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# The influence of merger and convection on an anticyclonic eddy trapped in a bowl

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## Abstract :

We investigate the impact of several parameters on the lifecycle of an anticyclonic eddy lying in a topographic depression (a bowl), similar to the Lofoten Vortex and the Rockall Trough eddy cases. We observe that the vortex merger with submesoscale coherent vortices generated at depth allows the eddy to grow in size, and intensify at depth. Wintertime convection is also shown to directly intensify the eddy by deepening isopycnals. Also, convection indirectly affects the shape of the eddy. It enhances the number of merger (1) at the surface, with small vortices generated in the convectively-deepened surface mixed-layer, and (2) at depth, because the vertical distance between the main eddy's core and small companion vortices is reduced, thus increasing the merging efficiency. These processes altogether contribute to the maintaining of the eddy. On the other hand, the bottom drag is the main process contributing to the decay of the eddy. Our study thus shows that the sustaining for several years of such eddies trapped in a bowl is mainly due to the balance between merger and bottom drag.

## Highlights

► Anticyclonic eddies lying in a topographic depression live for several years. ► Small SCVs generated nearby the anticyclonic eddy intensify the latter at their depth of generation. ► Wintertime convection is efficient in intensifying the anticyclonic eddy. ► The sustaining for several years of such eddies is due to the balance between merger, convection, and bottom drag.

**Keywords** : Mesoscale, Vortex, Merger, Convection

## 1 1. Introduction

2 Mesoscale eddies are a prominent feature of the ocean circulation. They have a strong  
3 influence on biological activity (Chelton et al., 2011), tracer transport (Zhang et al., 2014),  
4 and physical and chemical properties of the water column (Dong et al., 2014). In some  
5 regions, semi-permanent eddies can be seen throughout the year, at a nearly constant po-  
6 sition. Among other examples, two particular cases are the Lofoten Vortex (LV), and the  
7 Rockall Trough eddy (RT eddy). These two semi-permanent eddies have the peculiarity to  
8 be anticyclonic, and located above a topographic depression – a bowl. The formation of  
9 such vortices has recently been examined by Solodoch et al. (2021). Authors showed using  
10 idealized simulations that successive merging events form a permanent anticyclone lying in  
11 the topographic depression. The dynamics of the resulting vortex depends on the ratio of  
12 eddy’s vorticity to topography’s potential vorticity. However, the mechanisms that sustain  
13 semi-permanent anticyclones in bowl-like topography such as the LV and the RT eddy are  
14 not yet fully understood.

15 The LV can be found in the Lofoten Basin in the Nordic Seas. It appears as a large  
16 anticyclone at the center of the basin. It was first detected by *in situ* data between 1970 and  
17 1990 (Ivanov & Korablev, 1995). The LV is intensified between 700 and 900 m depth and has  
18 a radius of about 30 km (Yu et al., 2017). Two processes are candidate to explain the long  
19 lifetime of the LV. First, from observational data, Ivanov & Korablev (1995) and Bosse et al.  
20 (2019) argued that wintertime intensification resulting from convection plays a determinant  
21 role in sustaining the LV. Second, model studies showed that the LV is sustained by the  
22 merger and alignment with smaller vortices generated by unstable boundary currents (Köhl,  
23 2007; Trodahl et al., 2020). In the current state of knowledge, the relative importance of  
24 each process is not clear. One of the aim of the present study is to give new answers to this  
25 question.

26 The RT eddy is located in the Rockall Trough, off Ireland in the North Atlantic. It has  
27 a clear signature at the sea surface (Heywood et al., 1994; White & Heywood, 1995; Volkov,  
28 2005; Xu et al., 2015), but also at depth with high values of eddy available potential energy  
29 (Roullet et al., 2014). This eddy is less sampled than the LV and less known. However,  
30 thanks to recent *in situ* deployments, it has been shown that it is intensified at depth, with

31 a maximum azimuthal velocity of  $\sim 0.3 \text{ m s}^{-1}$  near 500 m depth (Smilenova et al., 2020). It  
32 has a radius of approximately 40 km and can reach down to 1500 m. Its lifecycle, as well as  
33 the mechanisms that sustain it are yet poorly documented. However, recent model studies  
34 by Le Corre et al. (2019) and Smilenova et al. (2020) have shown some evidences that 1)  
35 the RT eddy formation is the result of successive mergers of deeply generated submesoscale  
36 vortices along the Porcupine Bank, 2) the merger of the RT eddy with these small vortices as  
37 well as wintertime convection sustain the RT eddy, and allow it to remain semi-permanent  
38 in the Rockall Trough.

39 In this paper, we investigate the impact of several parameters on the lifecycle of an  
40 anticyclonic eddy lying in a topographic depression. In particular, we discuss the impact of  
41 merger and convection on the lifetime and shape of the anticyclone. To explore the parameter  
42 space, we use an idealized approach based on the Rockall Trough Eddy case. This allows to  
43 1) discuss on the general behavior of anticyclonic eddies in a bowl, and 2) give insights in  
44 the particular case of the Rockall Trough Eddy that is yet poorly documented. In section  
45 2 we present the methods, the numerical simulation setup and the diagnostics performed  
46 on outputs. In section 3 we present the results of our study, the impact of the different  
47 parameters on the vortex dynamics. In section 4 we summarize and discuss the results.

## 48 **2. Methods**

### 49 *2.1. The numerical simulations*

50 In this section, we present the idealized simulations performed for this study. The aim of  
51 these simulations is to simulate schematically the dynamics occurring in the Rockall Trough  
52 area: a semi-permanent anticyclone (the RT eddy) lying in a bowl-like topography, fed by  
53 anticyclonic Submesoscale Coherent Vortices (SCVs) generated hundreds of kilometers away  
54 from the main eddy (hereafter, the main eddy designates the eddy that lies approximately  
55 in the center of the bowl-like topography, and merges with smaller SCVs). We detail each  
56 aspect of the simulation in the following subsections.

