



1 **Intermediate water masses, a major supplier of oxygen for the eastern tropical Pacific ocean**

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4

5 **Abstract**

6 It is well known that Intermediate Water Masses (IWM) are sinking in high latitudes and ventilate
7 the lower thermocline (500 – 1500 m depth). We here highlight how the IWM oxygen content and
8 the IWM pathway along the Equatorial Intermediate Current System (EICS) towards the eastern
9 tropical Pacific ocean are essential for the supply of oxygen to the lower thermocline and the
10 Oxygen Minimum Zones (OMZs). To this end, we assess here a heterogeneous subset of ocean
11 models characterized by a horizontal resolution ranging from 0.1° to 2.8°. Subtropical oxygen
12 levels in the lower thermocline, i.e., IWM are statistically correlated with tropical oxygen levels and
13 OMZs. Sensitivity simulations suggest that the oxygen biases of the subtropical IWM oxygen levels
14 contribute to oxygen biases of the tropical thermocline : an increase of the IWM oxygen by 60
15 mmol.m⁻³ results in a 10 mmol.m⁻³ increase in the tropical ocean in a timescale of 50 years. In the
16 equatorial regions, the IWM recirculates into the Equatorial Intermediate Current System (EICS).
17 By comparing tracer and particle release simulations, we show that a developed EICS increases
18 eastern tropical ventilation by 30 %. Typical climate models lack in representing crucial aspects of
19 this supply: biases in IWM properties are prominent across climate models and the EICS is
20 basically absent in models with typical resolutions of ~1°. We emphasize that these biases need to
21 be reduced in global climate models to allow reliable projections of OMZs in a changing climate.

22

23 **1. Introduction**

24 Oxygen levels in the ocean are characterized by high values in the high latitudes and the subtropical
25 gyres, while concentrations decrease to close to zero in the tropical oceans in the Oxygen Minimum
26 Zones (OMZs). While OMZs are natural features, climate change is potentially responsible for their
27 expansion (Breitburg et al., 2018), leading to a reshaping of the ecosystems and a potential loss of
28 biodiversity. In order to perform robust projections there is a need to better understand the processes
29 at play that are responsible for the supply of oxygen to the OMZ. We focus here on the Pacific
30 ocean, where the largest OMZs are located (Karstensen et al., 2008; Paulmier and Ruiz-Pino. 2009)

31

32 Oxygen rich waters are supplied into the ocean by subduction processes (Karstensen et al., 2008).
33 Oxygen solubility increases with lower temperatures, thus waters formed in the Southern Ocean
34 and in the North Pacific are characterized by particularly high oxygen values. In particular, the



36 Antarctic Intermediate Water (AAIW) (Molinelli, 1981) ventilates large areas of the lower
37 thermocline of the Pacific Ocean (Sloyan and Rintoul., 2001) and is characterized by oxygen values
38 larger than 300 mmol.m^{-3} at subduction time (Russel and Dickson, 2003). The oxygenated core of
39 the AAIW in the tropical Pacific is located at about 500-1200 m depth at 40°S (Russell and
40 Dickson, 2003) and with this at a depth directly below the depth of the OMZs in the eastern Pacific;
41 the Pacific AAIW mixes down to 2000 m depth with the Pacific Deep Water (PDW) as determined
42 by the OMP (Optimum Multiparameter) analysis (Pardo et al., 2012; Carrasco et al., 2017). The
43 oxygen rich ($> 200 \text{ mmol.m}^{-3}$ at 40°S) AAIW spreads from its formation side in the Southern Ocean
44 to the subtropical regions ; conversely the oxygen poor PDW (below 150 mmol.m^{-3}), extends till
45 3000m depth and recirculates poleward (Koshlyakov and Tarakanov, 2003). The northern part of
46 the Pacific basin is characterized by the North Pacific Intermediate Water (NPIW) (Talley, 1993)
47 confined to the northern Pacific conversely to the AAIW, which spreads far northward as its
48 signature reaches 15°N (Qu and Lindstrom., 2004). AAIW, NPIW and the upper part of the PDW
49 are oxygenated water masses occupying the lower thermocline between 500 and 1500 m depth. We
50 will refer to these waters as Intermediate Water Masses (IWM) in the following.

51

52 In the subtropics, the IWM (more particularly the AAIW) circulates into the intermediate flow of
53 the South Equatorial Current and the New Guinea Coastal Undercurrent (Qu and Lindstrom, 2004)
54 where it retroflects in the zonal equatorial flows of the Southern Intermediate Countercurrent
55 (SICC) and Northern Equatorial Intermediate Current (NEIC) within about $\pm 2^{\circ}$ off the equator
56 (Zenk et al., 2005; Kawabe et al., 2010) (Fig 1). These currents are part of the Equatorial
57 Intermediate Current System (EICS) constituted by a complex system of narrow jets extending
58 below 500 m in the lower thermocline (Firing, 1987; Ascani et al., 2010; Marin et al. 2010; Cravatte
59 et al., 2012, 2017; Menesguen et al., 2019). While the existence of this complex jet system has been
60 shown to exist in particular using argo floats displacements (Cravatte et al., 2017) the spatial
61 structure and variability of the jets are still largely unknown. In addition, there is little knowledge
62 about their role in transporting properties such as oxygen.

63

64 The simulation of the supply of oxygen to the eastern tropical Pacific is a difficult task as it depends
65 on the realistic simulation of the IWM properties (in particular the oxygen content) and the IWM
66 pathway (through the EICS). It is known that current climate models, in particular CMIP5
67 (Coupled Model Intercomparison Project phase 5) models, have deficiencies in correctly
68 representing the IWM, and in particular the AAIW. They generally display too shallow and thin
69 IWM, with a limited equatorward extension compared to observations (Sloyan and Kamenkovich,



70 2007; Sallee et al., 2013; Meijers, 2014; Cabre et al., 2015; Zhu et al., 2018 for the south Atlantic
71 ocean). Discrepancies in the simulated properties of IWM compared to observations are due to a
72 combination of many errors in the climate models, including simulation of wind and buoyancy
73 forcing, inadequate representation of subgrid-scale mixing processes in the Southern Ocean, and
74 midlatitude diapycnal mixing parameterizations (Sloyan and Kamakovich, 2007; Zhu et al., 2018).
75 In addition, the representation of the EICS is lacking in coarse resolution models (Dietze and
76 Loeptien, 2013; Getzlaff and Dietze, 2013). Higher resolution (0.25°, 1/12°) configurations partly
77 resolve the EICS but with a smaller amplitude than observed (Eden and Dengler, 2008; Ascani et
78 al., 2015). The mechanisms forcing the EICS are complex and still under debate (see the review by
79 Menesguen et al., 2019).

80

81 In this study we focus on the impact of IWM (and of the deficiencies in the representation of their
82 properties and transport) on the oxygen content in the eastern tropical Pacific in a set of model
83 simulations. Section 2 gives an overview of all models that we used as well as of the sensitivity
84 simulations. Next, we assess to which extent the IWM modulate (or drive) the oxygen levels in the
85 eastern tropical (20°S - 20°N) Pacific ocean in this set of models. The role of the IWM depends i)
86 on the oxygen content of the IWM in the lower thermocline of the subtropical regions (section 3)
87 and ii) on the zonal recirculation of the oxygen by the EICS toward the eastern part of the basin
88 (section 4). We conclude in section 5.

89

90 **2. Analyzed models and experiments**

91 2.1 Mean state

92 We analyze the mean state of the oxygen fields, OMZ, EICS of the following model experiments
93 (see Table 1), which previously have been used in recent studies focusing on the understanding of
94 the tropical oxygen levels mean state or variability:

95 - a NEMO (Nucleus for European Modelling of the Ocean) configuration with a resolution of 2°,
96 refined meridionally to 0.5° in the equatorial region (NEMO2). The circulation model is coupled to
97 a simple NPZD (Nutrient Phytoplankton Zooplankton Detritus) biogeochemical model that
98 comprises 6 compartments (e.g used in Duteil et al., 2018; Duteil, 2019). The simulation has been
99 forced by climatological forcings based on the Coordinated Reference Experiments (CORE) v2
100 reanalysis (1948-2007) (Large and Yeager, 2009) and integrated for 1000 years.

101 - the UVIC (University of Victoria) model (e.g used in Getzlaff et al., 2016; Oschlies et al., 2017),
102 an earth System Model (ESM) that has a horizontal resolution of 1.8° latitude x 3.6° longitude. The
103 experiment has been integrated for 10000 years. The biogeochemical model is a NPZD-type model



104 of intermediate complexity that describes the full carbon cycle (see Keller Keller 2012 for a detailed
105 description). This model is forced by monthly climatological NCAR/NCEP wind stress fields.
106 - the GFDL (Geophysical Fluid Dynamics Laboratory) CM2-0 suite (Delworth et al., 2012; Griffies
107 et al., 2015, Dufour et al, 2015): the suite is based on the GFDL global climate model and includes
108 a fully coupled atmosphere with a resolution of approximately 50 km. It consists of three
109 configurations that differ in their ocean horizontal resolutions: GFDL1 with a nominal 1°
110 resolution, GFDL025 with a nominal 0.25° and GFDL01 with a nominal 0.1° resolution (e.g used in
111 Frenger et al., 2018 and Busecke et al., 2019 for studies on ocean oxygen). The climate models are
112 forced with preindustrial atmospheric pCO₂ concentrations. At simulation year 48, the simplified
113 ocean biogeochemistry model miniBLING is coupled to the models, with three prognostic tracers,
114 phosphate, dissolved inorganic carbon and oxygen (Galbraith et al., 2015). Due to the high
115 resolution of GFDL01, the integration time is limited. We here analyze simulation years 186 to 190.
116 The heterogeneity of the configurations that we analyze permits to determine whether the simulated
117 oxygen distributions of the models display systematic biases or are strongly configuration
118 dependent (e.g dependent on resolution).

