction only and has an upper bound of 0.83 (for
174 $\Gamma=0.2$) reached when all the tidal energy is dissipated locally through friction and buoyancy flux ($Q=1$). It seems
175 that this point was unnoticed in several recent publications which considered $q=1$ and $\Gamma=0.2$ for a fully local
176 dissipation of internal tide [Koch-Larrouy et al., 2007; Melet et al., 2011] or Lee waves energy [Melet et al., 2014].
177 A consequence of that is that the effective mixing efficiency used in these simulations was 17% larger than the
178 Osborn [1981] upper bound of 0.2. Because the odd upper bound of q has previously lead to some confusion we
179 prefer to use here the equivalent form Eq. 3 where we make use of the R_f instead of Γ and Q instead of q for its
180 simpler and more appealing upper bound of 1.

181 For both SL02 and KL07 parameterizations, $E(x,y)$ is inferred in this study from a finite element barotropic ocean
182 tidal model [Le Provost et al., 1994; Carrère and Lyard, 2003].

183 The two parameterizations SL02 and KL07 we employ were designed for different purposes. The main goal of the
184 SL02 parameterization was to reproduce the increase of the dissipation rate and turbulent mixing near the bottom
185 and driving the upwelling of abyssal waters. This deep mixing is believed to be mainly driven by the breaking of
186 very high vertical modes internal tides near the bottom [St. Laurent and Garrett, 2002]. By contrast, the main goal
187 of the KL07 parameterization was mainly to reproduce the correct water masses associated with the Indonesian
188 throughflow, which mainly requires to get the correct level of mixing within the pycnocline, while the bottom
189 mixing has a much weaker influence. This leads the KL07 parameterization to have a drastically different vertical
190 distribution of $F(z)$ and Q values compared to SL02.

191 The SL02 parameterization is based on experimental observations of Polzin et al. [1997] and prescribes an
192 exponential decay of the dissipation rate above the bottom with a typical 500 m length scale. However the rate of
193 energy dissipated by high vertical modes into heat by viscous friction near the bottom was estimated to represent
194 only 30% of the power converted from barotropic tides to internal tides [Polzin et al., 1997]. SL02 also considers
195 a mixing efficiency $\Gamma=0.2$ so that the total local dissipation fraction reaches $Q=(1+\Gamma)\times 0.3=0.36$ in that case. Most
196 of the internal tide energy escapes from the bottom layer as low modes [Polzin et al., 1997; St. Laurent and Garrett,
197 2002], and so the dissipation of energy associated with these lower modes is simply not taken into account in the
198 SL02 parameterization.

199 In the Indonesian seas these lower modes are trapped within the complex array of semi-enclosed basins and will
200 be dissipated locally. For this reason, we set the factor Q (and not q) to 1 in the KL07 parameterization. The fact
201 that all the energy dissipated locally was recently questioned by Nagai and Hibiya [2015] who used a high
202 resolution simulation with explicit tides to resolve internal tides and showed that Q could vary between 0.5 and 1
203 in various Indonesian straits, although they do not provide a value for the Timor Basin. The distribution $F(z)$ of
204 the dissipation rate in the vertical used in KL07 is also completely different from SL02. In KL07 It is based on a
205 simple scaling of the dissipation rate below the surface layer as $\propto N^2$ which better reflects the trapping and
206 dissipation of internal tides in the pycnocline. [The resulting strong dissipation rate in the pycnocline was essential
207 to reproduce the required water mass transformation in the KL07 simulations. The drawback is that this
208 parameterization may not reproduce the higher vertical modes energy dissipation near the bottom. Therefore the
209 KL07 parameterization likely underestimates the near bottom dissipation by higher modes as reproduced by the
210 SL02 parameterization. Indeed the dissipation rate of internal tide energy is very low above the bottom in KL07.
211 Figure 2 shows the ratio $R_D(z)$ between the integrated total dissipation rate below depth z , and the power input to
212 internal tides $E(x,y)$ computed as $R_D(z) = \int_H^z \rho(\varepsilon + B)dz / E(x,y) = Q \int_H^z F(z)dz$ for the two parameterizations
213 SL02 and KL07. This ratio was computed for an average stratification profile over the Timor Basin for KL07. The
214 fraction $R_D(z=1250 \text{ m})$ of $E(x,y)$ dissipated below the eastern sill depth (1250 m) is much higher in the SL02
215 parameterization where it varies between 0.26 and 0.32 (depending on the bottom depth) than in the KL02
216 parameterization where $R_D(z=1250 \text{ m})$ is nearly one order of magnitude smaller reaching 0.031. Therefore much
217 lower mixing is expected below 1250 m using the KL07 parameterization. As detailed in the next section the large
218 difference between the two parameterizations may be reduced by using a lower mixing efficiency above the bottom
219 in the SL02 parameterization.

220 Variable mixing efficiency

221 The common practice in large scale numerical simulations parameterizing internal tidal mixing, is to fix the value
222 of Γ to 0.2 (or equivalently R_f to 0.16) [Osborn, 1980]. However according to Osborn [1980], 0.2 represents a
223 maximum value. Arneborg [2002] for instance recommended a weaker value of 0.12 following observations of St.
224 Laurent and Schmitt [1999]. Direct numerical simulations [Shih et al., 2005, SH05 hereafter] and laboratory
225 experiments [Barry et al., 2001, B01 hereafter] have shown a dramatic decrease of Γ with increasing turbulent
226 intensity $I = \varepsilon / (N^2)$, where I represents the ratio of the destabilizing effect of turbulence versus the stabilizing effect

227 of viscosity and stratification. Note that $I=R^{4/3}$, where $R=L_0/\eta$ is the ratio of the Ozmidov scale $L_0= \omega^{-1/2}N^{-3/2}$ which
228 represents the spatial scale of the larger overturns of the turbulent cascade, to the Kolmogorov dissipative scale
229 $\eta=(\nu^3/\epsilon)^{1/4}$.

230 Therefore R represents the extension of the turbulent energy cascade, the so-called inertial sub-range. Based on
231 their simulations SH05 show that the mixing efficiency scales as $\Gamma=0.2$ for $R<30$ and decreases as $\Gamma=2R^{-2/3}$ for
232 $R>30$. B01 also obtain a decrease of Γ with R in their laboratory experiment with a different scaling law as $\Gamma=9$
233 $R^{-8/9}$ for $R>150$. The decrease of Γ with R which was obtained over a limited range of R in B01 and SH05 is still
234 debated [Gregg et al., 2012; Kunze et al., 2012]. However several recent in situ observations have confirmed this
235 behavior [Davis and Monismith, 2011; Bluteau et al., 2013; Lozovatsky and Fernando, 2013]. Bluteau et al. [2013]
236 notably suggest that the best scaling is obtained comparing their observations with B01, while Davis and
237 Monismith [2011, DM11 hereafter] find a scaling law as $\Gamma=4.5R^{-2/3}$ consistent with SH05 although with a different
238 coefficient.

239 In our simulations the factor R depends on the choice of the internal tidal mixing parameterization. For KL07, the
240 prescribed dependence $\propto N^2$ implies that I and R are almost constant within the water column with a value $R\sim 150$.
241 Applying the SH05 scaling, we find that mixing efficiency is close to 0.07, while the B01 and DM11 scaling
242 suggest a value of 0.1 and 0.16 respectively. The exponential decrease of Γ in the SL02 parameterization results in
243 a strong increase of R above the bottom. Typically we find that R increases from $R\sim 300$ at 1200 m depth to
244 $R\sim 10000$ near the western sill depth (1890 m). Following the three previous scalings, we find that Γ (resp. R_f)
245 varies in the range $0.04\delta 0.1$ (resp. 0.039 to 0.09) at 1250 m to $0.02\delta 0.04$ (resp. 0.0196 to 0.039) at the sill depth.
246 Given the strong decrease of R suggested for the SL02 parameterization and the relatively large uncertainty in this
247 decrease, we have considered two simple additional sensitivity tests for the simulations using the SL02
248 parameterization where the mixing efficiency Γ (resp. R_f) was lowered to 0.1 (resp. 0.09) and 0.05 (resp. 0.047)
249 compared to the usual 0.2 value of Osborn [1980]. Note that to our knowledge the only internal tidal mixing
250 parameterizations considering a decrease of the mixing efficiency are Melet et al. [2013] and Lavergne et al.
251 [2015]. Melet et al. [2013] have suggested a simple empirical decrease of the mixing efficiency with N as $\Gamma=0.2$
252 $N^2/(\omega^2+N^2)$, with ω the angular velocity of the Earth. An advantage of this formulation is that it prevents the
253 diffusion from going to infinity when the stratification goes to 0. However the Melet et al. [2013] formulation
254 affects only very low abyssal stratification. In the Timor Basin, the deep stratification N^2 is more than two orders
255 of magnitude larger than ω^2 and this formulation has a negligible influence on the mixing efficiency. Lavergne et

256 al. [2015] have implemented the Shih et al. [2005] scheme for a global model and have shown a reduction by a
257 factor of two of the overturning circulation.

258 Simulations runs

259 Five 40-year simulations were performed to assess the role of the tidal mixing parameterization on the deep ocean
260 circulation in the Timor Basin (Table 1). The simulations are almost identical, they differ only regarding the tidal
261 parameterization applied (or not) in the Timor Basin: NOTIDE has no tidal mixing parameterization, TIDE-KL
262 uses the KL07 parameterization, in TIDE-SL02 the SL02 parameterization is applied over the Timor Basin using
263 the Osborn [1980] value $R_f = 0.16$ ($\Gamma = 0.2$), TIDE-SL01 and TIDE-SL005 are the same as TIDE-SL02 with a
264 reduced mixing R_f of 0.09 ($\Gamma = 0.1$), and 0.047 ($\Gamma = 0.05$), respectively. In all the simulations, the KL07
265 parameterization is applied everywhere else in the Indonesian Seas. This choice was made in order to modify only
266 the deep circulation in the Timor Basin without affecting the overall equilibrium reached in the KL07 simulations.
267 In the following δ TIDE simulations δ is used to refer to all the simulations that include an internal tide
268 parameterization (TIDE-KL, TIDE-SL02, TIDE-SL01, TIDE-SL005).

269 **4. Circulation in deep Timor Basin**

270 In the following we focus on the transport below 1250 m as for this depth range the circulation is constrained by
271 a closed boundary at the eastern sill (1250 m depth) and exchanges are mainly restricted to the Indian Ocean
272 through the deep western Timor sill (1890 m). Figure 3a shows the INSTANT transport measurement at the
273 western Timor sill integrated over three depth intervals [1250 δ 1600 m], [1600 δ 1890 m (sill depth)] and [1250 δ
274 1890 m].

275 For the three depth intervals, the instantaneous volume transport $IT(t)$ shows large intra-seasonal variations of
276 several Sv (standard deviation of 1.02 Sv for [1250 δ 1890 m]) with alternating inflow and outflow in the Timor
277 Basin. The inflows and outflows are in phase over the three depth intervals indicating that the instantaneous ASV
278 direction is most of the time consistent over the whole water column below 1250 m depth. Some of these large
279 intra-seasonal reversals are likely associated with strong Kelvin waves remotely forced in the equatorial Indian
280 Ocean and propagating eastward along the south boundary of the Indonesian seas while also radiating energy in
281 the Indonesian seas through the Lombok, Makassar, Ombai and Timor Passage [Sprintall et al., 2000; Wijffels and
282 Meyers, 2004; Sprintall et al., 2009; Drushka et al., 2010].

283 The cumulative average transport: $CT(t) = \frac{1}{t} \int_0^t IT(\tau) d\tau$ (where $t=0$ is June 1st 2005) is represented in Figure 3b for
284 the three depth intervals. Note that the initial value $CT(0)=IT(0)$ whereas the final value $CT(T)$ represents the
285 transport averaged over the full record length T . The cumulative average $CT(t)$ is used here to assess the time
286 interval averaging needed for a stable mean flow to emerge. The signature of the strong intra-seasonal variations
287 is smoothed when the averaging period t reaches 5-6 months, after which a residual average inflow below 1600 m
288 and an average outflow above 1600 m are revealed. The inflow reaches the value of 0.163 Sv when averaged over
289 the full observational period (June 2005 to December 2006) with an uncertainty range [0.11, 0.25] Sv. This inflow
290 is almost exactly compensated by an outflow of -0.161 Sv with uncertainty range [-0.14, -0.21] Sv between 1250
291 and 1600 m so that the time-averaged transport is almost zero (0.002 Sv) between 1250 m and the bottom. These
292 observational estimates of the transport show that a mean deep circulation is present in the Timor Basin.

293 The instantaneous transport between 1250 m and the western sill indicates a strong intra-seasonal variability for
294 the five simulations (not shown) with similar standard deviations of 1.14, 1.06, 1.08, 1.11 and 1.08 Sv for the
295 NOTIDE, TIDE-KL, TIDE-SL01, TIDE-SL01 and TIDE-SL005 runs respectively, in good agreement with the
296 1.02 Sv variability of the observations.