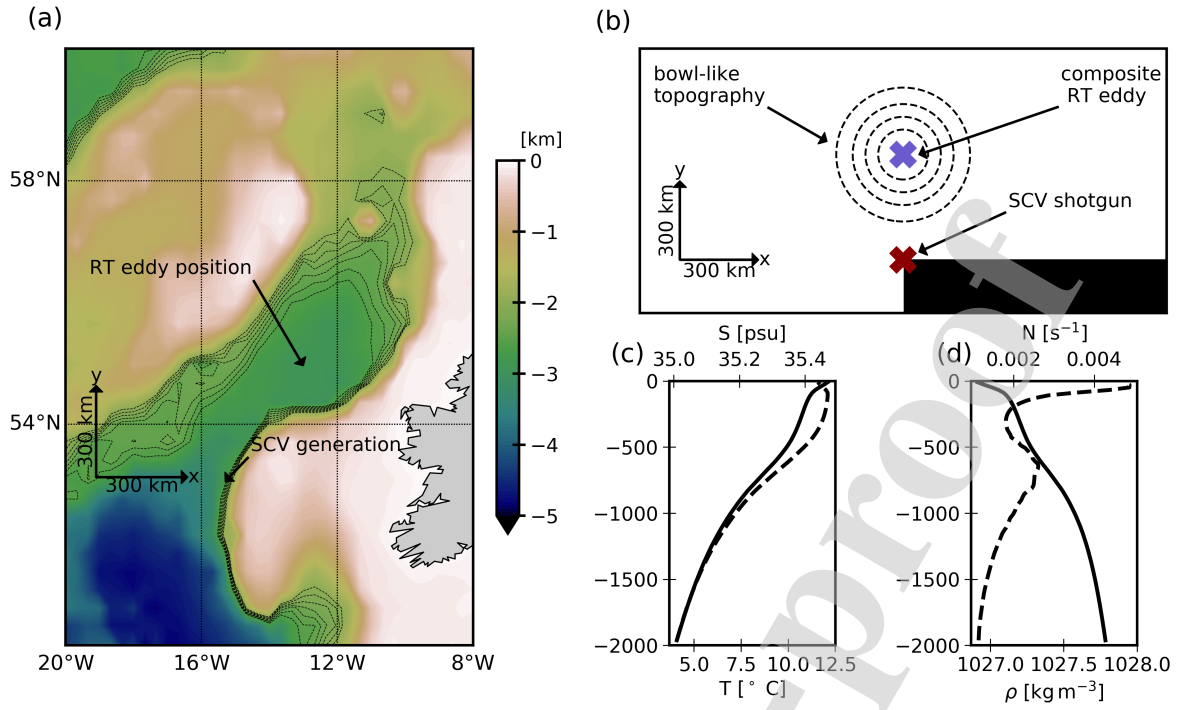


Figure 1: a) Rockall Trough bathymetry; dashed contours show isobaths 2500 to 2000 m depth with a 100 m interval. b) Scheme of idealized simulation setup; dashed contours show same isobaths as in a). c) Climatological background temperature (solid) and salinity (dashed) used in idealized simulations. d) Climatological background potential density (solid) and corresponding Brunt-Väisälä frequency (dashed) used in idealized simulations.

### 57 2.1.1. Numerical setup and domain

58 The simulations rely on a 3D primitive equation framework. They are performed using  
 59 the Coastal and Regional Ocean Community model CROCO (Shchepetkin & McWilliams,  
 60 2005). This model solves the hydrostatic primitive equations for the velocity, temperature,  
 61 and salinity, using a full equation of state for seawater (Shchepetkin & McWilliams, 2011).  
 62 The simulations integrate the primitive equations for about 7 and a half years. The nu-  
 63 merical settings are similar to previous simulations performed in an idealized context (see,  
 64 *e.g.*, Ménesguen et al., 2018): horizontal advection terms for tracers and momentum are dis-  
 65 cretized with fifth-order upwind advection schemes (UP5); the explicit horizontal viscosity  
 66 and diffusivity are set to zero, since the UP5 scheme damps dispersive errors; the vertical  
 67 advection is discretized with a fourth-order centered parabolic spline reconstruction (Splines  
 68 scheme). Further discussion about these parameterizations can be found in Klein et al.

69 (2008) or Ménesguen et al. (2018). Vertical mixing of tracers and momentum is done using  
 70 a K-profile parametrization (KPP, Large et al., 1994), and the effect of bottom friction is  
 71 parameterized through a logarithmic law of the wall (with the same parameters than in *e.g.*  
 72 Gula et al. (2015) or Le Corre et al. (2020)). Some simulations are run without this bottom  
 73 drag to study its impact on the vortex dynamics. Simulations have 64 terrain-following ver-  
 74 tical levels, which are stretched such that the resolution increases in the depth range where  
 75 the main eddy lies, giving  $\Delta z \sim 20$  m from surface to 1000 m depth, and  $20 < \Delta z < 90$  m  
 76 below. The horizontal resolution is  $\Delta x = 5$  km.

77 The domain is chosen so that it represents schematically the RT area, see Fig. 1(a,b).  
 78 The domain is 2000 km and 1000 km wide zonally and meridionally, respectively. A bowl-  
 79 like topography is placed at the center of the domain, to represent the RT topographic  
 80 depression. It is modeled by a Gaussian function

$$h = h_0 + h_1 \exp(-r^2/(2R)^2),$$

81 with  $r = \sqrt{(x - x_0)^2 + (y - y_0)^2}$ ,  $x_0 = 1000$  km,  $y_0 = 600$  km,  $R = 100$  km,  $h_0 = 2000$  m,  
 82 and  $h_1 = 500$  m, such that the simulation is 2000 m deep everywhere, except in the bowl  
 83 where it reaches 2500 m deep. The background stratification is the average stratification  
 84 in the RT area, see Fig. 1(c,d). It is defined as the average stratification in the RT from  
 85 Le Corre et al. (2020)'s simulation. A return to this background stratification is set in the  
 86 boundaries. At these boundaries a 10 km wide sponge layer avoid the generation of spurious  
 87 boundary dynamics.