119

120 2.2 Sensitivity simulations

121 In order to disentangle the different processes at play we perform two different sets of sensitivity
122 simulations with one of the models, NEMO. NEMO allows to test effects of increasing the ocean
123 resolution and to integrate the model over a relatively long time span. All sensitivity experiments
124 are integrated for more than 50 years (1948 to 2007) using the CORE (Coordinated Ocean-Ice
125 Reference Experiments) v2 interannual (Large and Yeager, 2009) forcings. This time scale permits
126 the recirculation from the interior subtropical regions to the tropical area (as suggested in the model
127 study by SenGupta and England, 2007).

128

129 2.2.1 Oxygen restoring to observations in the subtropical regions

130 In the first set of experiments the focus is on the role of the lower thermocline oxygen content for
131 the ventilation of the eastern equatorial Pacific. We use NEMO2, the oceanic component of the
132 IPSL-CM5A (Mignot et al., 2013), that is part of CMIP5. NEMO2 shows mid-latitudes oxygen
133 biases consistent with CMIP5 models. We compare three experiments :
134 - NEMO2-REF: the experiment is integrated from 1948 to 2007 starting from the spinup state
135 described in 2.1.



136 - NEMO2-30DEG: the oxygen boundaries are restored to observed oxygen concentrations (WOA)
137 at the boundaries 30°N and 30°S : the mid-latitude supply of oxygen by the IWM is therefore
138 correctly represented.

139 - NEMO2-30DEG1500M: same as NEMO2-30DEG ; in addition oxygen is also restored at the
140 depth interface of 1500m, mimicking a correct oxygen state of the deeper water masses (lower part
141 of the AAIW, upper part of the PDW)

142

143 We focus with the above three experiments on the transport of IWM oxygen levels to the tropical
144 ocean and the OMZs. The respiration rate (oxygen consumption) is identical in NEMO2-REF,
145 NEMO2-30DEG and NEMO2-30DEG1500M in order to avoid compensating effects between
146 supply and respiration that depends on biogeochemical parameterizations (e.g Duteil et al., 2012).
147 We aim to avoid such compensating effects to ease interpretation and be able to focus on the role of
148 physical transport.

149

150 2.2.2 Conservative Tracer Release in oxygenated waters

151 In a second set of experiments, we performed tracer release experiments in a coarse 0.5°
152 (NEMO05) and high resolution 0.1° (NEMO01) configurations of NEMO (Table 1) to examine the
153 transport of oxygenated IWM from the subtropical regions into the oxygen deficient tropics.
154 NEMO01 is a configuration based on NEMO05 and where a 0.1° two-way nest has been
155 embedded in the whole Pacific Ocean, from 49°S to 31°N (Czeschel et al, 2011). In these
156 experiments, we initialized the regions with climatological (WOA) oxygen levels greater than 150
157 mmol.m⁻³ with a value of 1 (and 0 when oxygen is lower than 150 mmol.m⁻³). In the model
158 simulations, the tracer is subject to the same physical processes as other physical and
159 biogeochemical tracers, i.e. advection and diffusion but it does not have any sources and sinks. The
160 experiments have been integrated for 60 years (1948 – 2007) using realistic atmosphere forcing
161 (COREv2). NEMO05 and NEMO01 display a similar upper ocean circulation (Fig 5) but NEMO05
162 does not simulate a developed EICS contrary to NEMO01.

163

164 In order to complement the tracer experiment we performed Lagrangian particle releases.
165 Lagrangian particles allow to trace the pathways of water parcels due to the resolved currents, and
166 to track the origin and fate of water parcels. They are not affected by subgrid scale mixing
167 processes. The particles are advected offline with 5 daily means of the NEMO05 and NEMO01
168 currents. The NEMO01 circulation fields have been interpolated on the NEMO05 grid in order to
169 allow a comparison of the large scale advective patterns between NEMO01 and NEMO05. We used



170 the ARIANE tool (Blanke and Raynaud, 1997). A first particle release has been performed in the
171 eastern tropical OMZ at 100°W in the tropical region between 5°S - 5°N, a second release has been
172 performed in the western part of the basin at 160°E. The particles have been released in the lower
173 thermocline at 1000 m and integrated backward in time from 2007 to 1948 in order to determine
174 their pathways and their location of origin. We released 120 particles every 5 days during the last
175 year of the experiment, for a total of 8760 particles.

176

177 **3. Intermediate water properties and oxygen content**

178 3.1. IWM Oxygen levels in models

179 The IWM subducted in mid/high latitudes are highly oxygenated waters. As part of the deficient
180 representation of IWM, the subducted “oxygen tongue” (oxygen values up to 240 mmol.m⁻³) is not
181 reproduced in most of the models part of CMIP5 (Fig 8 from Cabre et al., 2015, Fig 4 from Takano
182 et al., 2018) and in the models analyzed here (Fig 2a), with biases of about 20-60 mmol.m⁻³
183 (NEMO2, GFDL1, GFDL025, GFDL01). UVIC, a coarse resolution model, shows oxygenated
184 waters in the lower thermocline at mid latitudes (30°S-50°S); the oxygenation however arises due to
185 a too large vertical diffusion from the mixed layer rather than by an accurate representation of the
186 water masses. GFDL01, even though still biased low, presents larger oxygen values than the coarser
187 resolution models GFDL1, GFDL025 and NEMO2. A possible explanation is a better
188 representation of the water masses and in particular the AAIW in eddy-resolving models (Lackhar
189 et al., 2009). The IWM oxygen maximum is apparent at 30°S throughout the lower thermocline in
190 observations (Fig 2b), consistent with the circulation of IWM with the gyre from the mid/high
191 latitude formation regions towards the northwest in subtropical latitudes, and followed by a
192 deflection of the waters in the tropics towards the eastern basin.

193

194 Consistent with the low oxygen bias of models at subtropical latitudes (Fig 2b), models also feature
195 a bias in the tropical ocean (20°S-20°N) by 20 – 50 mmol.m⁻³ (Fig 2a, Fig 2c) at intermediate depths
196 in the eastern part of the basin (similarly to CMIP5 models, as shown by Cabre et al., 2015). The
197 basin zonal average of the mean oxygen level in the lower thermocline (layer 500 - 1500) m at 30°S
198 and in the eastern part of the basin (average 20°S – 20°N, 160°W-coast; 500-1500 m) are positively
199 correlated (Pearson correlation coefficient R=0.73) (Fig 2d), suggesting a large role of the IWM in
200 controlling the oxygen levels in the tropical oceans.

201

202 The models presenting the poorest oxygenated water at 30°S display the largest volume of OMZs
203 (GFDL025 and GFDL1), though the negative correlation (Pearson correlation coefficient R=-0.52)



204 is less pronounced between the volume of the OMZs and the mean oxygen levels in the layer 500 -
205 1500 m at 30°S (Fig 2e). Reasons of this weaker correlation are due to the OMZs being a result of
206 several processes next to oxygen supply by IWM, e.g, vertical mixing with other water masses
207 (Duteil et al., 2011), isopycnal mixing in the upper thermocline (Gnanadesikan et al., 2013; Bahl
208 al., 2019), supply by the upper thermocline circulation (Shigemitsu et al., 2017; Busecke et al.,
209 2019). A correlation, even weak, suggests a major role of the IWM in regulating the OMZ volume
210

211 In order to better understand the role of IWM entering the subtropical domain from higher latitudes
212 for the oxygen levels in the eastern tropical Pacific Ocean, we perform sensitivity experiments (see
213 2.2.1) in the following.

214

215 3.2 Sensitivity of tropical IWM oxygen to subtropical and deep oxygen levels

216 3.2.1 Oxygen levels in the lower thermocline

217 The difference of the experiments NEMO2-30DEG – NEMO2-REF (average 1997-2007) allows to
218 quantify the effect of model biases of IWM at mid latitudes (30°N/30°S). As we restore oxygen to
219 observed levels at 30°S/°N (see 2.2.1), the difference shows a large anomaly in oxygen levels at
220 30°S (more than 50 mmol.m⁻³) at lower thermocline level (500 – 1500 m) corresponding to the
221 missing deep oxygen maximum (Fig 3). The northern negative anomaly results from a deficient
222 representation of the north Pacific OMZ, i.e., modeled oxygen is too high for NPIW. The northern
223 low and southern high anomalies spread towards the tropics at intermediate depth. A fraction of the
224 positive oxygen anomaly recirculates at upper thermocline level due to a combination of upwelling
225 and zonal advection by the tropical current system (for instance the EUC at thermocline level is a
226 major supplier of oxygen as shown in observations by Stramma et al., 2010 and in ocean models by
227 Duteil et al., 2014, Busecke et al., 2019).

228

229 The difference NEMO2-30DEG1500M – NEMO2-30DEG (Fig 3f-h) shows a deep positive
230 anomaly in oxygen, as oxygen levels are lower than in observations by 30-40 mmol.m⁻³ in the
231 eastern tropical regions. This anomaly is partially transported into the lower thermocline (500 -
232 1500 m). It shows that a proper representation of the deep oxygen level (> 1500 m) is important for
233 a realistic representation of the lower thermocline and OMZs. Causes of the oxygen bias of the
234 deeper water masses are beyond the scope of this study but may be associated with regional
235 (tropical) issues, such as an improper parameterization of respiration (e.g a too deep
236 remineralisation) (Kriest et al., 2010), or a misrepresentation of deeper water masses.