297 To compare both the vertical structure and the convergence of the deep circulation toward its mean equilibrium
298 state in the data and in the simulations we have computed the cumulative average stream function as a function of
299 the depth z : $CT(t,z) = \frac{1}{t} \int_0^t IT(\tau,z) d\tau$, where $IT(\tau,z)$ is the instantaneous transport between the sill depth and the
300 depth z (Figure 4). The 0 iso-value of $CT(t,z)$ marks the limit of the deep circulation cell, while the depth of its
301 maximum marks the transition between inflow and outflow. In the NOTIDE simulation there is no deep inflow
302 and the mean circulation is almost zero below 1500 m depth. In contrast, all simulations that include a tidal
303 parameterization show that a deep circulation emerges after a time averaging of a few months. Although the
304 vertical distribution of the mixing is very different in the TIDE-KL and TIDE-SL02, -SL01, -SL005 simulations,
305 the structure of the deep circulation cell is relatively robust. The mean deep cell extends between the western sill
306 depth (1880 m in the simulations compared to 1890 m in the observations) and up to 1500 m depth in the TIDE-
307 KL simulation and up to 1350 m in the TIDE-SL02 simulation. The cell extension is reduced to 1400 m and 1500
308 m depth when the mixing efficiency is decreased to 0.1 and 0.05 in the TIDE-SL01 and TIDE-SL005 simulations,
309 respectively. Note that the smaller vertical extent in the simulations compared to the observations results partially
310 from a deeper eastern sill (1350 m) than in the observations (1250 m). The intensity of the transport shows a wide
311 range of variability; the maximum inflow transport reaches 0.07 Sv for the TIDE-KL run which is about 50 %

312 smaller than the observed inflow (0.16 Sv), and 0.28 Sv in the TIDE-SL02 run which is about 75 % larger than
 313 the observed inflow. The inflow transport decreases to 0.15 Sv in the TIDE-SL01 run which is very close to the
 314 observations, while the TIDE-SL005 run has an inflow transport of 0.09 Sv. Interestingly, both the depth extent
 315 and transport are very close in the TIDE-SL005 and TIDE-KL runs despite the very different distribution of the
 316 dissipation rate and vertical turbulent diffusion K_z .

317 Further insight into the structure of the deep circulation is given by the time averaged zonal stream function along

318 the strait for the five simulations (Figure 5), defined as: $\Psi(x,z) = \int_{\text{bottom}}^z \int_{y_s}^{y_N} u(x,y,z) dy dz$

319 where u is the eastward velocity and y_s, y_N the south and north boundaries of the basin at the depth z . The
 320 magnitude of $\Psi(x,z)$ represents the transport between the bottom and the depth z for each longitude x , while the
 321 intensity of the zonally averaged vertical and horizontal velocities are given (within a constant factor) by the
 322 negative of the horizontal and the vertical derivative of $\Psi(x,z)$ respectively. In all the simulations with tidal
 323 parameterizations (Figure 5 b-e), the stream lines show an overturning circulation below the eastern model sill
 324 depth of 1350 m (red dashed line), with an inflow of decreasing intensity along the strait. As previously noted, the
 325 smallest deep transport is observed for the TIDE-KL and TIDE-SL005 for which the smallest mixing intensity is
 326 expected. This reduced transport occurs with a slightly reduced vertical extension of the circulation cell below
 327 1500 m. An upwelling with maximum intensity around $z \sim 1600$ m is observed between 123°E and 128°E in the
 328 TIDE-SL02 and TIDE-SL01 simulations, while this upwelling has a deeper maximum (around $z \sim 1700$ m) in the
 329 TIDE-KL and TIDE-SL005 simulations. In the NOTIDE simulation the circulation is nearly zero below 1500 m
 330 (Figure 5a).

331 A more detailed view of the vertical structure of the section and time averaged ASV at the western Timor sill is
 332 shown in Figure 6c for INSTANT observations and for the five simulations (averaged over the last ten years). For
 333 INSTANT observations, the error estimate on the transport is computed following Sprintall et al. [2009]. To allow
 334 direct comparison, the modeled ASV is multiplied by a correcting factor equal to the ratio of the modeled sill width
 335 to the real sill width for each depth z . Below the eastern sill depth (i.e. for $z > 1250$ m) the observed mean velocity
 336 field shows an inflow below 1600 m with a maximum of 8 cm s^{-1} at 1800 m and an outflow above 1600 m. A
 337 similar structure of the mean ASV is found in all TIDE simulations, albeit with some important variations in the
 338 inflow velocity maximum ranging from 5.5 cm s^{-1} (in the TIDE-KL simulation) to 19 cm s^{-1} (in the TIDE-SL02
 339 simulation). In contrast, the inflow velocity in the NOTIDE simulation is almost zero.

340 The observational data and numerical simulations also suggest that the inflow is associated with a density decrease
341 across the sill. Figure 6b shows the potential density profiles at two positions corresponding to the model grid
342 points, just west of the western sill [122.75°E, 11.42°S] and within the Timor Basin [125°E, 10.44°S]. An estimate
343 of the observed density profiles at these positions was computed using objective interpolation with a 32 km radius
344 from neighboring CTD measurements. Moving from west to east across the sill, a density drop of $\sim 0.1 \text{ kg m}^{-3}$ is
345 observed below 1500 m depth in the interpolated CTD data. The amplitude of this density drop is comparable in
346 the TIDE-SL02 run (averaged over the last ten years of the run) where it reaches 0.075 kg m^{-3} and slightly decreases
347 for the runs with reduced mixing efficiencies TIDE-SL01 and TIDE-SL005 with 0.07 kg m^{-3} and 0.06 kg m^{-3}
348 respectively. The TIDE-KL and NOTIDE simulations show smaller density drops of 0.04 kg m^{-3} . This density
349 drop generates a pressure drop that drives the observed inflow in the Timor Basin. We first try to determine whether
350 this density driven overflow is critical and/or influenced by the rotation before performing further analysis on the
351 heat budget and mixing estimates.

352

353 **5. Overflow criticality and influence of rotation**

354

355 It is important to address whether the time mean overflow at the sill is in a state of hydraulic control. Hydraulic
356 control at a sill is associated with strong mixing processes [Tessler et al., 2010; Alford et al., 2013] that would not
357 be reproduced with the coarse model resolution we are using. The Froude number describes the criticality of the
358 flow and is defined as the ratio of the overflow advection speed to the phase speed of (reduced) gravity waves.
359 $Fr=U/(g\phi h)^{1/2}$, where U is the along strait velocity, $g\phi=\Delta\rho/\rho_0$ is the reduced gravity and h the overflow active layer
360 height. Hydraulic control takes place for $Fr=1$ which marks the transition from subcritical to supercritical flow.
361 For a continuous stratification the definition of the reduced gravity is ambiguous, so we follow Whitehead [1989]
362 who defined $\Delta\rho$ as the density drop upstream and downstream of the strait. Using Figure 6b we set $\Delta\rho$ in a large
363 range [0.05, 0.15] kg.m^{-3} considering the uncertainty associated with the space and time range of the
364 measurements. Following Whitehead [1989], the height h can be defined as the region of divergence between
365 upstream and downstream potential density profiles which gives $h=490 \text{ m}$ (see Figure 6b). We also take an
366 alternative definition of h as the height of the zero crossing of the ASV around 1550 m which gives $h=340 \text{ m}$ (see
367 Figure 6c). With both these definitions, h is in the range [340 490 m]. Finally we consider the depth average over
368 h of the ASV in the range [0.02, 0.03] m.s^{-1} for the time-mean flow with maximum instantaneous inflow reaching
369 0.12 m.s^{-1} according to the reported uncertainty [Sprintall et al., 2009]. With these numbers we estimate that Fr is

370 in the range [0.02, 0.07] for the mean flow with instantaneous value peaking at 0.27, and conclude that the flow is
371 always subcritical. Therefore there is no hydraulic control at the Timor sill over the INSTANT time period. This
372 further suggests that the strong vertical mixing in the basin is rather a cause than a consequence of the overflow,
373 although the moderate shear associated with the overflow could still slightly reinforce the vertical mixing.

374
375 Rotation effects can induce cross tilting of the isopycnals and cross section shear. Our experimental sampling and
376 the model resolution below 1250 m do not allow us to accurately characterize rotation effects, an issue which is
377 beyond the scope of this paper. Therefore in the following we only assess their significance in a general sense. The
378 basic effect of rotation can be estimated assuming a model of semi-geostrophic flow where the pressure gradient
379 balances the Coriolis force in the across strait direction [Pratt and Whitehead, 2007]. The simplest assumption is
380 to consider a constant velocity of the overflow, then the across basin tilting of the overflow interface is
381 $\Delta z = fLU/(g\phi)$, where L is the basin width at $z=h$. Given the uncertainties of h and $g\phi$ a moderate tilting across the
382 basin is found in the range $\Delta z = [13, 96]$ m. Considering the upper bound of Δz , a sensitivity test on the impact of
383 rotation was made by applying a simple linear tilting of the isopycnals/isotherms and the currents structure (not
384 shown), it was found that the inflow transport in the lower cell was changed by less than 1%, while the
385 compensating outflow varied from less than 20%. The resulting impact of rotation on heat budget and computation
386 of K_z developed in the next section did represent a minor source of uncertainty and was therefore not further taken
387 into account in the remainder of the analysis..

388

389 **6. Mixing and heat budget in the deep Timor Basin**

390 **6.1 Vertical turbulent diffusivity in the model runs**

391 Since the Timor is semi-enclosed, there are mixing processes which sustain the density drop across the sill. The
392 across sill density decrease at depth is coincident with an increase of the tidal currents. Figure 6a shows the M2
393 tidal ellipses interpolated from the nearby current meters and obtained from a harmonic analysis of the mooring
394 velocity data at the western sill over the [June 2005 to December 2006] period. The current ellipses show a strong
395 baroclinic structure with stronger velocities ($>10 \text{ cm s}^{-1}$) below 1600 m depth suggesting that baroclinic tides could
396 generate the strong mixing at depth. We have computed the vertical turbulent diffusivity K_z averaged over the
397 Timor Basin in all the simulations (Figure 7b). These mean K_z are defined as $\langle K_z \rangle = \langle K_z N^2 \rangle / \langle N^2 \rangle$ where $\langle \cdot \rangle$ is a
398 time and basin mean for each depth and N the stratification. The advantage of this definition of the mean is that it
399 is based on an average of turbulent buoyancy fluxes which is physically meaningful, whereas a direct average of

400 K_z has poor physical significance. In practice, a direct average of K_z may put too much weight on the strong K_z
401 present in very weakly stratified regions that have little impact on the turbulent fluxes. In the TIDE-SL02
402 simulation $\langle K_z \rangle$ is strong ($[2-8] \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$) and increases toward the bottom. The TIDE-SL01 and TIDE-SL005
403 simulations show a similar trend with a two-fold and four-fold decrease of $\langle K_z \rangle$ which is consistent with the
404 respective decrease of the mixing efficiency Γ (resp. R_f) to 0.1 (resp. 0.09) and 0.05 (resp. 0.047). The vertical
405 turbulent mixing is smaller in the TIDE-KL simulation ($[0.4-1] \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$) with a more even distribution
406 throughout the water column as expected for the KL07 parameterization. In contrast, $\langle K_z \rangle$ for the NOTIDE
407 simulation remains very weak throughout the water column with values one to two orders of magnitude smaller
408 than in all the TIDE simulations.

409 These results suggest that in the model a large vertical turbulent diffusivity induced by internal tides at the Timor
410 sill and in the Timor Basin generates a mixing of deep dense (and cold) water ($z > 1600 \text{ m}$) with less dense (and
411 warmer) overriding waters ($z < 1600 \text{ m}$). Therefore a density gradient and a pressure drop is generated across the
412 sill that drives the deep inflow between the sill and 1600 m depth. Because the basin is closed to the east below
413 1250 m depth, a return flow is generated above 1600 m. In this process an average upward velocity is also generated
414 over the Timor Basin (Figure 7a) which contributes to the upwelling of cold and dense water. At equilibrium, a
415 balance is achieved at the basin scale between the lateral inflow and upward transport of this dense and cold water
416 and the downward turbulent diffusive transport of less dense water and heat. In order to estimate the turbulent
417 diffusivity from the observations data set we apply a heat budget at the basin scale.