### 88 2.1.2. The SCV shotgun

89 In the RT, SCVs are generated along the Porcupine Bank (Smilenova et al., 2020). To  
 90 simulate this SCV generation, we designed a "SCV shotgun", that continuously generates  
 91 SCVs at a given depth during the simulation. It is placed at 300 km from the bowl-like  
 92 topography center. This distance is chosen so that it is similar to the one between the  
 93 Porcupine Bank and the RT eddy position in reality, see Fig. 1(a,b).

94 The SCV shotgun is based on the principle fully described in Deremble et al. (2016): at  
 95 boundary singularities such as corners, vorticity is injected into the domain even for free-slip

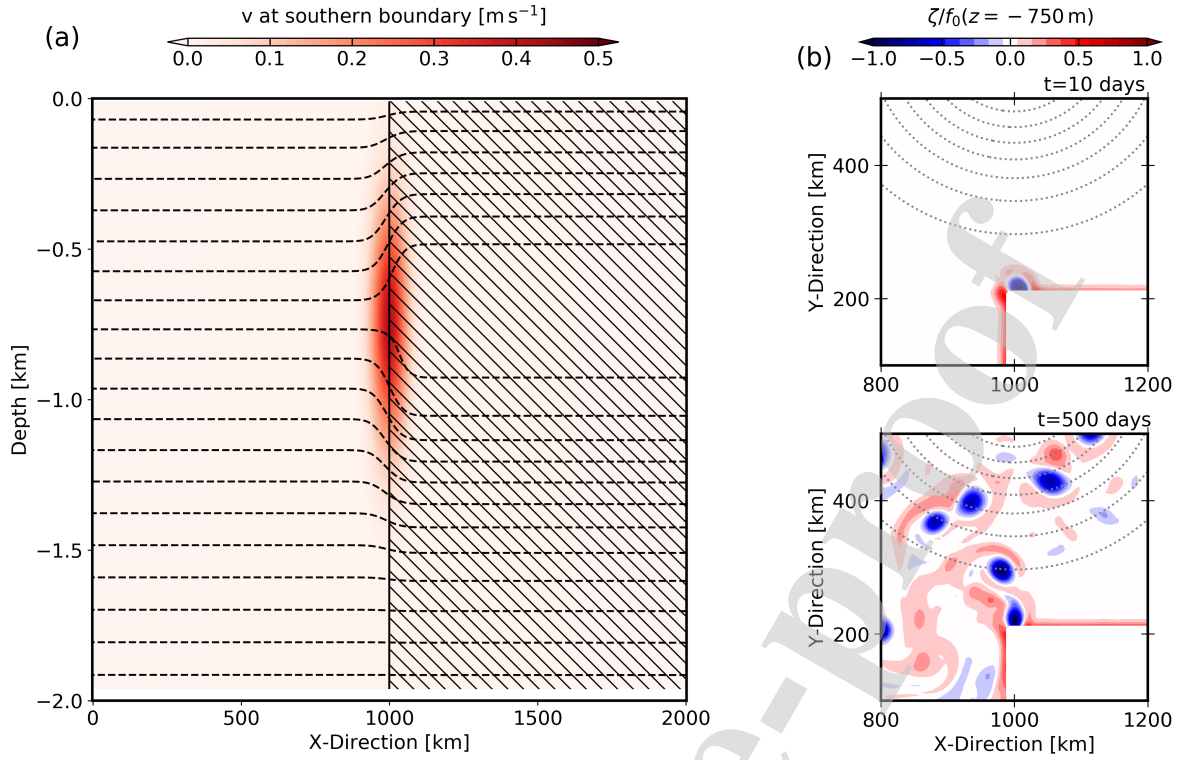


Figure 2: a) Meridional velocity at the southern boundary for the "middle" case; black lines indicate isopycnals with a  $0.5 \text{ kg m}^{-3}$  spacing; hatched area indicate the position of the mask. b) Snapshots of normalized relative vorticity at  $t=10$  and  $t=500$  days, at 750 m depth, showing the SCV generation at the mask corner in the "middle" case.

96 boundary conditions. We add a land mask forming a corner (*i.e.* a boundary singularity)  
 97 at the south of the domain, with a free-slip condition along this mask. Then, we impose a  
 98 meridional current at depth along the mask (see Fig. 2(a)), of the form:

$$v = v_0 \exp(-(x - x_0)^2 / (2L)^2) \exp(-(z - z_0)^2 / (2H)^2),$$

99 with  $v_0 = 0.4 \text{ m s}^{-1}$ ,  $L = 30 \text{ km}$ , and  $H = 200 \text{ m}$ . As discussed in Deremble et al. (2016),  
 100 the horizontal extension and intensity of generated SCVs are mainly controlled by the sub-  
 101 grid parameterization and horizontal discretization, such that  $L$  and  $v_0$  poorly control the  
 102 shape of SCVs. After sensitivity tests, we chose the aforementioned values for  $v_0$ ,  $L$ , and  
 103  $H$  such that the model stability is satisfying, and that the properties and the frequency of  
 104 generation of SCVs are similar to the one observed in realistic simulations of the Rockall

105 Trough (Smilenova et al., 2020), *i.e.* about 10 SCVs are generated each year. We also vary  
 106  $z_0 = [-1250, -1000, -750, -500, -250]$  m, to discuss the impact of the SCV depth on the merging  
 107 process. They are called "deep", "middle deep", "middle", "middle surf", and "surf" cases  
 108 respectively in the following. Note that the middle case is the one representative of RT  
 109 SCV generation (see *e.g.* Fig. 11 in Smilenova et al. (2020)). This current is geostrophically  
 110 adjusted with the density field at the southern boundary, see Fig. 2(a). Examples of SCV  
 111 generation in the middle case are shown in Fig. 2(b,c).