237



238 3.2.2 Oxygen budget and processes

239 To assess the processes that drive the oxygen content of the (sub)tropical lower thermocline, we
240 analyzed the oxygen budget in NEMO2-REF and NEMO2-30DEG. The budget is calculated as an
241 average between 500 and 1500m and shown in Fig.4. In NEMO2-REF, the physical oxygen supply
242 is balanced by the respiration. The oxygen supply in the model is divided into advection, i.e.,
243 oxygen transport associated with volume transport, and isopycnal diffusion, i.e., subgrid scale
244 mixing processes that homogenize oxygen gradients. Diapycnal diffusion is comparatively small
245 and can be neglected. The lower branches of the subtropical gyres transport the oxygen from the
246 eastern to the western part of the basin (Fig 4a,b). Downwellings from the oxygen-rich upper layer
247 supply the interior of the subtropical gyres (Fig 4c). At the equator, the EICS transport westward
248 oxygen-poor water originating in the eastern side of the basin (Fig 4a). Concomitantly, the
249 meridional advection term transports oxygen originating from the subtropics in the tropical regions
250 (Fig 4b), which is upwelled (Fig 4c). Isopycnal diffusion transfers oxygen from the oxygen-rich
251 gyres to the poor oxygenated regions (Fig 4d).

252

253 The anomalies generated at 30°S and 30°N by the restoring experiment generate a disbalance
254 between respiration (which remains identical in NEMO2-REF and NEMO2-30DEG) and supply.
255 This disbalance is most apparent in the tropics by an increase (south) or decrease (north) of
256 isopycnal diffusion (Fig 4g, Fig 3h). Changes in the advective terms can be found along the
257 equator: as the vertical gradient of oxygen decrease (the intermediate ocean being more
258 oxygenated), the vertical supply from the upper ocean decreases in the south (increases in the north)
259 subtropical gyre and decreases at the equator (Fig 4g). The meridional oxygen gradient between the
260 southern subtropical gyre and the equator strengthens, and so the meridional transport from the
261 subtropics to the equator, partly by the western boundary currents (Fig 4f). The changes in zonal
262 transport are comparatively small (Fig 4e). The total advective term does not show significant
263 change however (Fig 4g).

264

265 In the experiment NEMO2-30DEG1500, in complement to the isopycnal propagation of the
266 subtropical anomaly, the deep (> 1500 m) oxygen anomaly is upwelled in the eastern equatorial
267 (500 – 1500 m) part of the basin (see Fig 4i) showing a large increase in advective terms, mostly
268 due to an increase in vertical advection), consistent with the analysis by Duteil (2019) who showed
269 that vertical advection is the dominant process to supply oxygen from the lower to the upper
270 thermocline in the equatorial eastern Pacific Ocean in a similar NEMO2 configuration.

271



272 This simple set of experiment shows that in climate models oxygen in the lower thermocline (500 –
273 1500 m) ocean are partially controlled by properties of IWM that enter the tropics from higher
274 latitudes. This presumably applies to other (biogeochemical) tracers. IWM oxygen propagates
275 equatorward mostly by small scale isopycnal processes and the western boundary currents. Further,
276 upwelling in the tropics from deeper ocean layers (Pacific Deep Water, partially mixed with the
277 lower IWM) play an important role . Our budget analysis highlighted the importance of advective
278 processes in the equatorial region in the lower thermocline which we will examine more closely in
279 the following.

280

281 **4. Equatorial intermediate current system and oxygen transport**

282 4.1 Structure of the currents in the upper 2000 m in observations and models

283 The current structure of the models analyzed in this study (see section 2.1, Table 1) is shown in Fig.
284 5. In the upper layer, the broad westward drifting South and North Equatorial Currents (SEC, NEC)
285 characterize the equatorial side of subtropical gyres. In the thermocline, the eastward flowing
286 equatorial undercurrent (EUC), flanked by the westward flowing south and north counter currents
287 are present in all models. This upper current structure is well reproduced (i.e the spatial structure
288 and intensity are consistent with observations) across the different models (see 2.1 “Model
289 analyzed”) compared to observations. Previous studies already discussed the upper thermocline
290 current structure in the GFDL models suite (Busecke et al., 2019), NEMO (e.g Izumo, 2005,
291 Lübbecke et al., 2008), UVIC (Loeptien and Dietze, 2013); the upper thermocline will not be
292 further discussed in this study.

293

294 At intermediate depth, in the observations, a relatively strong (about 0.1 ms^{-1}) westward flowing
295 Equatorial Intermediate Current (EIC) is present below the EUC at about 400-600 m depth (Marin
296 et al., 2010). A complex structure of narrow and vertically alternating jets every 200 m, so-called
297 Equatorial Deep Jets (EDJ), extends below the EIC till 2000 m (Firing, 1987; Cravatte et al., 2012).
298 Laterally to the EIC, in the upper thermocline, the Low Latitude Subsurface Countercurrents
299 (LLSC) are observed. They include the North and South Subsurface Counter Currents (NSCC and
300 SSCC), located around $5^{\circ}\text{N}/5^{\circ}\text{S}$, and a series of jets between $5^{\circ}\text{N}/\text{S}$ and $15^{\circ}\text{N}/\text{S}$ (in particular the
301 Tschuya jets in the southern hemisphere, described by Rowe et al., 2000). Below the LLSCs, the
302 Low Latitude Intermediate Currents (LLICs) include the a series of westward and eastward zonal
303 jets (500–1500-m depth range) alternating meridionally from 3°S to 3°N ; the North and South
304 Intermediate Countercurrents (NICC and SICC) flow eastward at 1.5° – 2° on both flanks of the
305 lower EIC. The North and South Equatorial Intermediate Currents (NEIC and SEIC) flow westward



306 at about 3° (Firing, 1987). A detailed schematic view of the tropical intermediate circulation is
307 shown in a recent review by Menesguen et al. (2019) and in Fig 1.

308

309 In coarse resolution models, the intermediate current system is not developed and sluggish (even
310 missing in UVIC and GFDL1). NEMO2 and NEMO05 display a “primitive” EICS as the LLSCs
311 are not represented. High resolution models (GFDL025, GFDL01, NEMO01) display a more
312 realistic picture, even if the mean velocity is still weaker than in observations (smaller than $5 \text{ cm} \cdot \text{s}^{-1}$
313 1), where it reaches more than 10 cm^{-1} at 1000 m (Ascani et al., 2010; Cravatte et al., 2017). An
314 interesting feature is that the jets are broader and faster in NEMO01 than in GFDL01. Possible
315 causes include a different wind forcing, mixing strength or topographic features as all these
316 processes play a role in forcing the intermediate jets (see the review by Menesguen et al., 2019).
317 The intermediate currents are less consistent vertically in NEMO01 than in GFDL01, due to their
318 large temporal variability in NEMO01. A strong seasonal and interannual variability of the EICS
319 has been observed that display varying amplitudes and somewhat positions of the main currents/jets
320 (Firing, 1998; Gouriou et al., 2006; Cravatte et al., 2017). A clear observational picture of the EICS
321 variability is however not yet available. Outside the tropics (in particular south of 15°S), the interior
322 velocity pattern is similar in coarse and high resolution models, suggesting a similar equatorward
323 current transport at intermediate depth in the subtropics, in for instance NEMO05 and NEMO01.

324

325 4.2 Transport by the EICS

326 4.2.1 Tracer spreading towards the eastern tropical Pacific

327 We released a conservative tracer in the subtropical domain in well oxygenated waters (see 2.2.2) in
328 a coarse (NEMO05) and a high resolution configuration (NEMO01). The tracer does not have
329 sources or sinks and is advected and mixed as any other model tracer and allows to assess the
330 spreading of tracer (such as oxygen) from oxygenated waters into the oxygen deficient eastern
331 tropical Pacific.

332

333 The ventilation by the oxygen rich waters, and in particular the IWM, is illustrated by the tropical
334 tracer concentration after 50 years (Fig 6a-c) of integration (mean 2002-2007). Concentrations
335 decrease from the release location to the northern part of the basin, where the lowest values (below
336 0.1) are located in NEMO05 and NEMO01. The 0.1 isoline is however located close to the equator
337 in NEMO05 while it is found around 7°N in NEMO01. This feature is associated with a pronounced
338 tongue of high tracer concentration (> 0.2) between 5°N and 5°S in NEMO01. Such a tongue is



339 absent in NEMO05. The enhanced tracer concentration in the equatorial region suggests a stronger
340 equatorial ventilation in NEMO01.

341

342 The preferential pathways of transport are highlighted by the determination of the transit time it
343 takes for the tracer to spread from the oxygen rich regions to the tropical regions. We define a
344 threshold called $t_{10\%}$ when the tracer reaches a concentration of 0.1 (Fig 6d-f) (similar to the
345 approach of SenGupta and England, 2007). $t_{10\%}$ highlights a faster ventilation of the equatorial
346 regions in NEMO01 compared to NEMO05, as $t_{10\%}$ displays maximum value of 10 (western part)
347 to 30 years (eastern part) between 5°N/5°S in NEMO01 compared to 30 years to more than 50 years
348 in NEMO05. The southern “shadow zone” is well individualized in NEMO01 compared to
349 NEMO05 as the oxygen levels are high in the equator in NEMO01, suggesting a strong transport by
350 the EICS. While $t_{10\%}$ increases linearly at intermediate depth at 100°W in NEMO05 from 20°S to
351 the equator, suggesting a slow isopycnal propagation (consistent with the experiments performed
352 using NEMO2 in part 3.2), the tracer accumulation is faster in the equatorial regions than in the
353 mid-latitudes in NEMO01, suggesting a large role of advective transport, faster than a transport by
354 diffusive processes.