418

419 **6.2 Vertical turbulent diffusivity derived from the heat budget in the observations**

420 Below 1250 m the Timor Basin is a semi-enclosed sea as it is closed by the eastern sill, and hence we can apply a
421 heat transport balance at the basin scale to infer the strength of mixing in the observations. This method was
422 originally proposed and discussed by Hogg et al. [1982] and Whitehead and Worthington [1982]. These authors
423 use a heat budget below an isopycnal surface. Here we use a z -coordinate budget which requires further hypotheses
424 concerning the spatio-temporal variability of the heat fluxes, as discussed below. This choice is made because the
425 deeper isopycnal surface intersects the bottom which makes their areal computation difficult. An additional
426 advantage is that the z -coordinate budget can be easily tested and validated in the numerical simulations. For each
427 depth z , we compute the heat budget over the volume $V(z)$ enclosed between the bottom and the depth z with open
428 boundaries defined by the western sill vertical section of area $S_1(z)$ and the horizontal surface of area $S_2(z)$ as
429 illustrated in Figure 8. We decompose the vertical velocity W and the potential temperature :

$$430 \quad = \langle \theta \rangle_{S_2(z)} + \theta' \\ 431 \quad W = \langle W \rangle_{S_2(z)} + W'$$

431 where $\langle \theta \rangle$ denotes the average over surface $S_2(z)$ and long time τ equal to the 1.5 year full record length, and the
 432 primes represent the fluctuations around the average. Then for each depth z , the heat balance over the volume
 433 $V(z)$ of the basin and averaged over a long time interval τ gives:

$$434 \quad \iiint_V C_p \left\langle \rho \frac{\partial \theta}{\partial t} \right\rangle dV = S_1 C_p \langle \rho U \theta \rangle_{S_1, \tau} - S_2 C_p \langle \rho \rangle_{S_2, \tau} \langle W \rangle_{S_2, \tau} \langle \theta \rangle_{S_2, \tau} + S_2 C_p \langle \rho \rangle_{S_2, \tau} \langle W' \theta' \rangle_{S_2, \tau} + G \quad (5)$$

435 where U is the along-strait velocity, C_p is the specific heat capacity per unit mass of sea water and the z dependence
 436 was dropped to simplify the notation. The term on the left hand side gives the rate of change of heat content in
 437 the volume $V(z)$. The first term on the right hand side is the horizontal heat transport across $S_1(z)$, the second and
 438 third term represent the vertical heat transport across $S_2(z)$ decomposed in vertical advection and turbulent
 439 fluctuations, and the last term G is the geothermal heat flux integrated over the bottom surface. Assuming that the
 440 long term average of the heat rate of change vanishes, Eq. 5 reduces to:

$$441 \quad S_2 \langle \rho \rangle_{S_2, \tau} \langle W \rangle_{S_2, \tau} \langle \theta \rangle_{S_2, \tau} - S_1 \langle \rho U \theta \rangle_{S_1, \tau} = S_2 \langle \rho \rangle_{S_2, \tau} \langle W' \theta' \rangle_{S_2, \tau} + G/C_p \quad (6)$$

442 Since the strait is closed at its eastern end, we can estimate the vertical transport across S_2 from the continuity
 443 equation as:

$$444 \quad S_2 \langle W \rangle_{S_2, \tau} = S_1 \langle U \rangle_{S_1, \tau} \quad (7)$$

445 We then make the assumption that $\langle W' \theta' \rangle_{S_2, \tau}$ can be considered as a vertical turbulent temperature flux, and
 446 modeled using the turbulent vertical diffusion coefficient as:

$$447 \quad \langle W' \theta' \rangle_{S_2, \tau} = \langle K_z \rangle_{S_2, \tau} \frac{\partial \langle \theta \rangle_{S_2, \tau}}{\partial z} \quad (8)$$

448 With this definition, $\langle K_z \rangle_{S_2, \tau}$ is expressed as the mean turbulent flux divided by the mean potential temperature
 449 gradient. This definition is also consistent with the definition of the mean K_z in the simulations given in the
 450 previous section.

451 Finally using Eq. 6, 7 and 8 we can get an estimate of K_z through an advection-vertical diffusion balance as:

$$452 \quad \langle K_z \rangle_{S_2, \tau} = \frac{S_1 \langle \rho \rangle_{S_2, \tau} \langle U \rangle_{S_1, \tau} \langle \theta \rangle_{S_2, \tau} - S_1 \langle \rho U \theta \rangle_{S_1, \tau} - G/C_p}{S_2 \langle \rho \rangle_{S_2, \tau} \frac{\partial \langle \theta \rangle_{S_2, \tau}}{\partial z}} \quad (9)$$

453 The bottom surface integrated geothermal flux was computed using a constant geothermal heat flux of 60 mW m^{-2} .
 454 Although this value is an upper bound for the region [Emile-Geay and Madec, 2009] it has only a marginal

455 influence on $\langle K_z \rangle_{S_2}$. The computation of the density and potential temperature requires salinity measurements
 456 which were not available from the Timor mooring, we therefore infer salinity from the T-S relationship observed
 457 below 1250 m from the 131 deep CTD stations available in the Timor Basin. We find that the T-S relationship is
 458 almost linear with a correlation coefficient $r^2=0.95$ (not shown). Using this linear relationship to infer salinity, the
 459 resulting error in the in-situ density and potential temperature computation is then negligible.

460 From the mooring data we estimate all the terms in Eq. 9 except $\langle \theta \rangle_{S_2}$, $\langle \rho \rangle_{S_2}$, and $\frac{\partial \langle \theta \rangle_{S_2}}{\partial z}$. Hydrological
 461 measurements made over the Timor Basin during the mooring deployment period (from June 2005 to December
 462 2006) are too scarce in time and space to allow an estimate of these terms. Therefore we used all the available
 463 hydrological data in the region (Figure 1a) as well as the climatological values to estimate these mean terms as
 464 follows:

- 465 1) Direct ensemble average from CTD data for each depth
- 466 2) Objective interpolation of CTD data over the surface $S_2(z)$ for each depth and surface average
- 467 3) Direct ensemble average of climatological values of the density and the temperature for each depth as
 468 obtained from the World Ocean Atlas 2009 [Locarnini et al., 2010].

469 *Error estimates on the heat budget in the observations*

470 Since the three mean estimates are derived from the same set of CTD/bottle data the variation between the three
 471 averaging methods may not completely represent the standard error of the mean temperature/density. However, an
 472 estimation of the standard error on the mean climatological temperature is available in the World Ocean Atlas
 473 2009 data set. This uncertainty considers sparsity in time and space, as well as the variable quality of the data set
 474 [Locarnini et al., 2010].

475 We therefore calculate a maximum global uncertainty that is the sum of this climatological standard error and
 476 the variation obtained from the three estimates of the mean temperature.

477 This uncertainty is small for $\langle \theta \rangle_{S_2}$ and $\langle \rho \rangle_{S_2}$, but represents an important source of error for the mean temperature
 478 gradient near the bottom $\frac{\partial \langle \theta \rangle_{S_2}}{\partial z}$. A 95% confidence interval on the $\langle K_z \rangle_{S_2}$ estimates was computed from both

479 the uncertainty in these mean terms and the uncertainty of the transport estimates following Sprintall et al. [2009].

480 To compute the confidence interval, the uncertainty was propagated in Eq. 9 using a Monte Carlo method, i.e. by
 481 randomly selecting 1000 values of temperature and transport and computing the resulting variation in $\langle K_z \rangle_{S_2}$.

482 The random values of temperature and transport were selected from a Gaussian distribution truncated to $[-\sigma, +\sigma]$,

483 where the standard deviation of the distribution equals the half error range on transport or temperature/density.
484 In addition, we remind the reader that, as discussed in the previous section, we did not take into account the impact
485 of rotation in this computation.

486 Observed vertical turbulent diffusivity and validation of the method on the model runs

487 The mean vertical turbulent diffusivity estimated using the observations in Eq. 9 is given in Figure 7b. It is almost
488 constant with a value close to $1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$ down to 1800 m and an uncertainty range of $[0.6, 2.5] \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$. At
489 1850 m the mean K_z increases up to $3 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$ but this increase is not significant in regard to the very large
490 uncertainty near the bottom. Mean K_z estimated from the observations is up to two orders of magnitude larger than
491 the mean K_z in the NOTIDE simulation (Figure 7b) which further suggests the need to include the effect of the
492 internal tide in numerical models to reproduce the correct level of mixing. The mean K_z in the TIDE-SL02
493 simulation is larger by a factor of 2 to 4 between 1500 m and 1800 m depth compared to the upper bound estimate
494 of the observational K_z . This discrepancy is reduced to a factor of 1 to 2 for the TIDE-SL01 run that employs a
495 more plausible $\Gamma=0.1$ mixing efficiency and falls within the observational bounds for $\Gamma=0.05$ in the TIDE-SL005
496 run. The TIDE-KL run shows K_z values which are smaller than the observations by a factor 2 below 1500 m depth.
497 The most important assumption made in Eq. 9 to estimate the mean K_z from the observations, is that that the term
498 $\langle W' \epsilon' \rangle_{s_2}$, can be completely modeled as a turbulent flux. This assumption is not straightforward since fluctuations
499 are defined both around a time and surface average. In order to test this assumption we have computed a second
500 estimate of the mean K_z for all the TIDE simulations using the same heat budget method that was applied to
501 observations (i.e. Eq. 9 without the geothermal flux). The heat budget estimate of K_z obtained in this way is very
502 close to the prescribed K_z for all the TIDE runs except for the TIDE-KL run where there is a discrepancy of up to
503 2 (Figure 7b). The reason for the larger discrepancy in this run is unknown. For the most part these estimates
504 support the validity of Eq. 9.

505 **6.3 Comparison of the vertical upwelling velocity and buoyancy flux in the simulations and observations**

506 Turbulent buoyancy flux

507 The estimation of K_z using Eq. 3 in the model is strongly dependent on the value of the stratification N^2 and
508 therefore on vertical gradients of temperature and salinity that are poorly resolved in the simulations. Moreover,
509 the large K_z found below 1750 m depth in the TIDE-SL runs are associated with low values of N^2 , therefore, strong
510 K_z at depth will have a moderate impact on water mass transformation. The turbulent buoyancy flux $K_z N^2$ [W kg^{-1}]
511 (expressed here per unit mass) is less dependent on the vertical resolution and will actually drive the water mass
512 transformation. We have estimated the mean turbulent buoyancy flux $b = \langle K_z N^2 \rangle_{s_2}$, over the Timor Basin for each

513 depth in all simulations. $\langle K_z N^2 \rangle_{s_2}$ was also estimated from the observations data set as
 514 $\langle K_z N^2 \rangle_{s_2, \tau} \cong \langle K_z \rangle_{s_2, \tau} \langle N^2 \rangle_{s_2, \tau}$. Figure 7c shows that the buoyancy flux from the observations falls between the
 515 estimates from the TIDE-SL02 and TIDE-SL01 simulations. The discrepancy between observations and the TIDE-
 516 SL simulations observed for K_z below 1750 m depth disappears for the turbulent buoyancy flux. Note that the
 517 decrease of the buoyancy flux below 1750 m results from the null flux condition at the bottom.

518 Vertical upwelling velocity

519 It is also interesting to compare the averaged vertical velocity $\langle W \rangle_{s_2}$ in the model and its estimation from the
 520 observations data set using Eq. 7. In the observations, the vertical velocity is always positive: it increases from
 521 nearly zero at 1250 m depth to a maximum value of $4 \times 10^{-6} \text{ m s}^{-1}$ around 1600 m (Figure 7a) and decreases toward
 522 zero near the bottom. A similar behavior is obtained for the mean upwelling velocity in the TIDE simulations
 523 although the maximum is reached deeper at 1750 m. The largest upwelling velocity is found in the TIDE-SL02
 524 simulation reaching $6 \times 10^{-6} \text{ m s}^{-1}$, which is close to the estimate in the TIDE-SL01 simulation but lower than in the
 525 TIDE-SL005 run ($2.5 \times 10^{-6} \text{ m s}^{-1}$). The lowest vertical velocity is obtained in the TIDE-KL run ($2 \times 10^{-6} \text{ m s}^{-1}$) but
 526 this run also shows a smoother decrease above 1750 m. In the NOTIDE simulation, the vertical velocity is negative
 527 above 1700 m and close to zero below, which further implies that the average upwelling resulting from the inflow
 528 in the Timor Basin can only be reproduced when a parameterization of internal tidal mixing is included.

530 **6.3 Observed vertical turbulent diffusivity from Thorpe scales computation**

531 We give an independent estimate of the turbulent mixing using the statistical length scale of overturns known as
 532 the Thorpe scale [Thorpe, 1977].

533 In stratified turbulence, the scale of the largest eddies (i.e. the Ozmidov scale Lo) is limited by the work required
 534 to counter the buoyancy forces. At the Ozmidov scale, the balance between the buoyancy force resulting from the
 535 stratification and the inertial force associated with the turbulent motion results in the scaling [Gargett, 1988]:

$$536 \quad Lo \sim \left(\frac{\epsilon}{N^3} \right)^{1/2} \quad (10)$$

537 Thorpe [1977] and Dillon [1982] have shown that the Ozmidov scale is statistically closely related to the RMS
 538 length of potential density overturns detected within a turbulent patch (the Thorpe scale L_{Th}). Dillon finds in
 539 average that $Lo \sim 0.8(+/-0.4) L_{Th}$ which allows using Eq. 10 to determine ϵ .