### 112 2.1.3. The Rockall Trough anticyclone

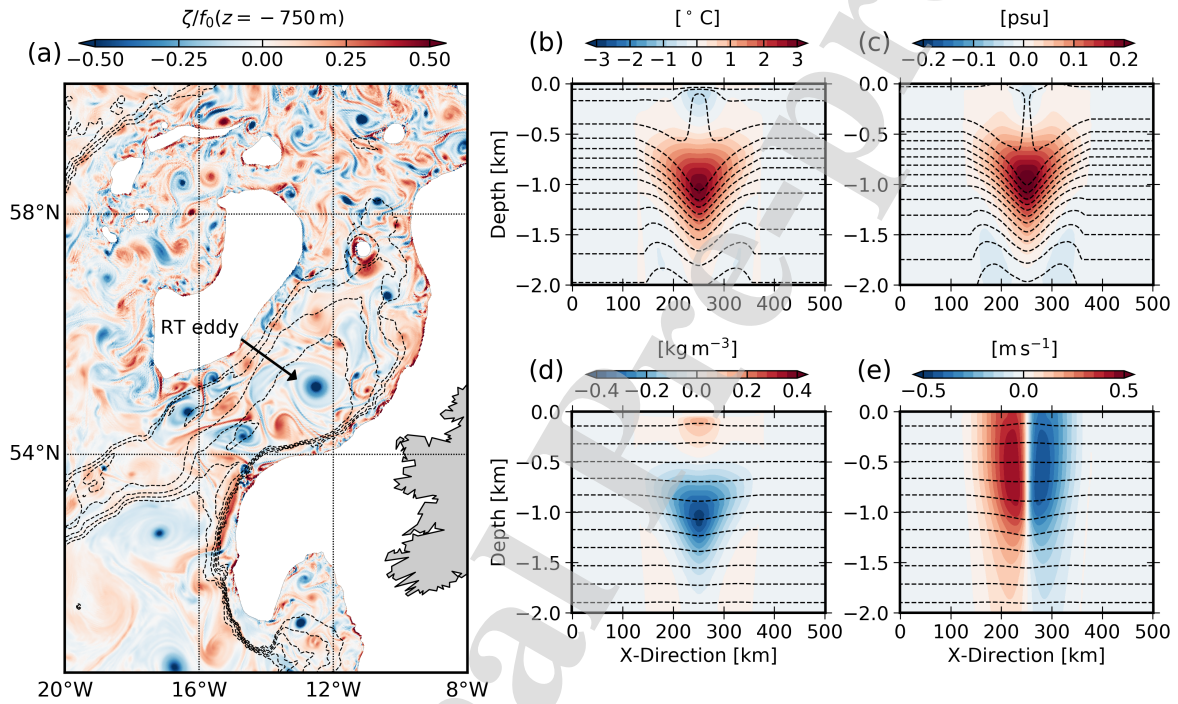


Figure 3: a) Snapshot of normalized relative vorticity at 750 m depth in the Rockall Trough area, from the realistic simulation (Le Corre et al., 2020) in which the composite anticyclone was extracted; dashed contours show isobaths from 3500 to 2000 m depth with a 250 m interval. (b,c,d,e) Temperature anomaly, salinity anomaly, density anomaly, and azimuthal velocity of the composite anticyclone; dashed contours show isolines of temperature (b), salinity (c), and density (d,e).

113 As discussed in the introduction, a semi-permanent anticyclonic eddy is present in the  
 114 RT throughout the whole year: the RT eddy. To simulate this presence, we add in some  
 115 simulations, at initialization, a composite anticyclone representative of the RT eddy above



116 the center of the bowl-like topography (at  $x = x_0$  and  $y = y_0$ , the blue cross position in Fig.  
117 1(b)).

118 This composite was extracted from a realistic simulation representing the Subpolar North  
119 Atlantic gyre. It is fully described in Le Corre et al. (2020). The 2011 vertical properties of  
120 the simulated RT eddy being close to ship-board Conductivity-Temperature-Depth (CTD)  
121 data collected in January 2011 (see the supplementary material of Smilenova et al. (2020)),  
122 this simulation is assumed to represent well the RT eddy dynamics. An example of RT eddy  
123 occurrence in the simulation is shown in Fig. 3(a). We tracked the RT eddy by following  
124 the maximum SSH value in the area. After isolating the eddy, we took its temporal main  
125 structure and azimuthally averaged it to obtain the main composite structure of the RT  
126 eddy (Fig. 3(b,c,d,e)).

#### 127 2.1.4. *Add convection*

128 The wintertime convection may play an important role in the intensification and the  
129 maintenance of anticyclonic eddies (Gelderloos et al., 2011; Bosse et al., 2016, 2019), in  
130 particular in high latitude areas such as the RT. To discuss the impact of the convection  
131 on the intensity of the RT eddy, we ran simulations with a surface net heat flux (SNHF)  
132 representative of the RT area. It is calculated as an average of the SNHF (from Carton  
133 et al., 2018) in the area of latitude and longitude comprised respectively between  $53^\circ\text{N}$  and  
134  $57^\circ\text{N}$ , and  $15^\circ\text{W}$  and  $11^\circ\text{W}$ . The annual variation of SNHF imposed in the simulations with  
135 convection is shown in Fig. 4.

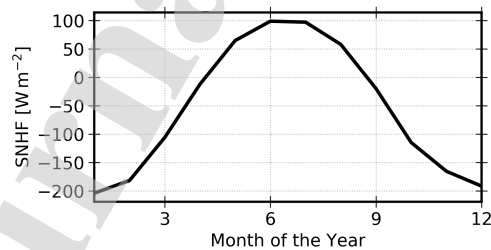


Figure 4: Surface net heat flux imposed in idealized simulations with convection.

136 *2.1.5. Sum up*

137 A total of 16 simulations have been run and analyzed, to study the impact of the dif-  
 138 ferent parameters (presence and depth of the SCV shotgun, presence of the RT eddy at  
 139 initialization, convection, bottom drag) on the RT eddy intensity and dynamics. The differ-  
 140 ent simulations are summarized in Table 1.