355

356 4.2.2 Equatorial lower thermocline water mass origin

357 Lagrangian Particles (see 2.2.3) allow us to understand the origin of the waters in the lower
358 thermocline. They also allow to disentangle the transport of the resolved currents of the EICS
359 (advection) from subgrid scale mixing processes, i.e. to assess the processes responsible for the
360 equatorial ventilation. Two releases have been performed in the eastern and western part of the
361 basin in order to assess the equatorial circulation in NEMO05 and NEMO01.

362

363 The particles of the first release in the eastern tropical Pacific (100°W, at 1000 m depth where the
364 EICS are located) origin from the intermediate eastern tropical pacific (IETP) ocean (160°W – coast
365 / 10°N-10°S / 200 – 2000 m) close to the region of release, in 60 % of the cases in NEMO05 and
366 50 % of the cases in NEMO01, at a time scale of 50 years (Fig 7a-c and 8a). In NEMO05, after 50
367 years, the particles originating outside the IETP come either from the upper (0 – 200 m) ocean (5
368 %), deep (> 2000 m) ocean (1%), higher (> 10°) latitudes (23 %), western (west of 160°W) part of
369 the basin (21 %) (Fig 8c). The largest difference between NEMO05 and NEMO01 is the much
370 larger amount of particles originating from the deep ocean in NEMO01 (8 % in NEMO01),
371 suggesting the presence of vertical recirculation cells at intermediate depths. The advection
372 processes are also considerably faster in NEMO01, in particular the zonal advective ones. The



373 relative difference between NEMO05 and NEMO1 is particularly strong 15 years after the release
374 (approximately corresponding to the $t_{10\%}$ at 1000 m at the equator in NEMO01), as already 10 %
375 of the particles originate outside the IETP, in regions where the oxygen levels are high, in NEMO01
376 while this fraction is close to 0 in NEMO05.

377

378 The particles of the second release (160°E, 1000 m depth) are originally located in the intermediate
379 western tropical Pacific Ocean (IWTP) (160°W – coast / 10°N-10°S / 200 – 2000 m) (Fig 7e-f).
380 After 50 years, all the particles originate outside of this box in NEMO01 (Fig 8b) (50 % originate in
381 the eastern basin, 23 % in the deep ocean, 24 % outside the equatorial band, 3 % in the upper 200
382 m) (Fig 8d) while only 70 % of the particles originate outside the IWTP in NEMO05 (39 % in the
383 eastern basin, 27 % outside the equatorial band, 2 % in the deep ocean and 2 % in the upper ocean).

384

385 The Lagrangian particle results point to a generally stronger ventilation at intermediate depth in
386 NEMO01 due to the EICS, which reinforce the connection between western / eastern part of the
387 basin and thermocline / deep ocean

388

389 4.3 Equatorial oxygen levels in models

390 Our analyses above permit to better understand the distribution of the oxygen levels at the equator
391 in a suite of models characterized by an increasing resolution, such as the GFDL model suite. The
392 striking difference between GFDL01 and GFDL025 / GFDL1 are the high oxygen levels in the
393 eastern part of the ocean below 1000 m in GFDL01 compared to GFDL025/GFDL1 (Fig 2). The
394 oxygen levels are also more homogeneous zonally in GFDL01, with a weaker east/west gradient,
395 consistent with the tracer experiment that we performed in 4.2. The oxygen distribution fits with the
396 mean kinetic energy of the intermediate currents below 1000 m (Fig 9a), especially in the eastern
397 part of the basin (Fig 9b). Resolving the EICS results in similar results as what Getzlaff and Dietze
398 (2013) achieve with a simple parameterization of the EICS (Fig 9). To compensate for the
399 “missing” EICS in coarse resolution models, they enhanced anisotropically the lateral diffusivity in
400 the equatorial region. Implementing this approach tends to homogenize oxygen levels zonally,
401 with an increase of the mean levels by 30-50 mmol.m^{-3} in the eastern basin and a decrease of
402 oxygen concentrations in the western basin.

403

404 A possibly not intuitive feature is that the oxygen levels are relatively similar in GFDL025 and
405 GFDL1, while the current system is relatively similar in GFDL025 and GFDL01 (see Fig 5 and Fig
406 9). An explanation lies in the relatively small net balance between large fluxes of respiration and



407 oxygen supply (Duteil et al., 2014). If the supply is slightly higher compared to the consumption by
408 respiration, it will lead to an increase of oxygen concentration. If it is slightly lower, the oxygen
409 levels will decrease. A small difference in supply (e.g slightly weaker currents) may therefore lead
410 to a large difference in oxygen levels when integrated over decades. For this reason, the impact of
411 the EICS is more visible below 1000 m as the respiration decreases following a power-law with
412 depth (Martin et al., 1987) and is therefore easier to offset even by a moderate oxygen supply.

413

414 **5. Summary and conclusions**

415 Intermediate water masses (IWM) are subducted in the Southern Ocean and transported
416 equatorward to the tropics by isopycnal processes (Sloyan and Kamenskovich, 2007; Sallee et al.,
417 2013; Meijers, 2014). At lower latitudes they recirculate into the lower thermocline of the tropical
418 regions at 500 - 1500 m and into the EICS (Zenk et al., 2005; Marin et al., 2010; Cravatte et al.,
419 2012; 2017; Ascani et al., 2015; Menesguen et al., 2019) (see schema Fig 1). We show here that the
420 representation of this ventilation pathway is important to take into account when assessing tropical
421 oxygen levels and the extent of the OMZ in coupled biogeochemical circulation or climate models.
422 Particularly, we highlight two critical, yet typical, biases that hamper the correct representation of
423 the tropical oxygen levels.

424

425 5.1 Subducted IMW properties and tropical oxygen

426 First, the current generation of climate models, such as the CMIP5 models, show large deficiencies
427 in simulating IMW. Along with an unrealistic representation of IMW volume and properties when
428 the waters enter the subtropics, the models also lack the observed prominent oxygen maximum
429 associated with IMW. Restoring oxygen levels to observed concentrations at 30°S/30°N and at
430 1500 m depth in a coarse resolution model, comparable to CMIP5 climate models in terms of
431 resolution and oxygen bias, shows a significant impact on the lower thermocline (500 – 1500 m)
432 oxygen levels : a positive anomaly of 60 mmol.m⁻³ translates into an oxygen increase by 10 mmol-
433 m⁻³ in tropical regions after 50 years of integration.

434

435 The equatorward transport of the anomaly in the subtropics is mostly due to isopycnal subgrid scale
436 mixing processes as shown by the NEMO2 budget analysis. While the models with differing ocean
437 resolutions may differ in their transport of IWM between the subtropical regions and the tropics, it
438 nevertheless suggests that mesoscale activity in higher resolution models is important to spread
439 IWM (e.g Xu et al., 2016). This possibly includes subsurface eddies that show a signature well into
440 the IWM depth range (Frenger et al., 2018, see their Fig 2).



441

442 5.2 IMW transport and Equatorial Intermediate Current System

443 Second, the Equatorial Intermediate Current System (EICS) is not represented in coarse resolution
444 models and only poorly represented in high resolution ocean circulation models (0.25° and 0.1°), as
445 its strength remains too weak by a factor of two (consistent with previous studies, e.g. Ascani et al.,
446 2015). The EICS transports the IWM that occupies the lower thermocline (500 – 1500 m depth) and
447 the recirculation of the IWM in the tropical ocean, as suggested by the observational study of Zenk
448 et al. (2005), and shown in our study.

449

450 We investigated the impact of the EICS on the oxygen supply with tracer release experiments: the
451 concentration of a conservative tracer that originates from the subtropical ocean, is, after 50 years,
452 30 % higher in the eastern equatorial (5°N - 5°S) Pacific in an ocean model with 0.1° resolution,
453 compared to an ocean model with 0.5° resolution. As the oxygen gradient along the equator is
454 similar to the gradient of the conservative tracer, we assume a similar enhancement of oxygen
455 supply by 30 % in the eastern equatorial Pacific at the same time scale. This means, if we account
456 for oxygen consumption due to respiration (about $1 \text{ mmol}\cdot\text{m}^{-3}\cdot\text{yr}^{-1}$ between 5°N - 5°S , see section
457 3.2), that the better resolved EICS in the higher resolution ocean leads roughly to higher
458 intermediate oxygen levels of $15 - 30 \text{ mmol}\cdot\text{m}^{-3}$ compared to the lower resolution ocean experiment
459 in a timescale of 50 years. Consistently, 0.1° -ocean GFDL01 model displays oxygen concentrations
460 larger by about $30 \text{ mmol}\cdot\text{m}^{-3}$ in the eastern equatorial lower thermocline (500-1500 m) compared to
461 the 1° -ocean GFDL1 configuration (with higher subtropical oxygen concentrations of IWM of 15
462 $\text{mmol}\cdot\text{m}^{-3}$ in GFDL01 at 30°S)

463

464 We would like to highlight two potential implications of our finding of the important role of the
465 EICS for the Pacific eastern tropical oxygen supply: i) First, we have shown that the intermediate
466 current system EICS is important for the connection between the western and eastern Pacific Ocean
467 at a decadal / multidecadal time scale. This suggests that the EICS modulates the mean state and the
468 variability of the tropical oxygen in the lower thermocline, and subsequently the whole water
469 column by upwelling of deep waters. ii) Second, we have found an enhancement of the connections
470 between equatorial deep ocean ($> 2000 \text{ m}$) and lower thermocline in the high resolution model
471 compared to the lower resolution model. This result is consistent with the studies of Brandt et al.
472 (2011, 2012), who suggested, based on observational data and on an idealized model, that
473 Equatorial Deep Jets, part of the EICS (see Fig 1b), propagate their energy upward and impact the
474 upper ocean properties of the ocean, including their oxygen content. Taken this into account, we



475 hypothesize that the Pacific Deep Water has a larger role than previously thought in modulating the
476 intermediate and upper ocean properties.