540 The determination of the Thorpe scales requires a careful handling of the noise in the measurements which
 541 generates spurious density inversions in the weakly stratified parts of the water column. For this reason we limit

542 our analysis to the 45 high vertical resolution CTD profiles (1m) of the WOD13 with good quality flag (code=0).
 543 To get rid of spurious inversions, we apply the method of Ferron et al. [1998] slightly modified by Gargett and
 544 Garner [2008]. The basic idea is to build a potential density profile that tracks only significant density variations
 545 in the original profile. We first determine the noise level σ_N of the measurements from the standard deviation of
 546 the detrended potential density over 15 m length sections taken in well mixed region of the profiles. We find that
 547 in average $\sigma_N=1.8\times 10^{-4}$ kg m⁻³ and define a threshold for significant variation as four times this noise level
 548 $\sigma_{th}\sim 7.5\times 10^{-4}$ kg m⁻³. In the part of the profile where no overturns are detected, we assume a low dissipation rate
 549 $\varepsilon=1\times 10^{-10}$ W kg⁻¹ typical of the background dissipation rate in the ocean far from boundaries [Waterhouse et al.,
 550 2014]. Since there is a considerable scattering and variation in the Thorpe estimates we have first averaged the
 551 values obtained from the Thorpe scales in 200 m bins over each profile and then build a mean dissipation rate $\langle \varepsilon \rangle$
 552 from an ensemble average over the 45 profiles. We determine a 95% confidence estimates on this ensemble
 553 average from a bootstrap method. The value of the buoyancy flux per unit mass b (in W kg⁻¹) and the K_z are then
 554 estimated as $b = \Gamma \langle \varepsilon \rangle$ and $\langle K_z \rangle = \Gamma \langle \varepsilon \rangle \langle N \rangle^{-2}$ choosing the canonical value of the mixing efficiency $\Gamma=0.2$ ($R_f=0.16$)
 555 to allow comparisons with tide_KL07 and tide_SL02 runs. Although there is a very large uncertainty in the Thorpe
 556 estimates of the buoyancy flux and K_z , the mean values obtained are in good agreement with the observational
 557 estimate obtained from the heat budget method (Figure 7 b and c) and therefore strengthens the previous
 558 conclusions.

559

560 **7. Conclusions and discussion**

561 Most of the deep sill passages in the Indonesian region are found along the ITF inflow route via the Sangihe Ridge,
 562 Halmahera Sea sill and Lifamatola Passage [Gordon et al., 2003]. For these sills the circulation is characterized by
 563 a strong overflow driven by a large scale density drop across the sill and often hydraulically controlled [see Van
 564 Aken et al., 2009]. In this situation strong mixing is generally observed as a result of the hydraulic control [Tessler
 565 et al., 2010; Alford et al., 2013]. In contrast for the deep overflow at the Timor sill, described here below 1250 m
 566 depth, there is no circulation associated with the deep part of the ITF [Gordon et al., 2003]. The deep overflow
 567 revealed in the INSTANT mooring data and TIDES simulations is driven by strong mixing in the Timor Basin that
 568 sustains a density drop across the sill. Simultaneously an upwelling of dense water is generated at the basin scale.
 569 In the absence of tidal mixing parameterization there is no mean circulation generated. Therefore the overflow in
 570 Timor is a result of strong mixing that is driven by an external process rather than a cause of the deep ocean mixing.

571 We provided estimates from observations data set of the average mixing rate K_z in the Timor Basin based on a
572 heat budget and a Thorpe scale analyses. The mixing rate is high $\sim 1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$, which is almost two orders of
573 magnitude larger than deep ocean estimates, but comparable to other estimations of K_z in regions of rugged
574 topography where elevated mixing is expected [see for instance Waterhouse et al., 2014 or Wunsch and Ferrari,
575 2004 for a review].

576 We have compared these estimates with results from numerical simulations. We first tested the case of a numerical
577 parameterization where there was no internal tide mixing parameterization, the K_z was one to two orders of
578 magnitude smaller than the observational estimate and the deep circulation was almost zero. We then considered
579 simulations with two internal tidal mixing parameterizations with drastically different vertical distributions of the
580 dissipation rate. The first was the KL07 parameterization, designed to reproduce the strong dissipation of energy
581 expected in the pycnocline from the trapping of internal tides energy in the Indonesian seas. The second was the
582 SL02 parameterization, designed to reproduce the strong dissipation above the bottom from the higher vertical
583 modes, but does not reproduce the strong fraction (64%) of energy carried away by low modes. We find that the
584 modelled K_z is about two times smaller than the observational estimate below 1500 m depth with the KL07
585 parameterization leading to a deep inflow transport 50 % smaller than in the observations. In contrast the
586 simulation with the SL02 parameterization overestimates the K_z by up to a factor 5 which results in a deep inflow
587 70 % larger than in the observations. According to several experimental [Barry et al., 2001], numerical [Shih et
588 al., 2005] and observational studies [Bluteau et al., 2013; Lozovatsky and Fernando, 2013; Davis and Monismith,
589 2011], the strong increase of the turbulence intensity above the bottom resulting from the SL02 parameterization
590 will lead to a strong decrease of the mixing efficiency Γ (resp. R_f) below the widely used value of 0.2 (resp. 0.16)
591 [Osborn, 1980]. Based on an estimate of this decrease we have considered two sensitivity tests simulations using
592 the SL02 parameterization with reduced Γ (resp. R_f) of 0.1 (resp. 0.09, TIDE-SL01) and 0.05 (resp. 0.047, TIDE-
593 SL005). The deep inflow in the TIDE-SL01 simulation is very close to the observations while it is 40 % smaller
594 in the TIDE-SL005 simulation.. These results are consistent with the turbulent buoyancy flux $K_z N^2$ estimates
595 which show that the TIDE-SL01 estimates fall within the observational bounds down to the sill depth suggesting
596 that this estimate with a reduced mixing efficiency of 0.1 is the closer to the observations. Note however that a
597 reduction of the Q factor and a reduction of R_f would have the same effect in the SL parameterization, therefore it
598 is not possible to firmly determine whether R_f is reduced or whether the Q factor proposed in the SL
599 parameterization is overestimated for the Timor Basin.

600 The comparison of the simulations with and without the internal tidal mixing parameterizations strongly suggests
601 that internal tides are the main source of mixing and drive the deep circulation in the Timor Basin. However, since
602 the model is of low resolution, mixing processes generated by mean flow boundary mixing are under-resolved,
603 notably near the sill, while barotropic tidal boundary mixing is not represented at all. This raises the question of a
604 possible compensation for such processes by the internal tide mixing parameterization in the model. Although the
605 exact estimation of these processes would require a higher resolution model and an explicit representation of tides,
606 we can argue that they should not represent the main mixing source. For instance in the TIDE-SL01 simulation for
607 which the deep transport in Timor Basin is very close to the observations, the maximum velocity at the sill (not
608 corrected by the ratio of the modeled sill width to the real sill width) is underestimated by a factor of 2. Assuming
609 a quadratic bottom drag in the form $C_d|U|U$ we can conclude that the turbulent dissipation rate resulting from mean
610 flow friction would be underestimated by a factor 8 in a thin layer near the bottom. Clearly this would not explain
611 the nearly two orders of magnitude difference in K_z observed up to 400 m above the sill between the NOTIDE
612 simulations and the observations. The barotropic tide within the deep region of the Timor Basin has moderate
613 amplitude [$< 20 \text{ cm s}^{-1}$, as seen in Ding et al., 2012]. This is probably sufficient to significantly increase the friction
614 and induce mixing but only in a thin shear layer above the bottom. Moreover, global estimates show that internal
615 tide generation in the deep ocean is about 30 times larger than direct barotropic tide dissipation through bottom
616 friction [Egbert and Ray, 2000]. Internal tides trapped in the Timor Basin will break locally and generate
617 significant mixing in the whole water column, a process intended to be captured by the KL07 with a focus on the
618 pycnocline and SL02 parameterization with a focus on the bottom dissipation. Therefore, although the model likely
619 under-estimates some mixing mechanisms, the breaking internal tide still appears as the best candidate to explain
620 the strong mixing process in the Timor Basin.

621 Our results also suggest that although the KL07 parameterization reproduces the needed strong dissipation and
622 mixing in the pycnocline, it likely underestimates the near bottom mixing and the resulting deep circulation. Recent
623 microstructure measurements performed in the energetic Halmahera Sea (north east of the Indonesian archipelago)
624 during the INDOMIX 2010 campaign [P. Bouruet-Aubertot et al., 2012; Koch Larrouy et al., 2015] show that
625 simple scaling of energy dissipation with internal tide energy as done in KL07 parameterization indeed
626 underestimates the dissipation rate below the pycnocline. The recent high resolution simulations in the Indonesian
627 seas including explicit internal tides by Nagai and Hibiya [2015] also show that the depth integrated K_z is larger
628 than what is predicted by the KL07 simulation in deep areas like the Timor Basin or the Banda sea, suggesting that
629 KL07 underestimates the strong increase of K_z in abyssal waters. This underestimation of the deep mixing will

630 have an impact on the deep circulation in the Indonesian seas as illustrated here in the case of the Timor Basin. A
631 basic correction to the KL07 parameterization would imply a redistribution of energy dissipation by higher modes
632 near the bottom and consequently a reduced dissipation in the pycnocline. This reduction would however be
633 modest if the fraction dissipated in the bottom layer by high modes is kept below 36 % as in the SL02
634 parameterization. Testing of such a corrected parameterization against higher resolution numerical runs and/or
635 direct microstructure measurements should be considered in the future. The exponential form and decay scale of
636 the SL02 parameterization also needs further testing with high resolution data and high resolution simulations,
637 since the resolution of the present measurements and simulations are not sufficient for this purpose. Note that
638 recent studies of internal tidal mixing parameterizations have suggested modifications to the SL02
639 parameterization [Declodt and Luther, 2012; Melet et al., 2013] but these studies focused on bottom mixing in
640 the open ocean and still neglected the large fraction of energy that radiates away as low modes.

641 The idea that the deep vertical circulation could be driven by abyssal vertical mixing was first discussed by Munk
642 [1966]. This question was also emphasized more recently by Huang [1999] who pointed out that deep circulation
643 could be a result rather than a cause of deep ocean mixing. This mechanism was verified in some recent numerical
644 simulations [Jayne, 2009; Saenko and Merryfield, 2011] and laboratory experiments [Whitehead and Wang, 2008].
645 However other studies point out that a large part of the upwelling of abyssal waters may result from wind driven
646 upwelling in the Southern Ocean, reducing the need for strong abyssal mixing [Toggweiler and Samuels 1995;
647 Hughes and Griffiths, 2006; Webb and Sugimoto, 2001]. A recent study using a Lagrangian ocean model in an
648 idealized basin by Haertel and Fedorov [2012] even suggests that the meridional overturning circulation could be
649 driven at leading order by adiabatic processes while diffusion represents only a first order perturbation increasing
650 the heat transport from 10 % to 20 %. The generation of a deep circulation by turbulent mixing as evidenced here
651 in the Timor Basin represents a good illustration of this process at a geophysical scale in a relatively well controlled
652 environment where we can minimize the temporal trend term in the heat budget equation by averaging over a
653 longer period.

654 In summary our observations based on both basin scale heat budget and a fine scale analyses of CTD profiles show
655 that the deep inflow in the Timor Basin is subcritical and driven by a strong mixing within the basin. Comparison
656 of observations with numerical models where internal tide mixing is parameterized (or not) strongly suggest that
657 internal tides breaking is the main driver of this mixing. The site appears as an ideal laboratory to study the mixing
658 parameterization and the representation of relatively fine scale circulation in coarse resolution models. Two
659 internal tide mixing parameterization were tested, one focusing the mixing in the pycnocline [KL07], the other one

660 focusing the mixing near the bottom [SL02]. The analyses reveal the need to take into account the mixing generated
661 near the bottom by higher mode dissipation (as modelled in the SL02 parameterization) to generate the deep inflow,
662 a process underestimated by the KL07 parameterization, although KL07 represents the mixing in the core of the
663 ITF [Koch Larrouy et al, 2007]. The simulations also show that the deep circulation is strongly sensitive to the
664 choice of the mixing efficiency. These results appeal for the design of an internal tide mixing parameterization
665 that consistently takes into account both the near bottom dissipation of higher modes and the dissipation of the
666 lower modes in the full water column

667 **References**

- 668 Alford, M., Gregg, M. and M. Ilyas (1999). Diapycnal mixing in the Banda Sea: results of the first
669 microstructure measurements in the Indonesian throughflow. *Geophysical Research Letters*, 26, 27416
670 2744.
- 671 Alford, M., H., Girton, J., B., Voet, G., Carter, G., S., Mickett, J., B., and J. Klymack (2013). Turbulent mixing
672 and hydraulic control of abyssal water in the Samoan Passage. *Geophysical Research Letters*, 40(17),
673 4666-4674, doi: 10.1002/grl.50684.
- 674 Arneborg, L. (2002). Mixing efficiencies in patchy turbulence. *Journal of Physical Oceanography*, 32(5), 1496-
675 1506.
- 676 Barnier, B., Madec, G., Penduff, T., Molines, J.M., Tréguier, A.-M., Le Sommer, J., Beckmann, A., Biastoch,
677 A., Boening, C., Dengg, J., Derval, C., Durand, E., Gulev, S., Remy, E., Talandier, C., Theetten, S.,
678 Maltrud, M., McClean, J. and B. DeCuevas (2006). Impact of partial steps and momentum advection
679 schemes in a the global ocean circulation model at eddy permitting resolution. *Ocean Dynamics*, 56(5-6),
680 543-567.
- 681 Barry, M. E., Ivey, G. N., Winters, K. B And J. Imberger (2001). Measurements of diapycnal diffusivities in
682 stratified fluids. *Journal of Fluid Mechanics*, 442(1), 267-291.
- 683 Blanke B. and P. Delecluse (1993). Variability of the tropical Atlantic ocean simulated by a general circulation
684 model with two different mixed layer physics. *Journal of Physical Oceanography*, 23:136361388.
- 685 Bluteau, C. E., Jones, N. L. and G. N. Ivey (2013). Turbulent mixing efficiency at an energetic ocean site.
686 *Journal of Geophysical Research: Oceans*, 118(9), 4662-4672.
- 687 Bouruet-Aubertot, P., Cuypers, Y., Ferron, B., Dausse, D., Menage, O., Atmadipoera, A., & Jaya, I. (2012,
688 December). Finescale parameterization of turbulent mixing and internal tides in the Indonesian
689 Throughflow from INDOMIX experiment. In *AGU Fall Meeting Abstracts (Vol. 1, p. 1872)*.