Name	SCV shotgun	RT eddy at initialization	Bottom Drag	Convection
dD	deep	—	yes	—
mD	middle	—	yes	—
m	middle	—	—	—
sD	surf	—	yes	—
RTD	—	yes	yes	—
RT	—	yes	—	—
RTDC	—	yes	yes	yes
dRTD	deep	yes	yes	—
dRTDC	deep	yes	yes	yes
mdRTD	middle deep	yes	yes	—
mRTD	middle	yes	yes	—
mRT	middle	yes	—	—
mRTDC	middle	yes	yes	yes
msRTD	middle surf	yes	yes	—
sRTD	surf	yes	yes	—
DC	—	—	yes	yes

Table 1: Parameters of the analyzed simulations

141 mRTDC is the simulation closest to the reality, as it includes all features and forcings  
 142 occurring in the RT area: SCVs generated at a realistic depth, convection, bottom drag, and  
 143 a anticyclonic eddy at the center of the bowl-like topography.

144 *2.2. Diagnostics*

145 We describe in this section the diagnostics performed on the simulation outputs.

### 146 2.2.1. Detection of the main eddy

147 In each simulation, we detect the main eddy using the Angular Momentum Eddy Detec-  
 148 tion and tracking Algorithm (AMEDA, Le Vu et al., 2018). One of the benefits of AMEDA  
 149 is that it does not depend on arbitrary thresholding, which would require a fine-tuning of  
 150 geometrical parameters. Also, the algorithm is robust with respect to the grid resolution  
 151 and can thus be applied to a wide variety of velocity fields (experimental, numerical, derived  
 152 from altimetry). This algorithm has been used and validated in previous –observational  
 153 and numerical– studies (Ioannou et al., 2017; Le Vu et al., 2018; Garreau et al., 2018;  
 154 de Marez et al., 2019; de Marez et al., 2020), see also an example of application of AMEDA in  
 155 <https://www1.lmd.polytechnique.fr/dyned/>. This algorithm works as follows: (a) from  
 156 the velocity fields, it computes the local normalized angular momentum (LNAM, Mkhinini  
 157 et al., 2014) and the local Okubo-Weiss parameter (LOW) at each point; (b) then, it seeks  
 158 LNAM local maxima where  $LOW < 0$ ; (c) if these maxima are surrounded by a closed stream-  
 159 line, they are flagged as eddy centers. A full description of the algorithm is presented in  
 160 Fig. 1 of Le Vu et al. (2018). In this study, the detection is done using daily velocity fields,  
 161 at -250, -500, -750, -1000, and -1250 m depth for surf, middle surf, middle, middle deep,  
 162 and deep cases respectively. Choosing the depth of detection as equal to the depth of SCV  
 163 generation ensures an accurate estimation of the radius increase of the main eddy when it  
 164 merges with SCVs. The main eddy’s edge is defined as its contour of maximal velocity. The  
 165 mean radius of this contour at a given time is  $R_{max}$ . We use this contour to compute volume  
 166 integrated quantities, assuming that the eddy is roughly cylindrical.

167 In simulations with a composite anticyclonic eddy at initialization, the main eddy is  
 168 simply the initial eddy, that we follow in time. For simulations with no eddy at  $t=0$ , the  
 169 main eddy is defined as the first SCV that reaches the center of the bowl-like topography  
 170 and then grow in size due to merging with other SCVs.

### 171 2.2.2. Kinetic energy budget

172 In the primitive equation framework, the kinetic energy (KE) equation can be obtained  
 173 by taking the inner product of the horizontal velocities with the momentum equations. It  
 174 follows:

$$\frac{1}{2}\partial_t u_i^2 + u_j \partial_j \left(\frac{1}{2}u_i^2\right) + w \partial_z \left(\frac{1}{2}u_i^2\right) = \frac{u_i}{\rho_0} \partial_i P + \mathcal{V}_i u_i + \mathcal{D}_i u_i + \mathcal{S}_i u_i, \quad (1)$$

175 with summation convention,  $i = 1, 2$ , and  $j = 1, 2$ ,  $u_i$  are the horizontal component of  
 176 velocity,  $\partial_i$  the components of the vector differential operator,  $\mathcal{V}_i$  the components of the  
 177 parameterized vertical mixing,  $\mathcal{D}_i$  the components of the horizontal diffusion, and  $\mathcal{S}_i$  other  
 178 sources and sinks (due to restoring, nudging, boundary conditions...). This equation is then  
 179 vertically integrated, and we define:

- 180 •  $\text{hadv} = \int dz u_j \partial_j \left(\frac{1}{2}u_i^2\right)$ ,
- 181 •  $\text{vadv} = \int dz w \partial_z \left(\frac{1}{2}u_i^2\right)$ ,
- 182 •  $\text{Prsgrd} = \int dz \frac{u_i}{\rho_0} \partial_i P$ ,
- 183 •  $\text{vmix} = \int dz \mathcal{V}_i u_i$ ,
- 184 •  $\text{hmix} = \text{explicit part of } \int dz \mathcal{D}_i u_i$ ,
- 185 •  $\text{hdiff} = \text{implicit part of } \int dz \mathcal{D}_i u_i$ ,
- 186 •  $\text{nudg} = \int dz \mathcal{S}_i u_i$ ,
- 187 •  $\text{cor} = \int dz (fuv - fvu)$ ,
- 188 •  $\text{vol} = \text{the depth integrated KE variations due to the grid breazing,}$
- 189 •  $\text{Drag} = \text{contribution of the bottom drag parameterization in the vmix term.}$

190 All these terms are computed online (Gula et al., 2016). The closed KE budget is:

$$\partial_t \int dz \frac{1}{2}u_i^2 = \text{hadv} + \text{vadv} + \text{Prsgrd} + \text{vmix} + \text{hmix} + \text{hdiff} + \text{nudg} + \text{cor} + \text{vol}. \quad (2)$$

191 We integrate these terms in time, such that for instance  $\int_0^t dt \text{ Drag}$  represents the contribution  
 192 of the bottom drag for the KE at a given time  $t$ . Finally, we horizontally integrate the results  
 193 in the main eddy's contour  $S$  (calculated by AMEDA). This allows to follow in detail which  
 194 physical mechanism is responsible for the evolution of the main eddy's KE.

### 195 3. Results

196 In this section, we describe the results of our study. We first explain qualitatively the  
197 course of a simulation representative of the RT, *i.e.*, the mRTD simulation. Then we discuss  
198 the impact of the different parameters on the evolution of the main eddy.