477

478 A pragmatic approach to account for the missing EICS is to increase diffusion anisotropically, with
479 increased zonal mixing in the tropics (Getzlaff and Dietze, 2013). This parameterization mimics a
480 more vigorous EICS and improves the simulated shape of the OMZ in climate models. However,
481 the prominent bias of IMW in climate models, and therefore of the water masses entering the EICS
482 is not accounted for with this parameterization. Furthermore such a parameterization improves the
483 mean state but does not reproduce the variability of the EICS.

484

485 5.3 Implication for biogeochemical cycles

486 The IWM are important supplier of oxygen to the tropical oceans, but also of nutrients (Palter et al.,
487 2010) as well as anthropogenic carbon (e.g Kathiwala et al., 2012), which accumulates in mode and
488 intermediate waters of the Southern Ocean (Sabine et al., 2004; Resplandy et al., 2013). The
489 mechanisms that we discussed here may therefore play a role in ocean carbon climate feedbacks on
490 time scales of decades to a century.

491

492 Finally, this study suggests that changes of the properties of the IWM may contribute to the still
493 partly unexplained deoxygenation of 5 mmol.m^{-3} / decade occurring in the lower thermocline of the
494 equatorial eastern Pacific Ocean (Schmidtko et al., 2017; Oeschlies et al., 2018). In addition to an
495 oxygen decrease in tropical regions, Schmidtko et al. (2017) showed a decrease of oxygen levels by
496 $2\text{-}5 \text{ mmol.m}^{-3}$ in the regions of formations of AAIW. Based on repeated cruise observations,
497 Panassa et al. (2018) highlighted an increase of the apparent oxygen utilization in the core of the
498 AAIW, related to a 5 % increase in nutrient concentrations from 1990 to 2014. The transport of this
499 modified AAIW, poorer in oxygen and richer in nutrients, toward the low latitudes both by small
500 scale processes (section 3) and at the equator by the EICS (section 4), may explain a significant part
501 of the occurring deoxygenation in the equatorial ocean. In complement to changes in the AAIW
502 properties, little is known about the variability and long term trend of the strength of the EICS, an
503 oceanic “bridge” between the western and the eastern part of the basin. A possible way forward
504 could be to perform idealized model experiments in high resolution configurations, aiming to assess
505 both the effect of the observed change in the AAIW properties and of a potential change of EICS
506 strength on oxygen levels.

507

508



509 **Data and code availability**

510 The code for the Nucleus for European Modeling of the Ocean (NEMO) is available at:
511 <https://www.nemo-ocean.eu/>. The code for the University of Victoria (UVIC) model is available
512 at [:http://terra.seos.uvic.ca/model/](http://terra.seos.uvic.ca/model/). The Lagrangian particles ARIANE code is available at
513 <http://stockage.univ-brest.fr/~grima/Ariane/>. The Coordinated Ocean-ice Reference Experiments
514 (COREv2) dataset is available at: <https://data1.gfdl.noaa.gov/nomads/forms/core/COREv2.html>.
515 The experiments data is available on request.

516

517 **Authors contributions**

518 OD conceived the study, performed the NEMO model and ARIANE experiments and analyzed the
519 data. IF preprocessed and helped to analyze the GFDL data. JG preprocessed and helped to analyze
520 the UVIC data. All authors discussed the results and wrote the manuscript.

521

522 **Competing interest**

523 The authors declare that they have no conflict of interest.

524

525 **Acknowledgments**

526 This work is a contribution of the SFB754 “Climate-Biogeochemistry Interactions in the Tropical
527 Ocean”, supported by the Deutsche Forschungsgemeinschaft (DFG). The NEMO simulations were
528 performed at the North German Supercomputing Alliance (HLRN). We would like to thank Markus
529 Scheinert (research unit “Ocean Dynamics”, GEOMAR) for his technical support in compiling the
530 NEMO code and for providing the high resolution NEMO input files. We would like to thank
531 GFDL for producing the CM2-0 suite that involved a substantial commitment of computational
532 resources and data storage.

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743 **Table and Figures**

744

745 Table 1 :

Model	Resolution	Atmosphere	BGC	Model Reference (circulation)	Model Reference (BGC)
Mean state comparison					
UVIC	2.8°	Coupled (temperature, humidity) Forced (NCEP/NCAR wind stress)	UVIC-BGC	Weaver et al., 2001	Keller et al., 2012
NEMO2	2° (0.5 eq)	Forced COREv2 “normal year”	NPZD-O2	Madec et al., 2017	Kriest et al., 2010 Duteil et al., 2014
GFDL1	1°	Coupled (~50 km)	miniBLING	Delworth et al, 2012, Griffies et al, 2015	Galbraith et al., 2015, Dufour et al, 2015
GFDL025	0.25 °	Coupled (~50 km)	miniBLING		
GFDL01	0.1°	Coupled (~50 km)	miniBLING		
Sensitivity experiments					
Model	Resolution	Atmosphere	BGC	Experiments	
NEMO2 (section 2.2.1)	2° (0.5 eq)	Forced COREv2 1948-2007	NPZD-O2	REF: control experiment 30N30S: O2 restoring to WOA at 30°N/30°S 30N30S1500M: O2 restoring to WOA at 30°N/30°S/1500m	
NEMO05 (section 2.2.2)	0.5°	Forced COREv2 1948 - 2007	Tracer release	Tracer initialized to 1 (O2 WOA > 150 mmol.m ⁻³) or 0 (O2 WOA < 150 mmol-m-3)	
NEMO01 (section 2.2.2)	0.1°	Forced COREv2 1948 – 2007	Tracer release		

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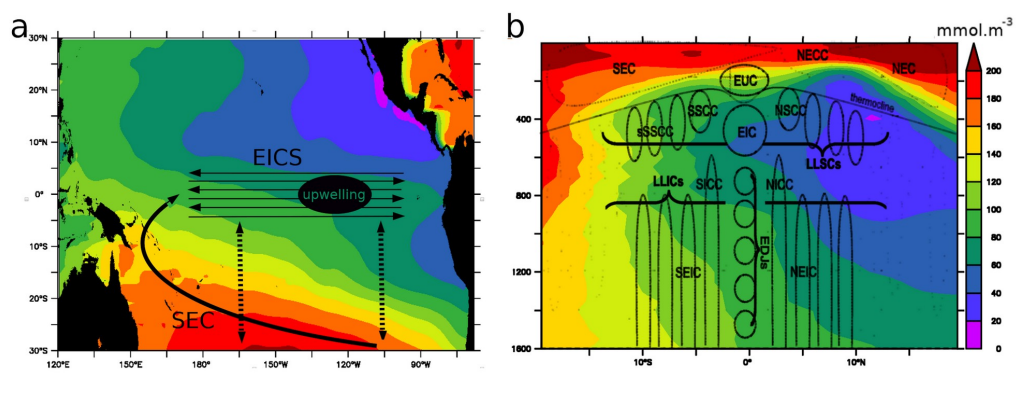
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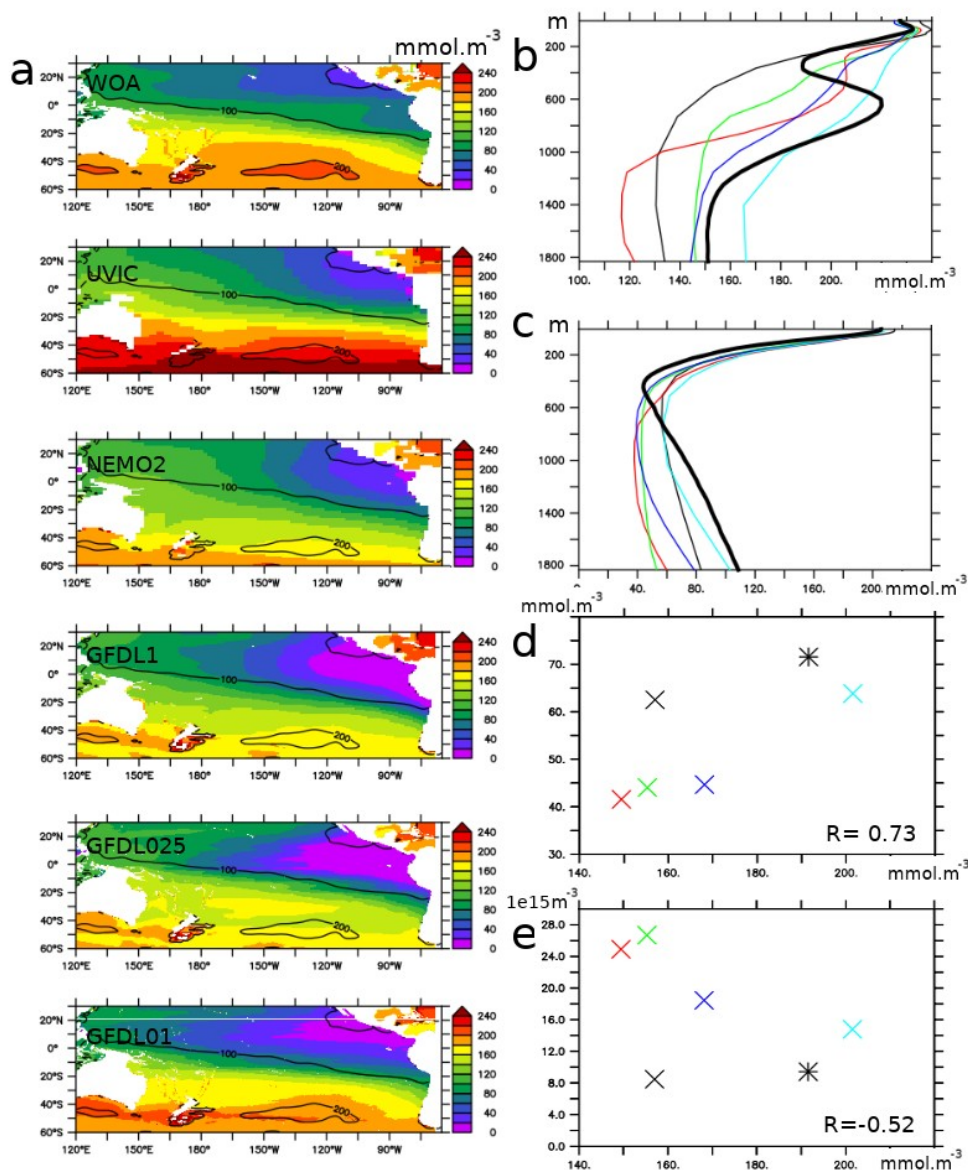