690 Boyer, T.P., J. I. Antonov, O. K. Baranova, C. Coleman, H. E. Garcia, A. Grodsky, D. R. Johnson, R. A.
691 Locarnini, A. V. Mishonov, T.D. O'Brien, C.R. Paver, J.R. Reagan, D. Seidov, I. V. Smolyar, and M. M.
692 Zweng (2013). World Ocean Database 2013, NOAA Atlas NESDIS 72, S. Levitus, Ed., A. Mishonov,
693 Technical Ed.; Silver Spring, MD, 209 pp., <http://doi.org/10.7289/V5NZ85MT>

694 Brodeau, L., B. Barnier, T. Penduff, A.-M. Tréguier and S. Gulev (2010). An ERA40 based atmospheric forcing
695 for global ocean circulation models, *Ocean Modelling*, 31, 88-104, doi:10.1016/j.ocemod.2009.10.005.

696 Carrère, L., and F. Lyard (2003), Modeling the barotropic response of the global ocean to atmospheric wind and
697 pressure forcing: Comparisons with observations, *Geophys. Res. Lett.*, 30(6), 1275, doi:10.1029/
698 2002GL016473. Cowley, R., B. Heaney, S. Wijffels, L. Pender, J. Sprintall, S. Kawamoto, and R. Molcard
699 (2008), INSTANT Sunda Data Report Description and Quality Control, available at
700 http://www.marine.csiro.au/~cow074/INSTANTdataQC_v4.pdf

701 Davis, K. A. and S. G. Monismith, S (2011). The modification of bottom boundary layer turbulence and mixing
702 by internal waves shoaling on a barrier reef. *Journal of Physical Oceanography*, 41(11), 2223-2241.

703 Decloedt, T. and D. S. Luther (2012). Spatially heterogeneous diapycnal mixing in the abyssal ocean: A
704 comparison of two parameterizations to observations. *Journal of Geophysical Research: Oceans*, 117(C11),
705 1978-2012.

706 Dillon, T. M. (1982), Vertical overturns: A comparison of Thorpe and Ozmidov length scales, *J. Geophys. Res.*,
707 87(C12), 9601-9613, doi:10.1029/JC087iC12p09601.

708 Ding Y., Bao X., Yu H. and L Kuang (2012). A numerical study of the barotropic tides and tidal energy distribution
709 in the Indonesian seas with the assimilated finite volume coastal ocean model. *Ocean Dynamics*, 62, 515-
710 532.

711 Drushka, K., Sprintall, J., Gille S. T. and I. Brodjonegoro (2010). Vertical Structure of Kelvin Waves in the
712 Indonesian Throughflow Exit Passages. *Journal of Physical Oceanography*, 40, 1965-1987.

713 Dussin R., A.-M. Treguier, J.M. Molines, B. Barnier, T. Penduff, L. Brodeau and G. Madec (2009). Definition of
714 the interannual experiment ORCA025-B83, 1958-2007. LPO Report 902.

715 Egbert, G.D. and R.D. Ray (2000). "Significant dissipation of tidal energy in the deep ocean inferred from satellite
716 altimeter data," *Nature*, 405, 775-778.

717 Emile-Geay, J. and G. Madec (2009). Geothermal heating, diapycnal mixing and the abyssal circulation. *Ocean*
718 *Science* 5, 203-217.

- 719 Fer, I., Müller, M., & Peterson, A. K. (2015). Tidal forcing, energetics, and mixing near the Yermak Plateau.
720 Ocean Science 11, 287-304.
- 721 Ferron, B., Mercier, H., Speer, K., Gargett, A., & Polzin, K. (1998). Mixing in the Romanche fracture zone. Journal
722 of Physical Oceanography, 28(10), 1929-1945.
- 723 Fieux, M., C. Andrieu, P. Delecluse, A. G. Ilahude, A. Kartavtseff, F. Mantsi, R. Molcard and J. C. Swallow
724 (1994). Measurements within the Pacific-Indian oceans throughflow region, Deep-Sea Research, 41, 1091-
725 1130.
- 726 Gargett, A. E. (1988). The scaling of turbulence in the presence of stable stratification. Journal of Geophysical
727 Research: Oceans (1978-2012), 93(C5), 5021-5036.
- 728 Gargett, A., & Garner, T. (2008). Determining Thorpe scales from ship-lowered CTD density profiles. Journal of
729 Atmospheric and Oceanic Technology, 25(9), 1657-1670.
- 730 Gordon, A. L., C. F. Giulivi and A. G. Ilahude (2003). Deep topographic barriers within the Indonesian Seas, Deep
731 Sea Research, Part II, 50, 2205-2228.
- 732 Gregg, M. C., Alford, M. H., Kontoyiannis, H., Zervakis, V. and D. Winkel (2012). Mixing over the steep side of
733 the Cycladic Plateau in the Aegean Sea. Journal of Marine Systems, 89(1), 30-47.
- 734 Haertel, P. and A. Fedorov (2012). The ventilated ocean. Journal of Physical Oceanography 42, 161-164.
- 735 Heywood, K. J. Naveira Garabato A. C. and D. P. Stevens (2002). High mixing rates in the abyssal Southern
736 Ocean. Nature, 415, 1011-1014.
- 737 Hogg, N. Biscaye, P. E., Gardner, W. and W. J. Schmitz Jr (1982). On the transport and modification of Antarctic
738 Bottom Water in the Vema Channel. Journal of Marine Research, 40, 231-283.
- 739 Huang, R. X. (1999). Mixing and energetics of the oceanic thermohaline circulation. Journal of Physical
740 Oceanography, 29, 727-746.
- 741 Hughes G. O and R. W. Griffiths (2006). A simple convective model of the global overturning circulation,
742 including effects of entrainment into sinking regions. Ocean Modelling, 12, 46-79.
- 743 Jackson, P. R. and C. R. Rehmann (2003). Laboratory measurements of differential diffusion in a diffusively
744 stable, turbulent flow. Journal of Physical Oceanography, 33(8), 1592-1603.
- 745 Jayne, S. R. (2009). The impact of abyssal mixing parameterizations in an ocean general circulation model. Journal
746 of Physical Oceanography, 39, 1756-1775.
- 747 Kartadikaria, A. R., Miyazawa, Y., Varlamov, S. M., and K. Nadaoka (2011). Ocean circulation for the Indonesian
748 seas driven by tides and atmospheric forcings: Comparison to observational data. Journal of Geophysical
749 Research, 116(C9), C09009.

750 Katsumata, K., Wijffels, S. E., Steinberg, C. R. and R. Brinkman (2010). Variability of the semidiurnal internal
751 tides observed on the Timor Shelf. *Journal of Geophysical Research*, 115(C10), C10008.

752 Koch-Larrouy, A., G. Madec, P. Bouruet-Aubertot, T. Gerkema, L. Bessieres and R. Molcard (2007). On the
753 transformation of Pacific Water into Indonesian Throughflow Water by internal tidal mixing, *Geophysical*
754 *Research Letters*, 34, L04604, doi:10.1029/2006GL028405.

755 Koch-Larrouy A., G. Madec, D. Iudicone, A. Atmadipoera and R. Molcard (2008a). Physical processes
756 contributing to the water mass transformation of the Indonesian Throughflow, *Ocean Dynamics*, 58, 275-
757 288, doi: 10.1007/s10236-008-0154-5.

758 Koch-Larrouy A., G. Madec, B. Blanke and R. Molcard (2008b). Water mass transformation along the Indonesian
759 throughflow in an OGCM, *Ocean Dynamics*, 58, 289-309, doi: 10.1007/s10236-008-0155-4.

760 Koch-Larrouy, A., Atmadipoera, A., van Beek, P., Madec, G., Aucan, J., Lyard, F., ... & Souhaut, M. (2015).
761 Estimates of tidal mixing in the Indonesian archipelago from multidisciplinary INDOMIX in-situ data.
762 *Deep Sea Research Part I: Oceanographic Research Papers*.

763 Koch-Larrouy, A., Lengaigne, M., Terray, P., Madec, G., & Masson, S. (2010). Tidal mixing in the Indonesian
764 Seas and its effect on the tropical climate system. *Climate Dynamics*, 34(6), 891-904.

765 Kunze, E., Firing, E., Hummon, J. M., Chereskin, T. K., and A. M. Thurnherr (2006). Global abyssal mixing inferred
766 from Lowered ADCP shear and CTD strain profiles, *Journal of Physical Oceanography*, 36, 1553-1576.

767 Kunze, E., MacKay, C., McPhee-Shaw, E. E., Morrice, K., Girton, J. B. and S. R. Terker (2012). Turbulent mixing
768 and exchange with interior waters on sloping boundaries. *Journal of Physical Oceanography*, 42(6), 910-
769 927.

770 de Lavergne, C., Madec, G., Le Sommer, J., Nurser, A. G., & Naveira Garabato, A. C. (2015). The impact of a
771 variable mixing efficiency on the abyssal overturning. *Journal of Physical Oceanography*, (2015).

772 Le Provost, C., M. L. Genco, F. Lyard, P. Vincent, and P. Canceil (1994), Spectroscopy of the world ocean tides
773 from a finite element hydrodynamic model, *J. Geophys. Res.*, 99, 24,777-24,798.

774 Le Sommer J., T. Penduff, S. Theetten, G. Madec and B. Barnier (2009). How momentum advection schemes
775 influence current-topography interactions at eddy-permitting resolution. *Ocean Modelling*, 29, 1-14.
776 doi:10.1016/j.ocemod.2008.11.007.

777 Ledwell J. R., Montgomery E. T., Polzin K. L., St. Laurent L. C., Schmitt R. W. and J. M. Toole (2000).
778 Evidence for enhanced mixing over rough topography in the abyssal ocean. *Nature*, 403, 79-182.

779 Levitus, S., T. Boyer, M. Conkright, T. O'Brien, J. Antonov, C. Stephens, L. Stathopoulos, D. Johnson and R. Gelfeld
780 (1998). World Ocean database 1998, technical Report NESDID18, NOAA Atlas.

781 Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, H. E. Garcia, O. K. Baranova, M. M. Zweng and D.
782 R. Johnson (2010). *World Ocean Atlas 2009, Volume 1: Temperature*. S. Levitus, Ed. NOAA Atlas
783 NESDIS 68, U.S. Government Printing Office, Washington, D.C., 184pp.

784 Lozovatsky, I. D. and H. J. S. Fernando (2013). Mixing efficiency in natural flows. *Philosophical Transactions of*
785 *the Royal Society A: Mathematical, Physical and Engineering Sciences*, 371(1982).

786 MacKinnon, J., Johnston T.M.S. and R. Pinkel (2008). Strong transport and mixing of deep water through the
787 Southwest Indian Ridge. *Nature Geoscience*, 1, 755-758.

788 Madec G. and M. Imbard (1996). A global ocean mesh to overcome the north pole singularity, *Climate Dynamic*
789 12, 381-388.

790 Madec G. (2008). NEMO reference manual, ocean dynamic component: NEMO-OPA. Preliminary version,
791 Tech. Rep. 27, Note du pôle de modélisation, Institut Pierre Simon Laplace (IPSL), France, ISSN No
792 1288-1619.

793 Melet, A., Verron, J., Gourdeau, L. and A. Koch-Larrouy (2011). Equatorward pathways of Solomon Sea water
794 masses and their modifications. *Journal of Physical Oceanography*, 41(4), 810-826.

795 Melet, A., Hallberg, R., Legg, S. and K. Polzin (2013). Sensitivity of the ocean state to the vertical distribution
796 of internal-tide-driven mixing. *Journal of Physical Oceanography*, 43(3), 602-615.

797 Melet, A., Hallberg, R., Legg, S. and M. Nikurashin (2014). Sensitivity of the Ocean State to Lee Wave Driven
798 Mixing. *Journal of Physical Oceanography*, 44, 900-921.

799 Meyer, A., Sloyan, B. M., Polzin, K. L., Phillips, H. E., & Bindoff, N. L. (2015). Mixing variability in the
800 Southern Ocean. *Journal of Physical Oceanography*, 45(4), 966-987.