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## 199 3.1. Qualitative evolution of the simulations

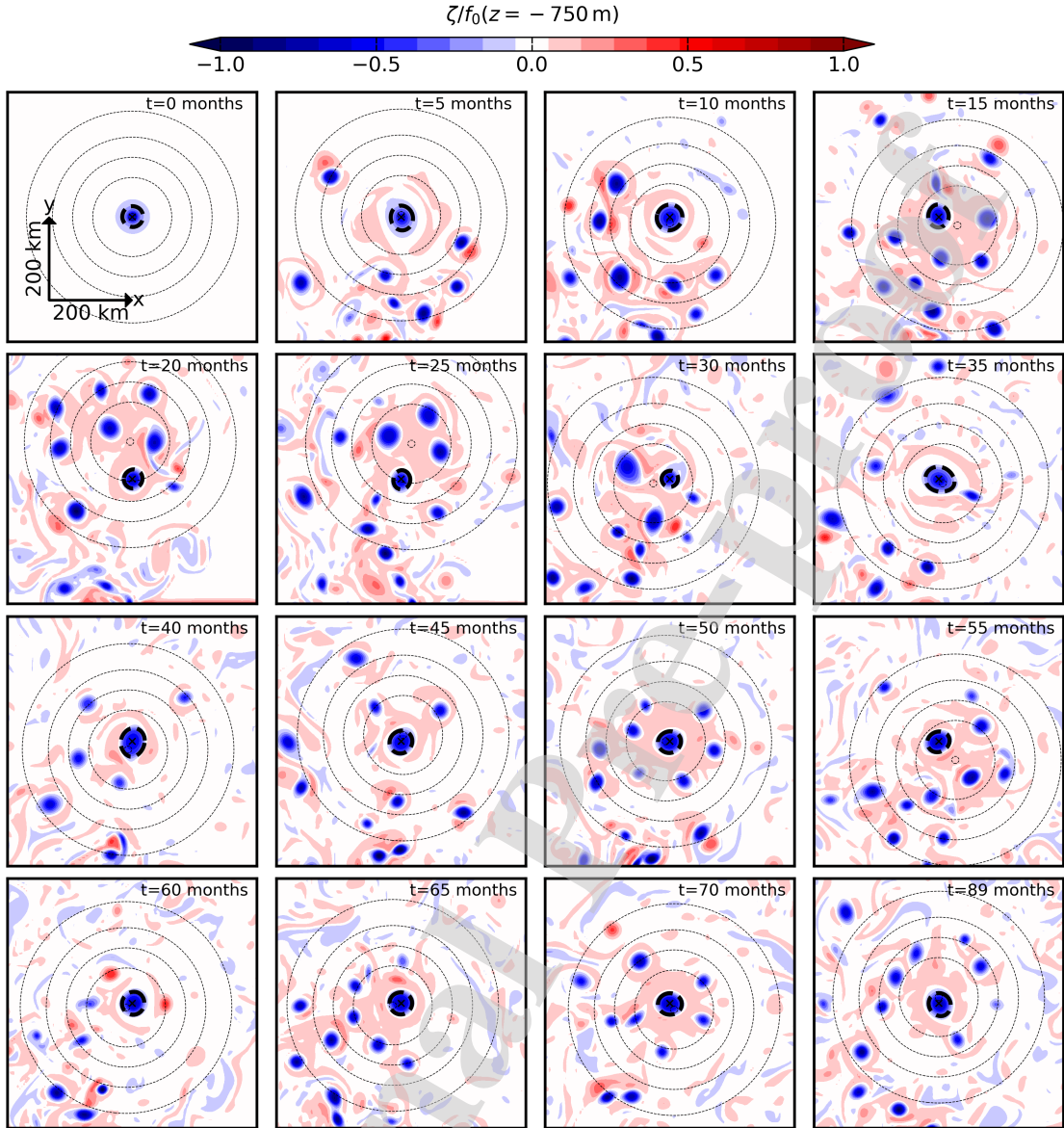


Figure 5: Snapshots of normalized relative vorticity at 750 m depth, in the mRTD simulation. Each panel is 600 km large, and is centered around the main eddy. Bold dashed contours indicate the contour of maximum velocity of the main eddy. Thin dashed contours show isobath from 2500 to 2000 m depth with a 100 m interval.

200 In the mRTD simulation, the main eddy is present from  $t = 0$  at the center of the bowl  
 201 topography (see Fig. 5 for the time evolution of the mRTD simulation). Because it is  
 202 anticyclonic, it is stuck in the center of the bowl to conserve its potential vorticity (Carnevale

203 et al., 1991). Thus, it cannot climb out without external disturbance. During the first year  
 204 of the simulation, the main eddy does not move from the center of the bowl. Its volume  
 205 slightly increases because of a azimuthal mode 2 and mode 4 destabilization, similar to the  
 206 one observed in de Marez et al. (2020a) (a vorticity tripole can be seen at  $t = 5$  months in  
 207 Fig. 5).

208 Simultaneously, SCVs are generated at the corner of the land mask. As shown in Derem-  
 209 ble et al. (2016), such boundary singularities generate dipolar structures. In our case, about  
 210 35 dipoles are generated each year. The anticyclonic pole is attracted by the bowl, while  
 211 the cyclonic pole tends to step aside. Indeed, on a slope, the topographic  $\beta$ -drift makes  
 212 anticyclones (resp. cyclones) drift downhill (resp. uphill) (LaCasce, 1998; Lam & Dritschel,  
 213 2001). This leads to the separation of about all dipoles in two monopoles of opposite vor-  
 214 ticities. Also, in some cases, the positive pole is rolled up around the anticyclone, leading  
 215 to a shielded anticyclonic SCV. In both cases, this results into anticyclonic SCVs attracted  
 216 by the bowl. However, they do not all reach the bowl (*i.e.*, the places where the floor is  
 217 deeper than 2000 m): only about ten anticyclonic SCVs per year manage to reach it. This  
 218 is mainly due to the fact that just after generation, SCVs merge between each other, and  
 219 subsequently converge to the center of the bowl.