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757 Figure 1 : a- schema summarizing the intermediate water masses (IWM) pathway from the
758 subtropics into the equatorial regions. EICS : Equatorial Intermediate Current System. SEC : South
759 Equatorial Current. Dashed line : isopycnal diffusive processes. Oxygen levels (mmol.m^{-3}) in the
760 lower thermocline (mean 500-1500m) are represented in color. b - schema (adapted from
761 Menesguen et al., 2019) illustrating the complexity of the EICS, extending below the thermocline
762 till more than 2000 m depth (see section 4.1 for a detailed description). Oxygen levels (mean 500 –
763 1500m) at 160°W are represented in color (mmol.m^{-3}).

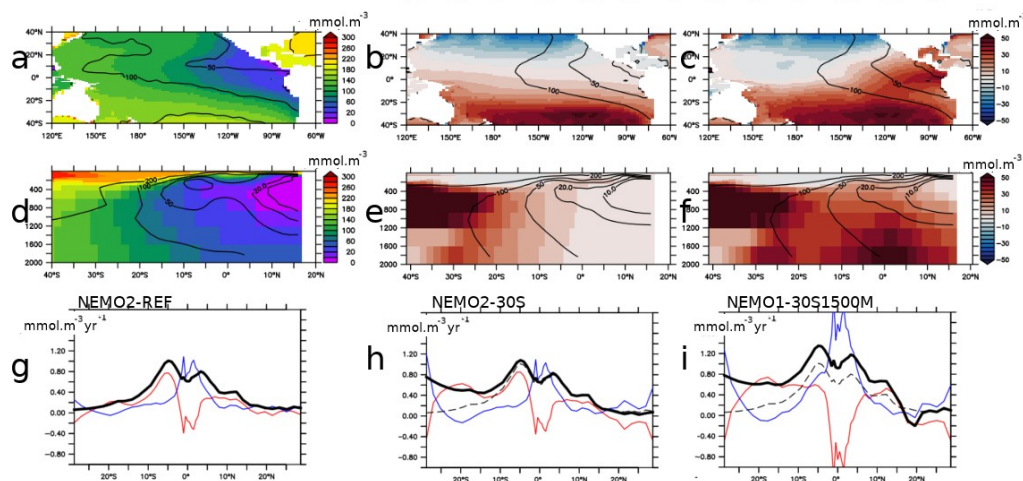


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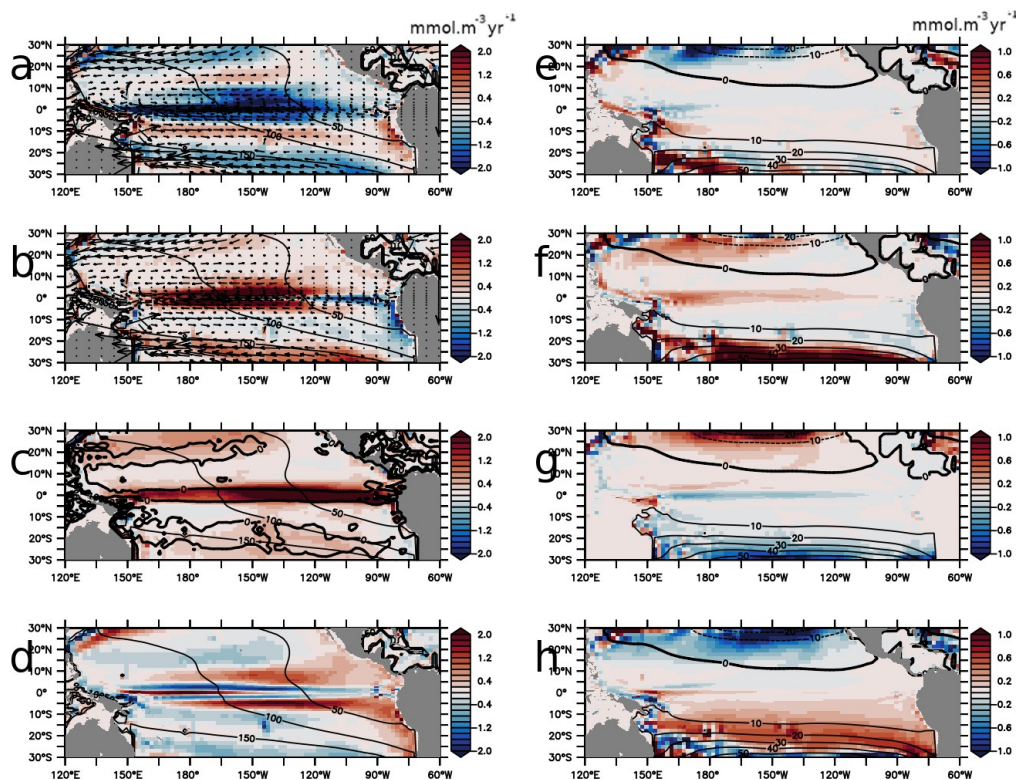
765 Figure 2 : a- oxygen levels (mmol.m^{-3}) in observations (World Ocean Atlas - WOA) (mean 500 –
 766 1500 m) and models (UVIC, NEMO2, GFDL1, GFDL025, GFDL01). Contours correspond to
 767 WOA values. b: average “30°S” (120°E-65°W, 30°S) c : average “tropics” (160°W-coast, 20°N-
 768 20°S). d: average “30°S” vs “tropics”. e: average “30°S” vs volume of tropical suboxic ocean
 769 (oxygen lower than 20 mmol.m^{-3}) regions ($1\text{e}15\text{m}^3$). UVIC : black, NEMO2 : cyan, GFDL1 : red,
 770 GFDL025, green; GFDL01 : blue, WOA: star.



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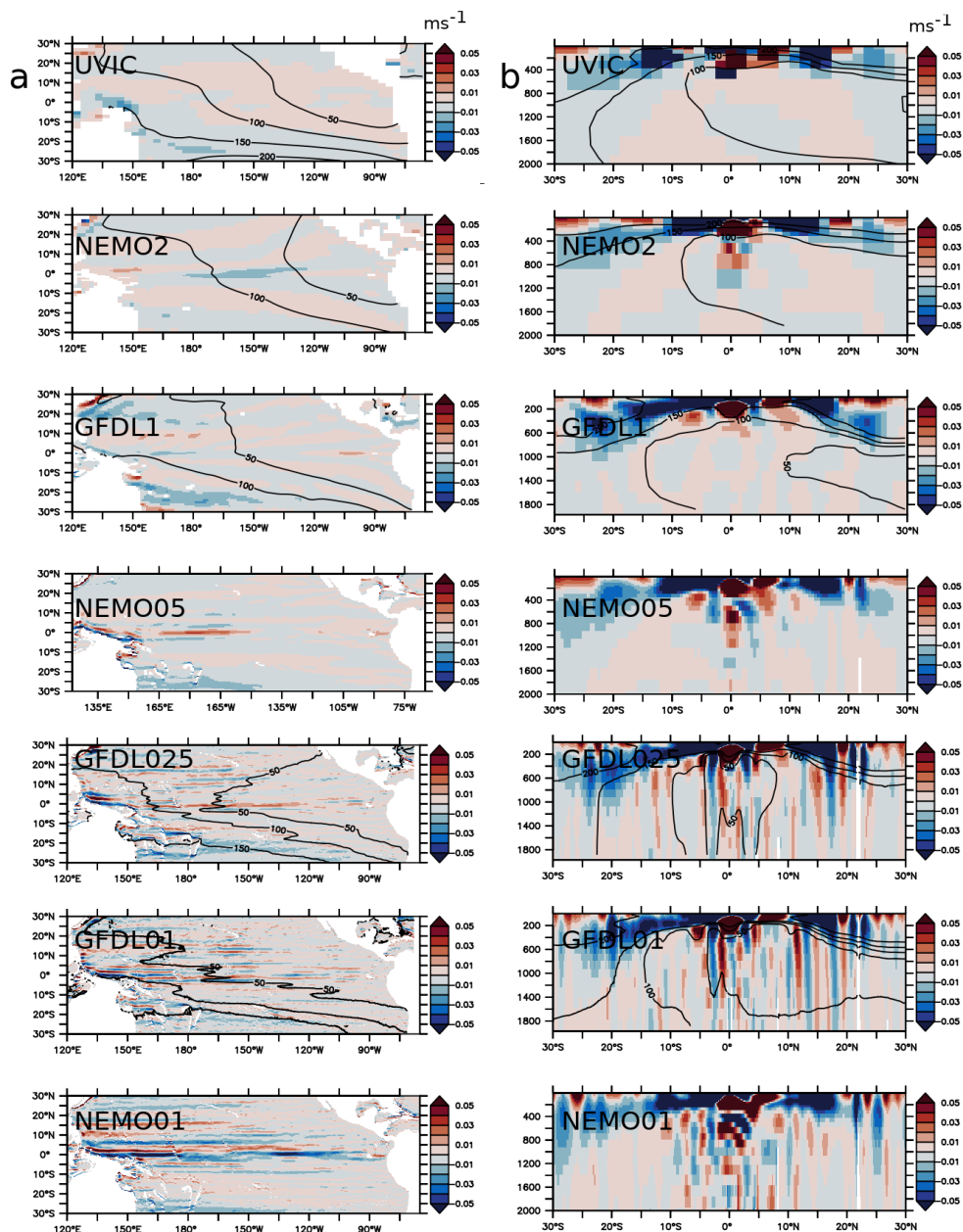