801 Molcard, R., M. Fieux and A. G. Ilahude (1996). The Indo-Pacific throughflow in the Timor Passage. *Journal of*
802 *Geophysical Research*, 101(C5), 12411-12420.

803 Müller, M., J. Y. Cherniawsky, M. G. G. Foreman, and J.-S. von Storch (2012). Global M2 internal tide and its
804 seasonal variability from high resolution ocean circulation and tide modeling, *Geophysical Research*
805 *Letter*, 39, L19607, doi:10.1029/2012GL053320.

806 Munk W. (1966). Abyssal recipes. *Deep-Sea Res.*13, 707-730.

807 Munk W and Wunsch C. (1998). Abyssal recipes II: energetics of tidal and wind mixing. *Deep-Sea Research*,
808 45, 1976-2009.

809 Nagai, T. and T. Hibiya (2015). Internal tides and associated vertical mixing in the Indonesian Archipelago.
810 *Journal of Geophysical Research: Oceans*, 120, 3373-3390.

811 Osborn T. R. (1980). Estimates of the local rate of vertical diffusion from dissipation measurements. *Journal of*
812 *Physical Oceanography*, 10, 83-89.

813 Penduff T., J. Le Sommer, B. Barnier, A.-M. Treguier, J.M. Molines and G. Madec (2007). Influence of
814 numerical schemes on current-topography interactions in $\frac{1}{4}^\circ$ global ocean simulations. *Ocean Science*, 3,
815 509-524.

816 Polzin, K. L., Toole, J. M., and R. W. Schmitt (1995). Finescale parameterizations of turbulent dissipation, *J.*
817 *Phys. Oceanogr.*, 25, 306-328.

818 Polzin, K. L., Toole, J. M., Ledwell, J. R. and R. W. Schmitt (1997). Spatial Variability of Turbulent Mixing in
819 the Abyssal Ocean. *Science*, 276, 5309, 93-96.

820 Pratt, L. J. and J. A. Whitehead (2007). *Rotating Hydraulics - Nonlinear topographic effects in the ocean and*
821 *atmosphere*, Springer-Verlag, 608pp.

822 Rudnick, D. L., Boyd, T. J., Brinard R. E., Carter G. S. and G. D. Egbert (2003). From tides to mixing along the
823 Hawaiian Ridge. *Science*, 301(5631), 355-357.

824 Saenko O. A. and W. J. Merryfield (2011). On the effect of topographically enhanced mixing on the global
825 ocean circulation. *Journal of Physical Oceanography*, 35, 826-834.

826 St. Laurent, L. and R. W. Schmitt (1999). The contribution of salt fingers to vertical mixing in the North Atlantic
827 tracer release experiment. *Journal of Physical Oceanography*, 29(7), 1404-1424.

828 St.. Laurent, L. and Garrett C. (2002). The Role of Internal Tides in Mixing the Deep Ocean. *Journal of Physical*
829 *Oceanography*, 32, 2882-2899.

830 St. Laurent, L., Simmons, H. L. and S. R. Jayne (2002). Estimating tidally driven mixing in the deep ocean.
831 *Geophysical Research Letter*, 29(23). doi: 10.1029/2002GL015633.

832 Shih, L. H., Koseff, J. R., Ivey, G. N. and J. H. Ferziger (2005). Parameterization of turbulent fluxes and scales
833 using homogeneous sheared stably stratified turbulence simulations. *Journal of Fluid Mechanics*, 525,
834 193-214.

835 Simmons, H. L., R. H. Hallberg, and B. K. Arbic (2004). Internal wave generation in a global baroclinic tide
836 model, *Deep Sea Research, Part II*, 51, 3043-3068.

837 Simmons, H. L., Jayne, S. R., Laurent, L. C. S., & Weaver, A. J. (2004) (b). Tidally driven mixing in a
838 numerical model of the ocean general circulation. *Ocean Modelling*, 6(3), 245-263.

839 Sprintall, J., A. L. Gordon, R. Murtugudde and R. D. Susanto (2000). A semiannual Indian Ocean forced Kelvin
840 wave observed in the Indonesian seas in May 1997, *Journal of Geophysical Research*, 105, 17217-17230.

841 Sprintall, J., S. Wijffels, A. L. Gordon, A. Field, R. Molcard, R. DwiSusanto, I. Soesilo, J. Sopaheluwakan, Y.
842 Surachman and H. Van Aken (2004). INSTANT: A new international array to measure the Indonesian
843 Throughflow. *Eos Trans. AGU*, 85(39), 3696376, doi:10.1029/2004EO390002.

844 Sprintall, J., S., Wijffels, R. Molcard and I. Jaya (2009). Direct Estimates of the Indonesian Throughflow
845 Entering the Indian Ocean: 2004-2006. *Journal Of Geophysical Research*, 114, C07001,
846 doi:10.1029/2008JC005257.

847 Sundfjord, A., Fer, I., Kasajima, Y and H. Svendsen (2007). Observations of turbulent mixing and hydrography
848 in the marginal ice zone of the Barents Sea. *Journal of Geophysical Research: Oceans*, 112(C5),
849 doi:10.1029/2006JC003524.

850 Tessler, Z. D., A. L. Gordon, L. J. Pratt, and J. Sprintall (2010). Transport and dynamics of the Panay Sill
851 overflow in the Philippine Seas. *Journal of Physical Oceanography*, 40, 2679-2695.

852 Thorpe, S. A. (1977). Turbulence and mixing in a Scottish loch. *Philosophical Transactions of the Royal Society
853 of London A: Mathematical, Physical and Engineering Sciences*, 286(1334), 125-181.

854 Thurnherr A. M (2006). Diapycnal mixing associated with an overflow in a deep submarine canyon. *Deep-Sea
855 Research II*, 53, 194-206.

856 Thurnherr, A. M., St Laurent, L. C., Speer, K. G., Toole, J. M. and J. R. Ledwell (2005) Mixing associated with
857 sills in a canyon on the mid-ocean ridge flank. *Journal of Physical Oceanography*, 35, 1370-1381.

858 Treguier, A.-M., B. Barnier, AP. De Miranda, J.M. Molines, N. Grima, M. Imbard, G. Madec, C. Messenger, T.
859 Reynaud and S. Michel (2001). An eddy-permitting model of the Atlantic circulation: Evaluating open
860 boundary conditions, *Journal Of Geophysical Research*, 106, 22115-22129.

861 Toggweiler J. R and B. Samuels (1995). Effect of Drake Passage on the global thermohaline circulation. *Deep-
862 Sea Research*, 42, 477-500.

863 Van Aken, H. M., I. S. Brodjonegoro and I. Jaya (2009). The deepwater motion through the Lifamatola Passage
864 and its contribution to the Indonesian throughflow. *Deep-Sea Research*, 56, 1203-1216,

865 Waterhouse, A. F., MacKinnon, J. A., Nash, J. D., Alford, M. H., Kunze, E., Simmons, H. L., ... and C. M. Lee
866 (2014). Global patterns of diapycnal mixing from measurements of the turbulent dissipation rate. *Journal
867 of Physical Oceanography*, 44(7), 1854-1872.

868 Webb, D. J. and N. Suginohara (2001). Vertical mixing in the ocean. *Nature*, 409(37).

869 Wijffels, S. and G. Meyers (2004). An intersection of oceanic waveguides: Variability in the Indonesian
870 Throughflow region, *Journal of Physical Oceanography*, 34, 1232-1253.

871 Whitehead, J. A. and L. V. Worthington (1982). The flux and mixing rates of Antarctic Bottom Water within the
872 North Atlantic. *Journal of Geophysical Research*, 87, 7902-7924.

- 873 Whitehead, J. A. (1989). Internal hydraulic control in rotating fluids - applications to oceans. *Geophysical and*
874 *Astrophysical Fluid Dynamics*, 48(1), 169-192.
- 875 Whitehead, J. A. and W. Wang (2008). A Laboratory Model of Vertical Ocean Circulation Driven by Mixing .
876 *Journal of Physical Oceanography*, 38, 1091-1106.
- 877 Wunsch C. And R. Ferrari (2004). Vertical mixing, energy, and the general circulation of the oceans. *Annual*
878 *Review of Fluid Mechanics*, 36, 281-314.
- 879

880 **Figures and tables**

881 **Tables**

	NOTIDE	TIDE-KL	TIDE-SL02	TIDE-SL01	TIDE-SL005
Timor Basin	No tidal mixing parameterization	KL07 parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=1$	St. Laurent parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=0.36$	St. Laurent parameterization $R_f=0.09$ ($\Gamma=0.1$) $Q=0.36$	St. Laurent parameterization $R_f=0.047$ ($\Gamma=0.05$) $Q=0.36$
Indonesian seas (except Timor Basin)	KL07 parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=1$	KL07 parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=1$	KL07 parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=1$	KL07 parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=1$	KL07 parameterization $R_f=0.16$ ($\Gamma=0.2$) $Q=1$

882

883 **Table 1** Mixing parameterization (KL07 or SL02) and mixing efficiency (R_f/Γ) used in and outside the Timor
884 Basin for the different runs.