220 At the start of the second year of simulation, SCVs start to interact with the main eddy.  
 221 These SCVs have two effects.

222 First, the SCVs' velocity field slightly disturb the main eddy, resulting in small displace-  
 223 ments of the main eddy in the bowl. From this date, the main eddy can thus be found at  
 224 tens of kilometers from the center of the bowl, see *e.g.*  $t = 15, 25,$  or  $55$  months, in Fig. 5.

225 Second, SCVs merge with the main eddy. In this simulation —*i.e.*, during about 7 years  
 226 and a half— 41 merging events between SCVs and the main eddy are observed. Two kinds  
 227 of merger occur. On the one hand, small SCVs that have experienced a few merging with  
 228 other SCVs before reaching the bowl, are attracted by the main eddy and steered around it.  
 229 They are finally absorbed by the main eddy which eventually grows by aggregating vorticity  
 230 and azimuthal velocity outside of its core (Sutyrin & Radko, 2019; Sutyrin, 2019). This  
 231 mechanism is referred to as *Vortex Thinning* in the literature. A vortex thinning event  
 232 results in a small increase of the eddy volume; examples can be seen at  $t = 15$  or  $70$  months

233 in Fig. 5. On the other hand, the main eddy can merge with SCVs that have grown in size  
234 due to successive merging with other SCVs. This results in situations where the main eddy  
235 is either of the same size of the vortex it merges with, or smaller than it. In both cases, the  
236 merging is roughly symmetric, and it results in an abrupt increase of volume of the main  
237 eddy. An example can be seen between  $t = 30$  and 35 months in Fig. 5.

238 These mergers can also be called alignment, because the main eddy's core density is not  
239 necessarily the same than its companion's one (see *e.g.* Nof & Dewar (1994), and sections  
240 7.1.3 and 7.2.4 of Lilly et al. (2003) that discuss cases of vortex alignment in the Labrador  
241 Sea). It can be mentioned that, in a stratified fluid, mergers must manifest as alignment  
242 because there will always be small differences in the core density of the two eddies. The  
243 impact of this three-dimensional view is discussed in the next sections. Also, we detail in  
244 the following the physical mechanism that occur in the simulations, and lead to changes of  
245 volume for the main eddy.

### 246 3.2. Mechanisms of eddy growth

247 In all simulations, as in *e.g.*, mRTD (see Fig. 5), the main eddy growth is intermittent  
248 and occurs at specific moments of the simulation. As discussed in the introduction, this  
249 eddy growth can be attributed to two mechanisms: merging with vortex companions, and  
250 convection. We detail in this section these physical mechanisms.



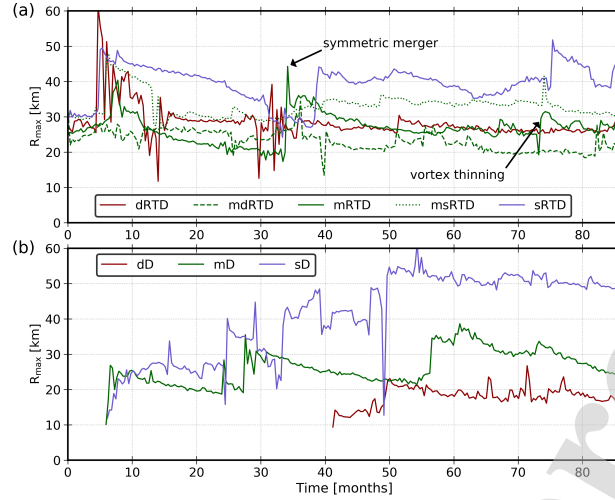


Figure 6: Evolution of  $R_{max}$  during simulations, for different initialization depth of the SCV shotgun (a) with and (b) without the RT composite at the center of the bowl. All simulations include bottom drag parameterization.

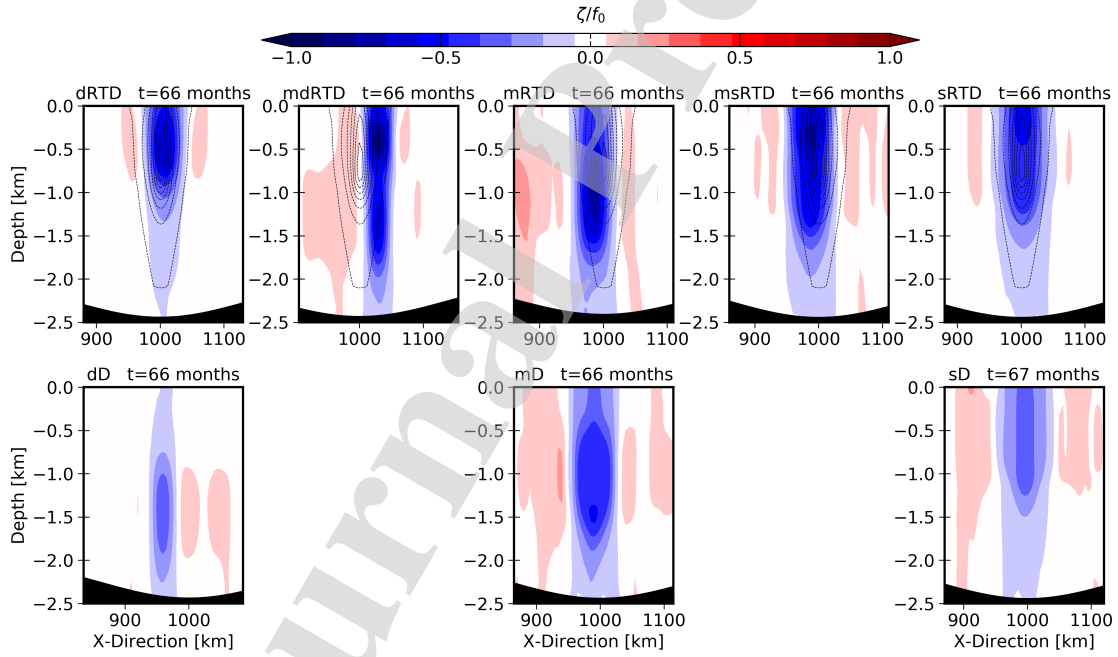


Figure 7: Vertical sections of normalized relative vorticity passing through the center of the main eddy after  $\sim 5$  and a half years of simulation, for simulations shown in Fig. 6. Thin contours in the top row show the contours of normalized vorticity through the center of the RT composite at initialization; note that line and color contours are shown for the same vorticity values.