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775 Figure 3 : a- Oxygen (mmol.m^{-3}) in the experiments NEMO2-REF (color) and World Ocean Atlas
776 (contour) (average 500-1500 m). b- Oxygen (mmol.m^{-3}) difference (average 500 - 1500 m) between
777 the experiments NEMO2-30S minus NEMO2-REF and c- NEMO2-30S1500M minus NEMO2-
778 REF (contour NEMO2-REF). d- Oxygen (mmol.m^{-3}) in the experiments NEMO2-REF (color) and
779 World Ocean Atlas (contour) (100°W). e- Oxygen (mmol.m^{-3}) difference (100°W) between the
780 experiments NEMO2-30S minus NEMO2-REF and f- NEMO2-30S1500M minus NEMO2-REF
781 (contour NEMO2-REF). g-i : basin zonal average (average 500 - 1500 m) of the oxygen total
782 supply (bold) ($\text{mmol.m}^{-3}.\text{year}^{-1}$), advective processes (blue) and isopycnal diffusion (red) in g -
783 NEMO2-REF, h- NEMO2-30DEG, i-NEMO2-30DEG1500M. The dashed line in is the oxygen
784 total supply in NEMO2-REF.
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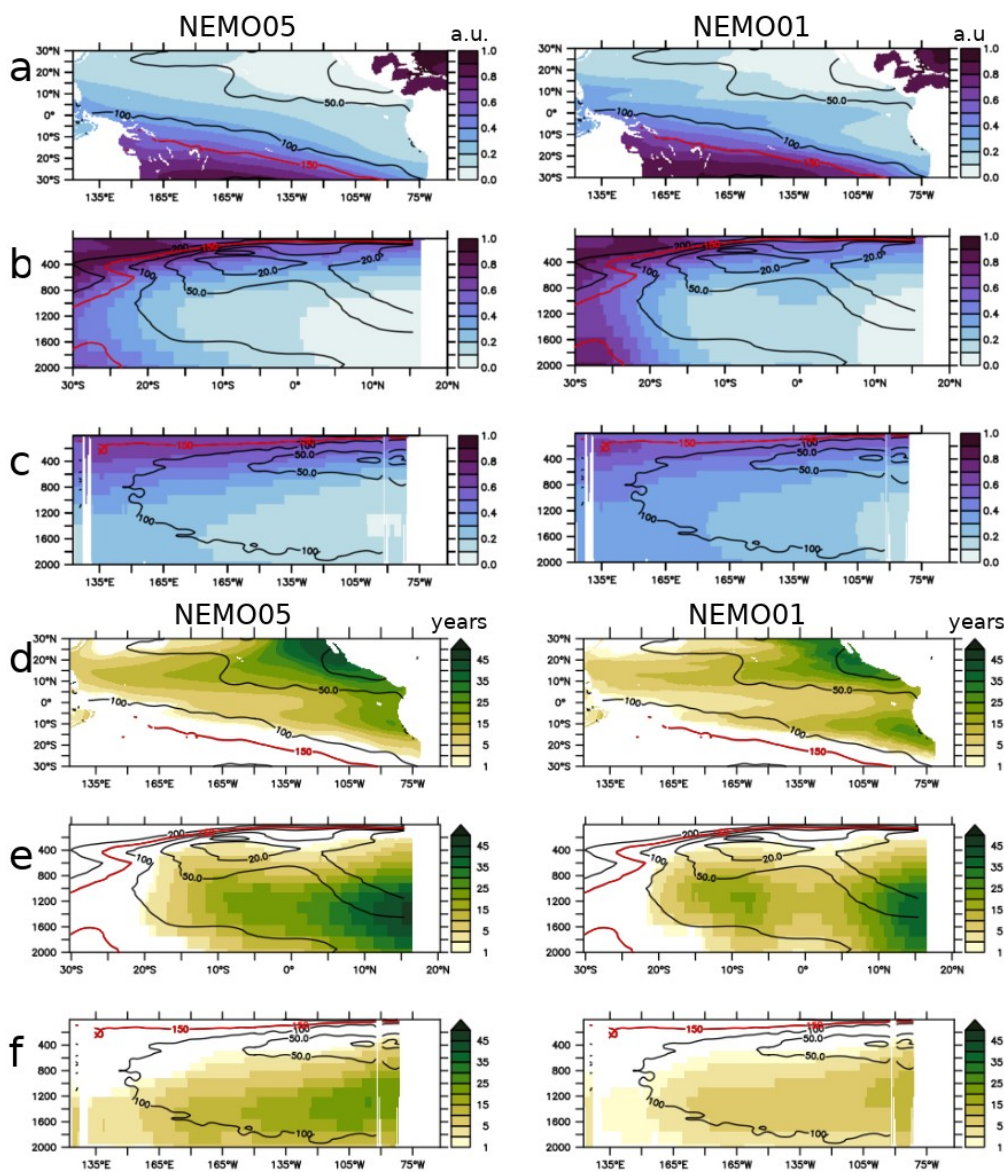
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787 Figure 4 : Oxygen supply processes ($\text{mmol.m}^{-3}.\text{year}^{-1}$ – average 500 - 1500m) in NEMO2-REF : a-
788 zonal advection, b– meridional advection, c- vertical advection, d- isopycnal diffusion. The
789 meridional and zonal currents are displayed as vectors in a,b and the vertical current as contour in c.
790 Oxygen levels (mmol.m^{-3}) are displayed in contour. Difference in oxygen supply processes
791 ($\text{mmol.m}^{-3}.\text{year}^{-1}$ – average 500-1500m) between NEMO2-30DEG and NEMO2-REF : e- zonal
792 advection, f – meridional advection, g- vertical advection, h- isopycnal diffusion. The NEMO2-
793 30DEG – NEMO2-REF oxygen anomaly (mmol.m^{-3}) is displayed in contour.



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795 Figure 5 : mean currents velocity (ms^{-1}) at a- 1000 m depth b- 100°W in UVIC, NEMO2,
796 NEMO05, GFDL025, GFDL01, NEMO01. The mean oxygen levels (mmol.m^{-3}) (when coupled
797 circulation-biogeochemical experiments have been performed – see Table 1) are displayed in
798 contour