885

886

887

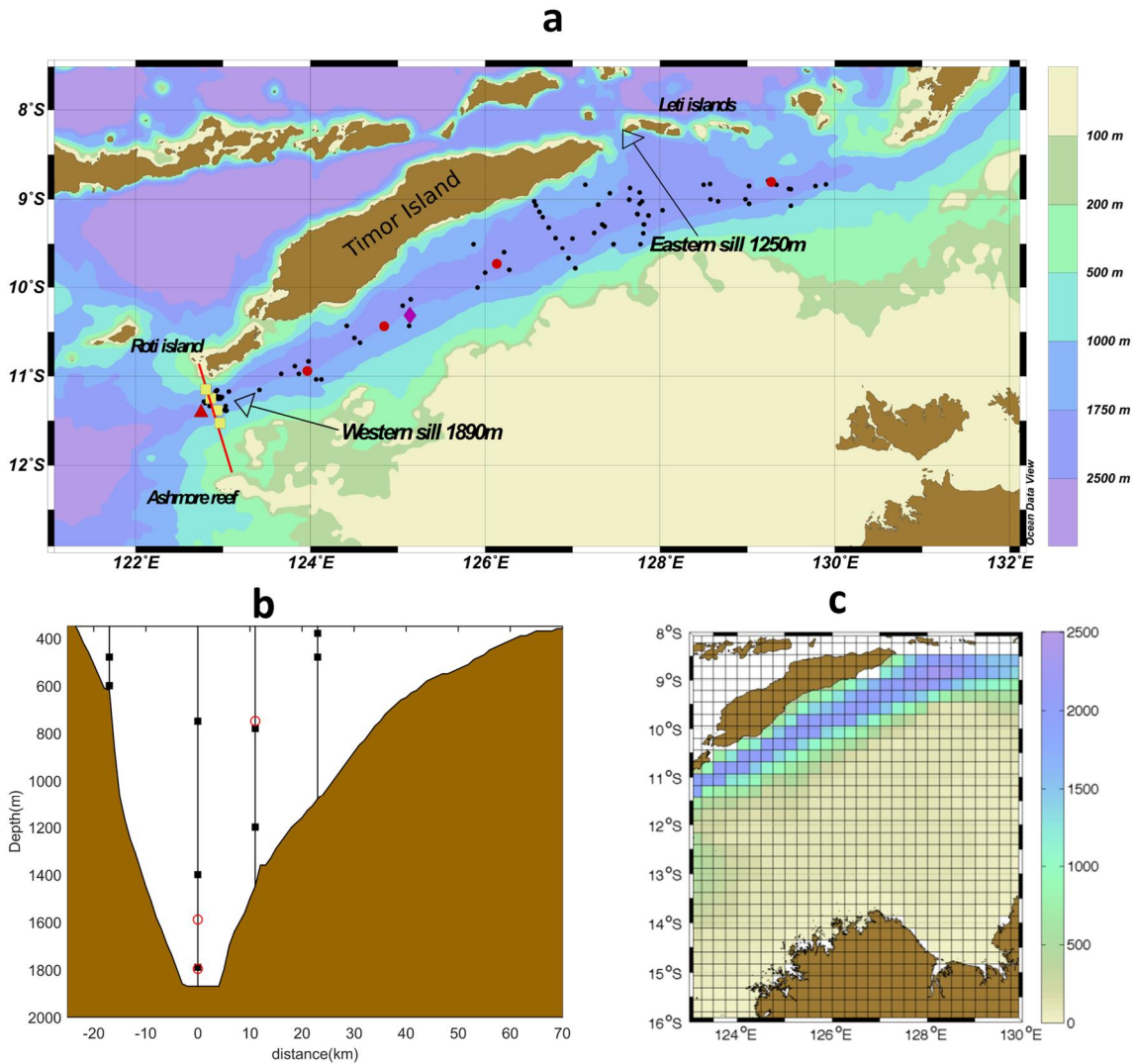
888

889

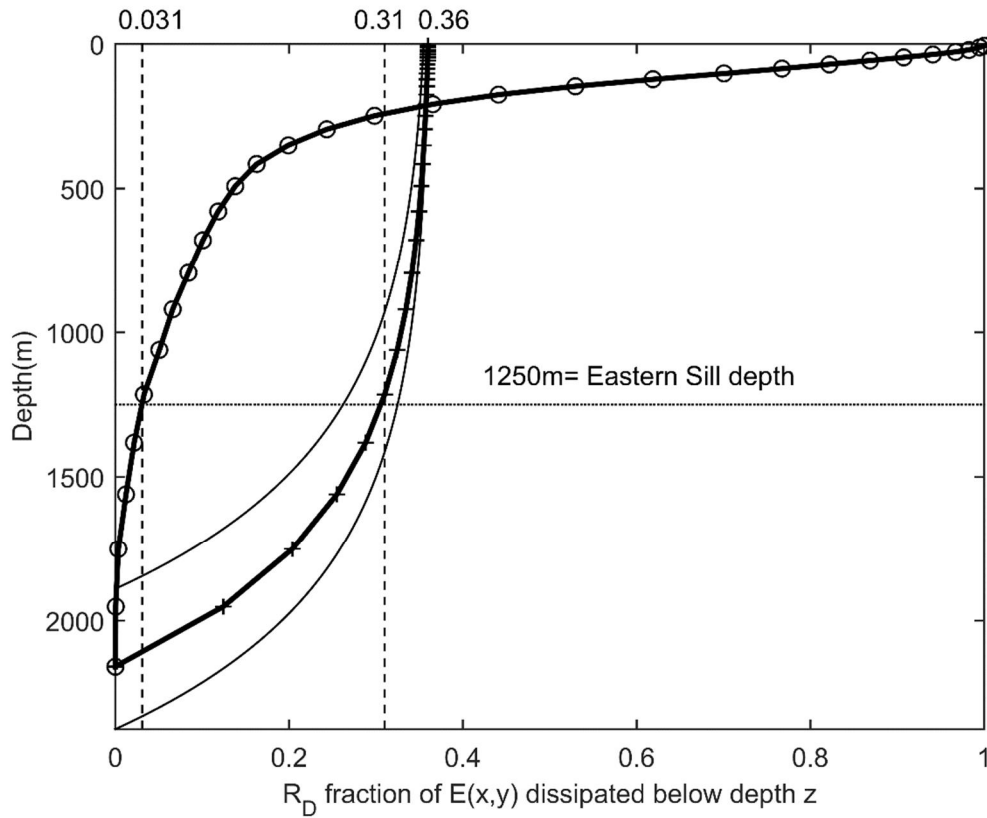
890

891

892



894
 895 **Figure 1.** Panel a: Bathymetry, yellow squares show the location of moored current-meter and temperature
 896 sensors arrays in the western Timor Sill (~1890 m sill depth) from INSTANT program 2004-2006, black dots
 897 show the location of CTD/OSD data obtained from the World Ocean Database 2013, the red dots show
 898 additional CTD from ATSEA 2010 campaign, the red triangle and purple diamond show the interpolation
 899 location of potential density data shown in Figure 6b. Panel b shows the cross-section bathymetry between Roti
 900 Island and Ashmore Reef and the mooring lines below 400 m. Black squares indicate the location of Aanderaa
 901 RCM07/RCM09 velocity and temperature sensors, open red circles indicate the location of Seabird SBE37/39
 902 temperature/conductivity sensors.
 903



904

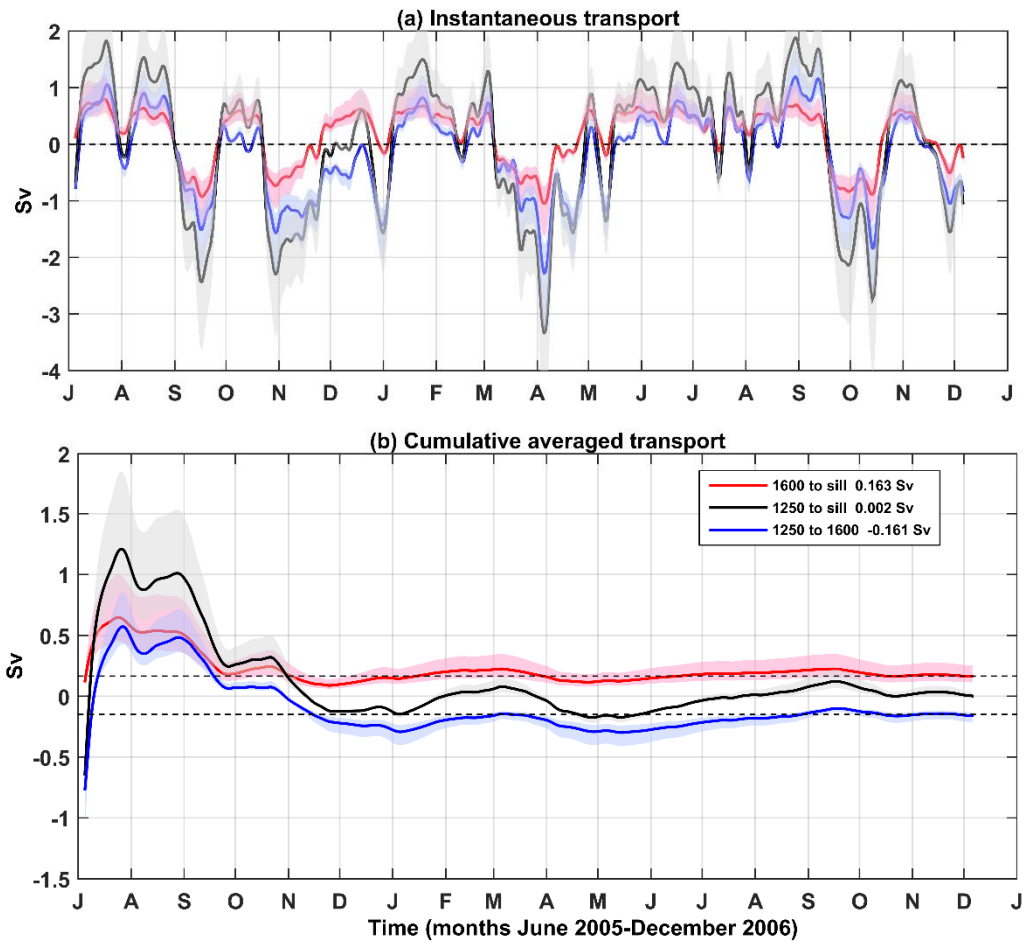
905

906

907

908

Figure 2 Ratio $R_D(z)$ of the integrated total dissipation rate (B_+) below depth z to the power input to internal tides for the Koch-Larrouy et al. [2007] parameterization (\circ symbol) and the St. Laurent et al. [2002] parameterization ($+$ symbol) considering a bottom depth of 2150 m (thick line), 1890 m (upper thin line) or 2375 m (lower thin line).



909

910

911 **Figure 3.** Instantaneous (a) and cumulative average (b) of the volume transport estimate from INSTANT

912 observations above western Timor Sill between 1250 and 1890 m sill (black), 1250 and 1600 m (red), and 1600

913 and 1890 m sill (blue). Shaded areas represent the error bars as estimated from various interpolation methods for

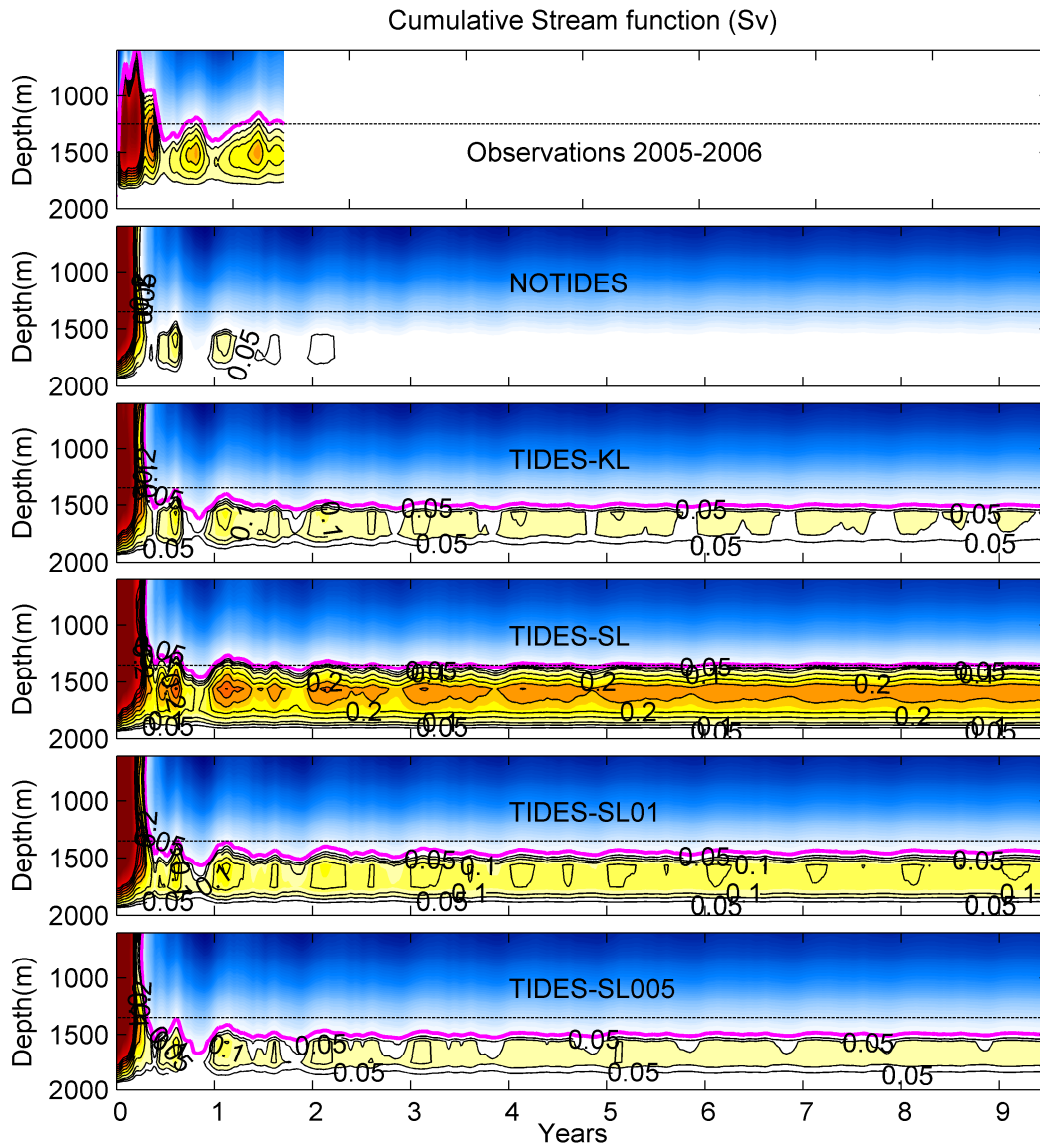
914 the velocity (see text for details).

915

916

917

918



919

920

Figure 4 Cumulative averaged transport below depth z for observations, NOTIDE, TIDE-KL, TIDE-SL02,

921

TIDE-SL01 and TIDE-SL005 simulations. The magenta line represents the limit of zero transport and bounds

922

the deep circulation cell, the black dotted line represents the Eastern sill depth (1350 m in the simulations and

923

1250 m in the observations).

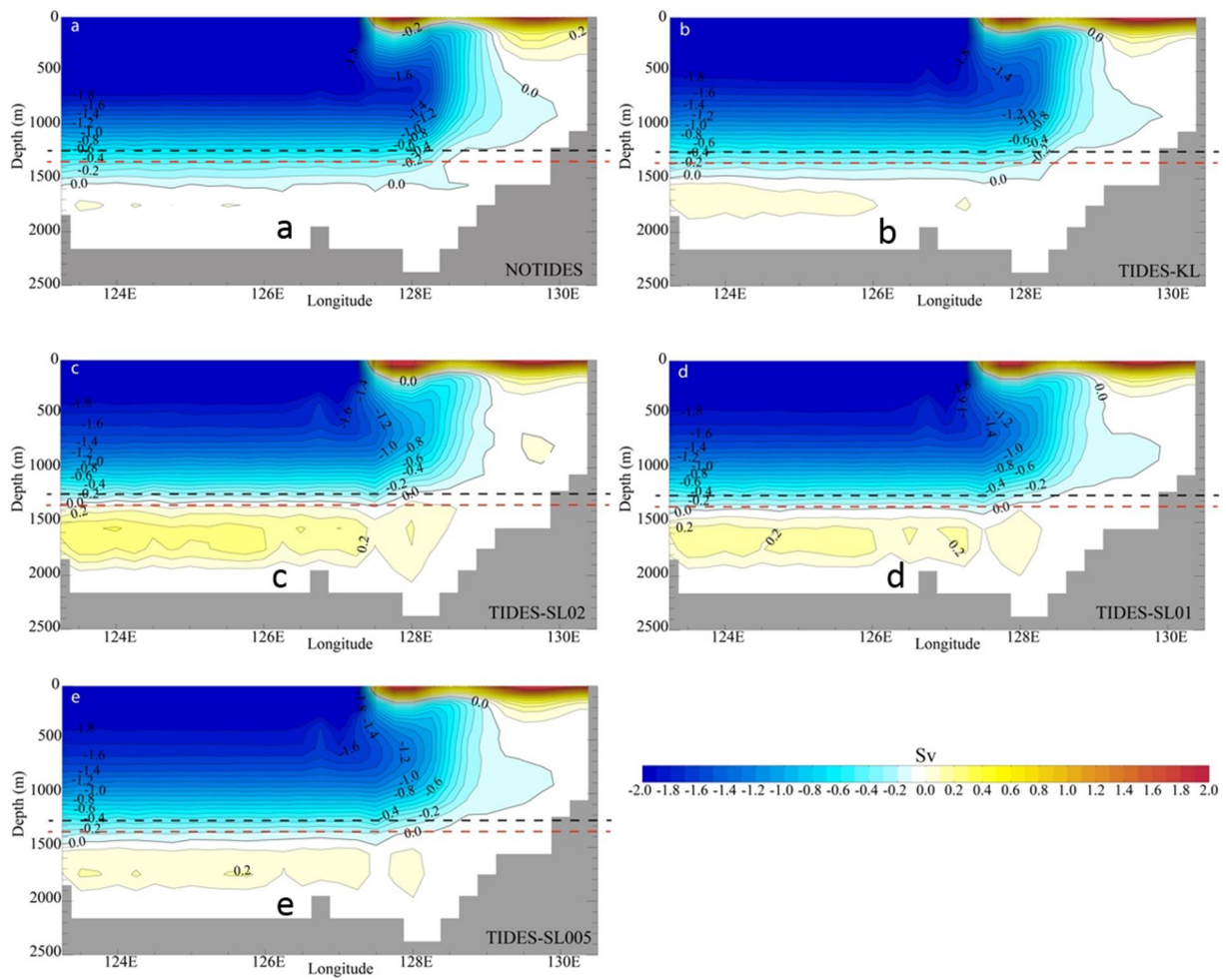
924

925

926

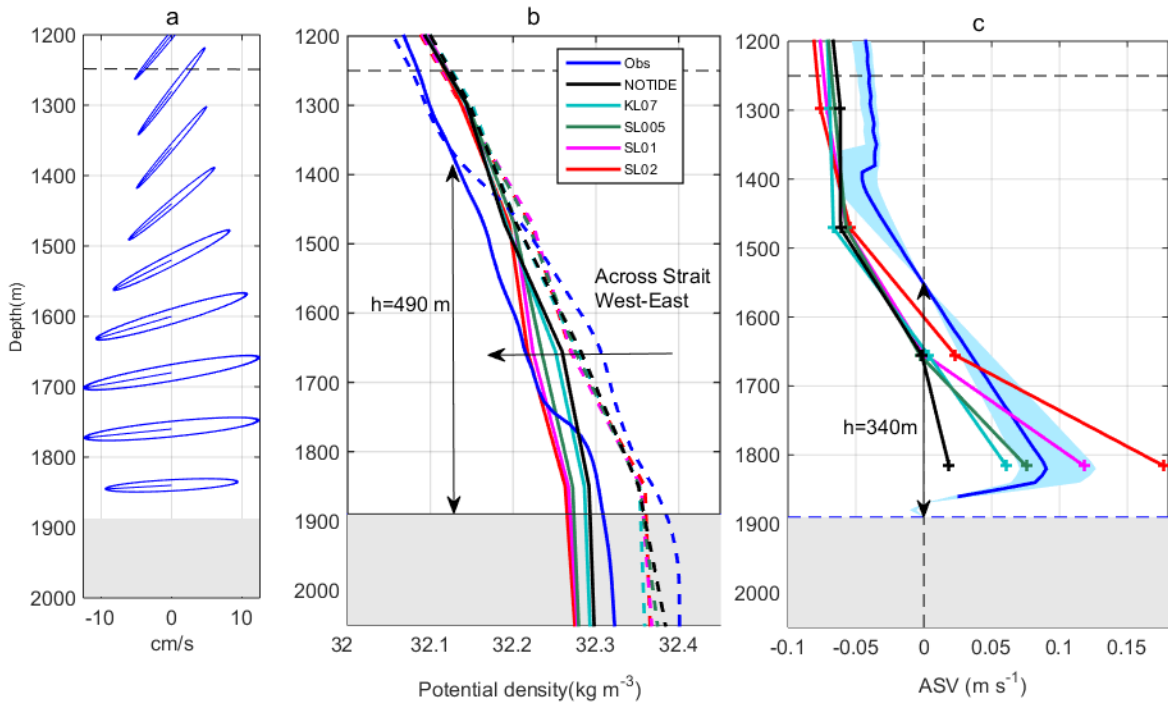
927

928



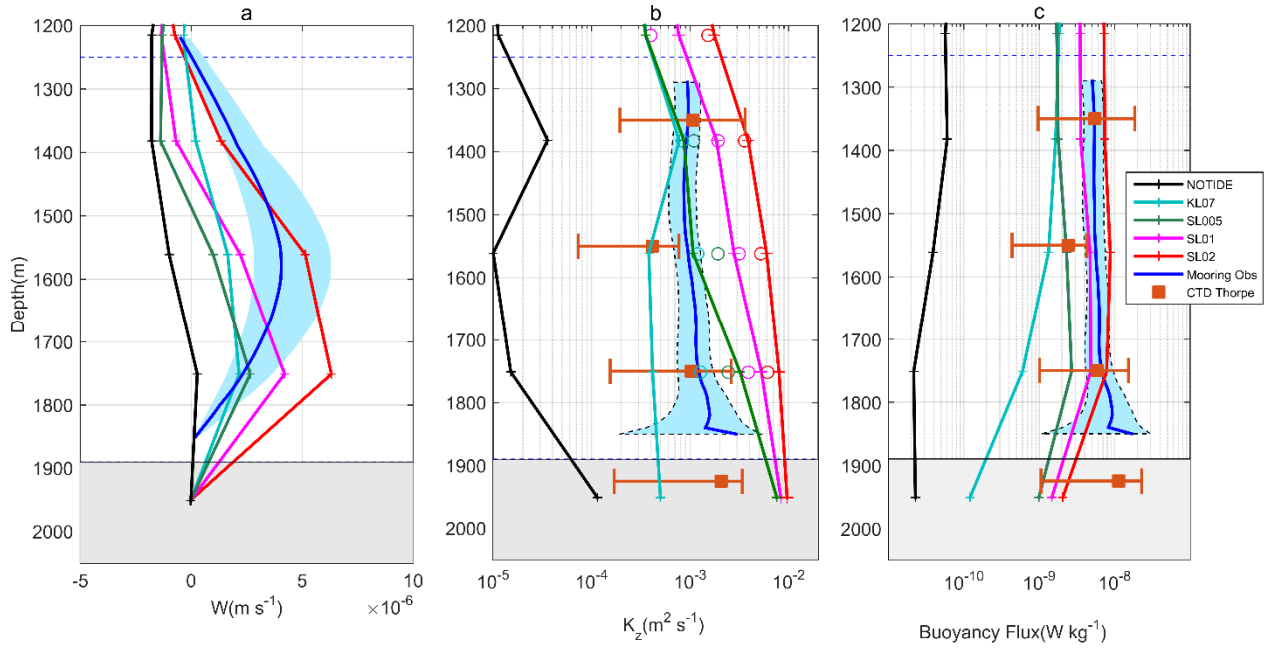
929
 930
 931
 932
 933
 934
 935

Figure 5. Zonal stream function calculated over the Timor Basin region (see Figure 1c) averaged over the last decade of simulation for NOTIDE (a), TIDE-KL07 (b), TIDE-SL02 (c), TIDE-SL01 (d), TIDE-SL005 (e) experiments. Labels of the contours indicate the zonal transport in Sv. Black and red dashed lines indicate the real and numerical eastern sill depth, respectively.



936
 937
 938
 939
 940
 941
 942
 943
 944
 945
 946
 947
 948
 949
 950
 951

Figure 6. (a) M2 Tidal ellipses from interpolated currents. (b) Potential (1000 db) density profile upstream the western sill [122.75°E, 11.42°S] in dashed line and downstream the sill within the Timor Basin [125°E, 10.44°S] in solid lines (the two position are indicated in Figure 1a). (c) Mean along strait velocity (ASV). For each panel, averages of the observations (blue), results from the NOTIDE (black), TIDE-SL02 (red), TIDE-SL01 (magenta), TIDE-SL005 (dark green) and TIDE-KL (light blue) simulations averaged over the last ten years of the run. The gray shading in (c) represents an error bar as estimated from various interpolations used for the velocity [Sprintall et al., 2009]. In each panel, the blue dashed line represents the western (1250 m) Timor sill depth, the gray band represents depth below the depth below the eastern sill (1890 m).



952

953

954

955

956

957

958

959

960

961

962

963

964

965

966

967

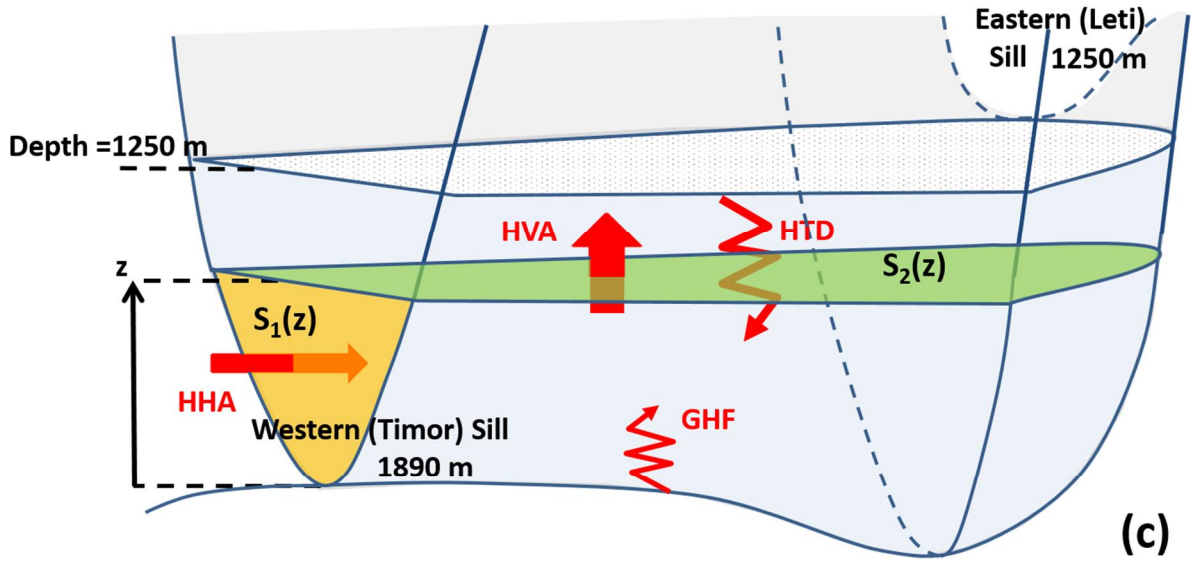
968

969

970

Figure 7 (a) Averaged vertical velocity across the S_2 surface ($S_2(z)$ is the horizontal surface across which the vertical heat transport across is decomposed in the vertical advection and turbulent fluctuation calculations (see text for details and sketch of Fig.8)), (b) mean vertical turbulent diffusivity K_z across S_2 surface, (c) mean buoyancy fluxes across S_2 surface. For each panel, averages of the observations (blue), results from the NOTIDE (black), TIDE-SL02 (red), TIDE-SL01 (magenta), TIDE-SL005 (dark green) and TIDE-KL (light blue) simulations averaged over the last ten years of the run. The circle markers in (b) indicates the K_z obtained for all the TIDE simulations by the heat budget estimation and the brown squares with 95% error bars in (b) and (c) the estimates from the Thorpe scale analysis. In (a), the gray shading represents an error bar as estimated from various interpolations used for the velocity, in (b) and (c) the gray shading is a 95 % confidence interval as estimated from the velocity and hydrological data error bounds (see text for details). In each panel, the blue dashed line represents the western Timor sill depth (1250 m) and the gray band represents depth below the eastern sill (1890 m)

971
972
973



$$\text{HHA} : S_1 C_p \langle \rho U T \rangle_{S_1, \tau} \quad \text{HVA} : -S_2 C_p \langle \rho \rangle_{S_2, \tau} \langle W \rangle_{S_2, \tau} \langle T \rangle_{S_2, \tau}$$

$$\text{HTD} : S_2 C_p \langle \rho \rangle_{S_2, \tau} \langle W' \theta' \rangle_{S_2, \tau} \cong S_2 C_p \langle \rho \rangle_{S_2, \tau} K_z \left\langle \frac{\partial \theta}{\partial z} \right\rangle_{S_2, \tau}$$

974
975
976
977

Figure 8 Schematic of the heat transport budget below 1250 m depth in the Timor Basin. HHA: Heat Horizontal Advection, HVA: Heat vertical advection, HTD: Heat Turbulent Diffusion, GHF: Geothermal Heat Flux.