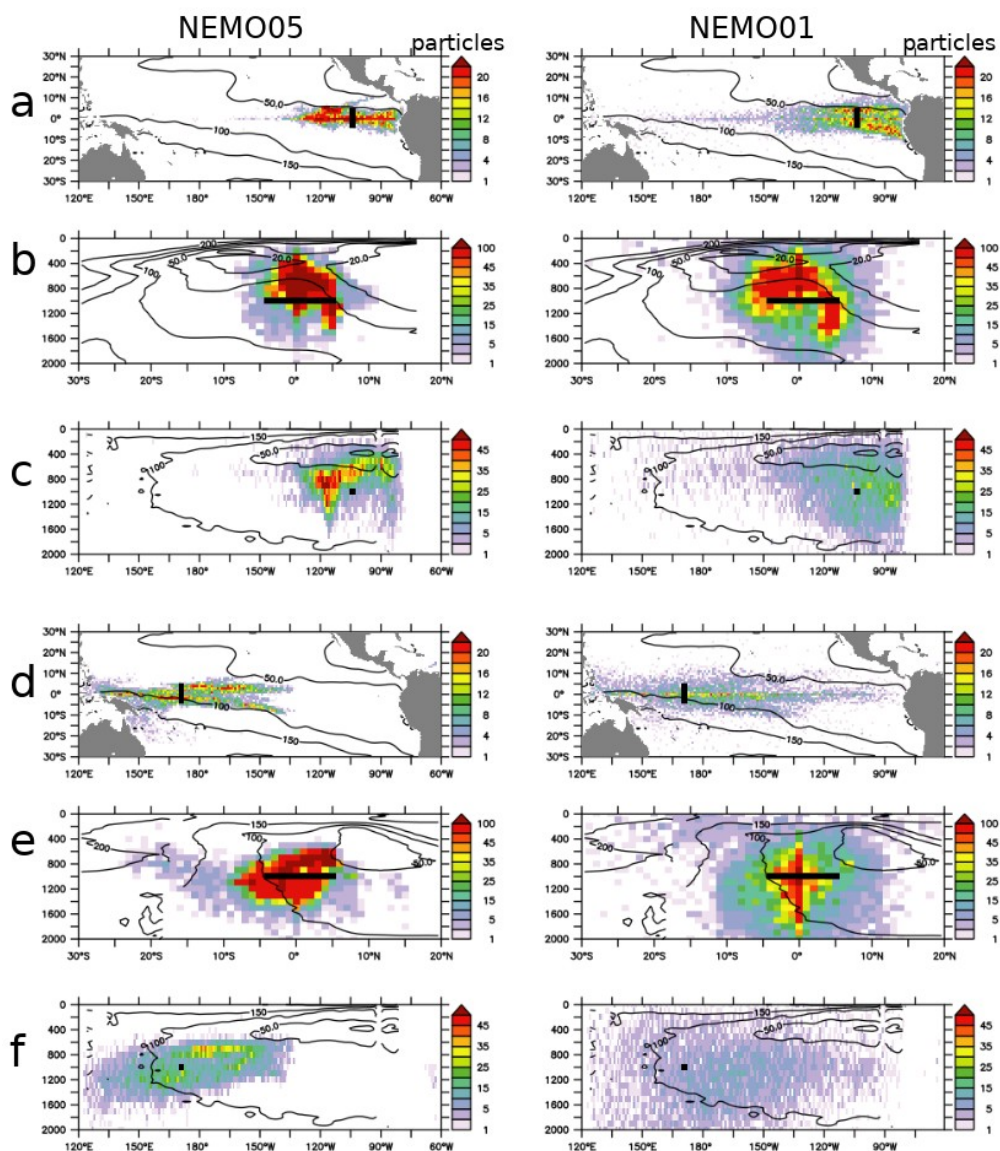


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800 Figure 6: a-c : tracer concentration (arbitrary unit) after 60 years integration in NEMO05 (left) and
801 NEMO01 (right). a- average 500-1500m, b- section 100°W, c- equatorial section. d-f: Time (years)
802 at which the released tracer reaches the concentration 0.1 ($t_{10\%}$) in NEMO05 and NEMO01. d-
803 average 500-1500m, e- section 100°W, f- equatorial section. In all the subpanels, the WOA oxygen
804 levels are displayed in contour. The red contour is the WOA 150 mmol.m^{-3} oxygen isoline, used to
805 initialize the tracer level.



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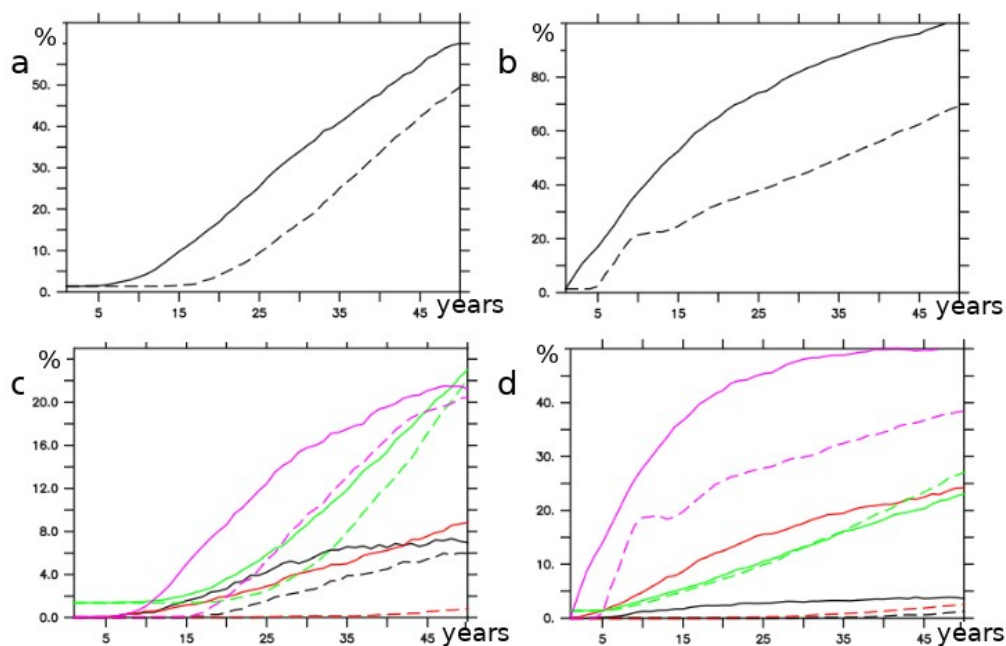


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808 Figure 7 : Density (number of particles in a $1^{\circ} \times 1^{\circ} \times 100\text{m}$ depth box) distribution of the location of
809 released Lagrangian particles (15 years backward integration starting from the final experiment
810 state) in NEMO05 (left) and NEMO01 (right). The release location is identified in bold and is
811 located at $100^{\circ}\text{W}/5^{\circ}\text{N}$ - $5\text{S}/1000$ m depth. a- vertical integrated density; b- zonal integrated density;
812 c- meridional integrated density. d-f : Similar to a-c but with a release location located at
813 $160^{\circ}\text{E}/5^{\circ}\text{N}$ - $5^{\circ}\text{S}/1000$ m depth. The mean oxygen levels are displayed in contour.



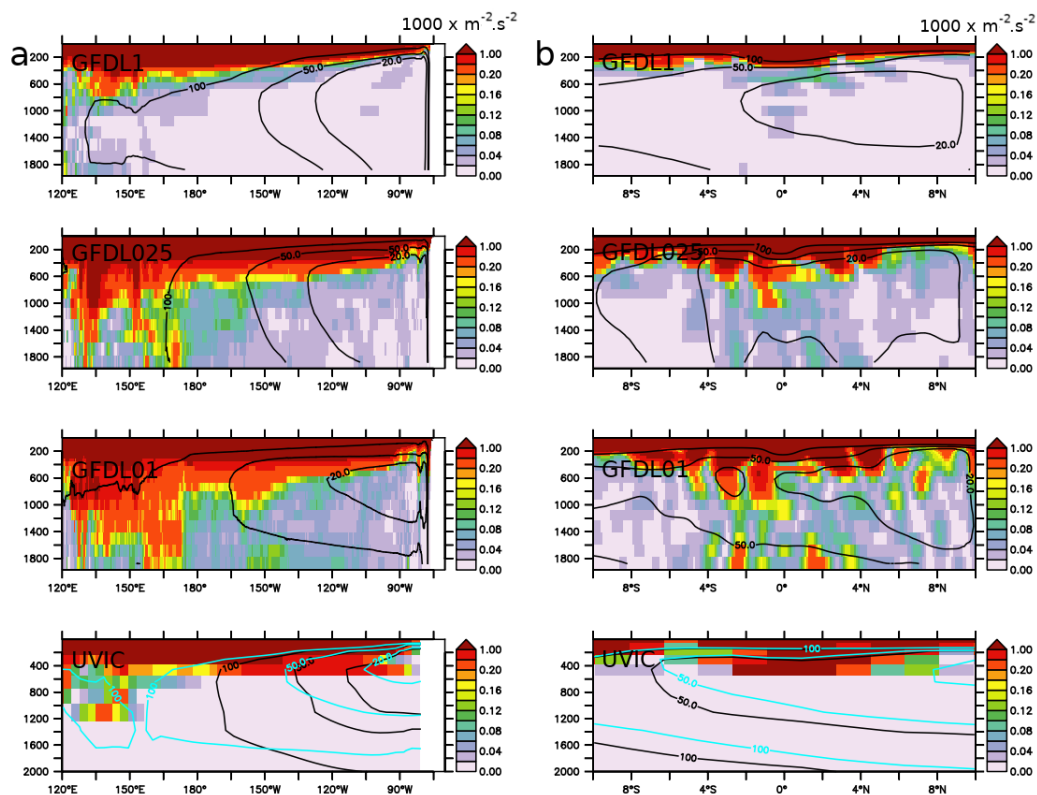
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820 Figure 8 : a : percentage of particles originating from outside the Intermediate Eastern Tropical
821 Pacific (IETP) ocean region (release 100°W / 5°N-5°S / 1000 m) or b- originating from outside the
822 Intermediate Western Tropical Pacific (IWTP) ocean region (release 160°E / 5N°5S / 1000 m) in
823 NEMO01 (black) and NEMO05 (dash). c,d ; percentage of particles originating from outside the
824 IETP (c) and the IWTP (d): upper ocean (< 200 m) (black), deep ocean (> 2000 m) (red),
825 subtropical region (> 10°N/S) (green), panel c: western (west of 160°W) - or panel d: eastern (east
826 of 160°W) part of the basin (magenta).

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837 Figure 9 : a - Mean Kinetic Energy ($\text{m}^2 \cdot \text{s}^{-2} \times 1000$) (average 10°N-10°S) in GFDL01, GFDL025,
838 GFDL01, UVIC, b - similar to a. but average 160°W-coast. Oxygen levels ($\text{mmol} \cdot \text{m}^{-3}$) are displayed
839 in contour. The blue contour corresponds to UVIC GD13 (Getzlaff and Dietze, 2013, including an
840 anisotropic increase of lateral diffusion at the equator)