252 To discuss the impact of merger on the main eddy, we analyze the time evolution of  $R_{\max}$   
 253 in different simulations, see Fig. 6. Both kinds of merging events discussed in the previous  
 254 section for mRTD simulation –vortex thinning and symmetric merger– can be seen in Fig.  
 255 6(a). They appear as steps in the time evolution of  $R_{\max}$ , at *e.g.*,  $t = 35$  or 75 months (in  
 256 the mRTD simulation, see arrows in Fig. 6(a)). In all simulations, symmetric mergers have  
 257 a greater impact on  $R_{\max}$ , as it can double the radius of the main eddy in a few days.

258 In all simulations with the RT composite at initialization (Fig. 6(a)), the radius of the  
 259 main eddy oscillates around about 30 km. After periods of radius decrease, the merging  
 260 efficiently increases the radius. The horizontal extension of the main eddy is thus similar to  
 261 its initial one after 7 years of simulation. It can be noticed that in the sRTD simulation,  
 262 the main eddy appears to have a larger radius than in other simulations. Even with this  
 263 difference, the purely horizontal view described by the time evolution of  $R_{\max}$  is qualitatively  
 264 similar in dRTD, mdRTD, mRTD, msRTD, and sRTD.

265 The main difference between the simulations is the final vertical shape of the main eddy,  
 266 *e.g.*, after 66 months (Fig. 7). In the dRTD simulation, the main eddy has a 3D shape  
 267 similar to the initial RT composite. This is due to the fact that SCVs have difficulty to  
 268 merge with the main eddy. Oppositely, in the sRTD and msRTD simulations, SCVs easily  
 269 merge with the main eddy, and thus drastically modify its 3D shape. The final shape of the  
 270 main eddy differs from the RT composite because it is intensified at the surface, with a larger  
 271 horizontal extension due to the numerous mergers it experienced. In the mRTD and mdRTD  
 272 simulations, the merging events lead to an important intensification of the eddy intensity at  
 273 depth. In the mdRTD simulation, the final shape of the main eddy is a double-core eddy,  
 274 with two vorticity maxima, at  $\sim 500$  m depth and  $\sim 1300$  m depth.

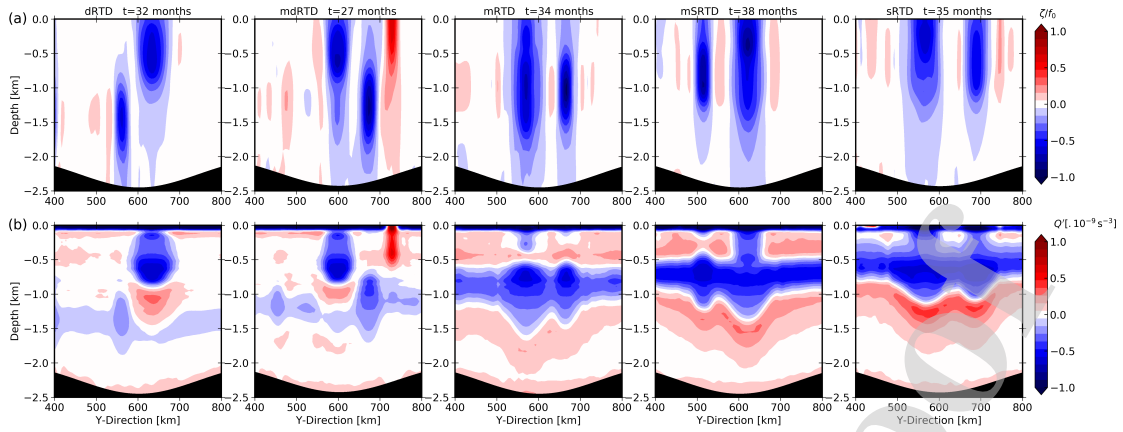


Figure 8: Vertical section of (a) normalized relative vorticity and (b) associated PV anomaly, at times just before a merger of the main eddy with a companion eddy, in dRTD, mdRTD, mRTD, msRTD, and sRTD simulations. All sections pass through the center of both eddies.

275 The merging efficiency is not the same in all simulations because of (1) the background  
 276 stratification, and (2) the vertical structures of the main eddy and the SCVs it merges  
 277 with. Indeed, Verron et al. (1990); Verron & Valcke (1994); Corr  ard & Carton (1999)  
 278 altogether showed, using 2-layer numerical simulations, that the merger (or alignment) of  
 279 two like-signed vortices depends on their shape before the merging. Vortices can be separated  
 280 into two kinds: PVI (potential vorticity initialization) vortices and RVI (relative vorticity  
 281 initialization) vortices. PVI vortices are represented by a patch of constant potential vorticity  
 282 in a single layer –and in some cases a vertical dipole of potential vorticity–, associated with  
 283 nearly barotropic relative vorticity. RVI vortices have a constant relative vorticity in a single  
 284 layer. Corr  ard & Carton (1999) showed that PVI vortices easily align together while RVI  
 285 vortices do not. Verron et al. (1990); Verron & Valcke (1994) showed that the ambient  
 286 stratification plays a different role in the merging depending on the vortex shape: RVI  
 287 vortex merger strongly depends on the stratification while PVI vortex merger does not. If  
 288 the stratification is weak, RVI vortices form a pair of heton-like structures, that repel each  
 289 other. If the stratification is stronger, the ambient flow is more barotropic, and merger is  
 290 easier. In a configuration more realistic than the 2-layer quasi-geostrophic model, like in our  
 291 study, the distinction between RVI or PVI vortices can be tricky because of the Gaussian  
 292 vertical shape that eddies often take (McWilliams, 1985).

293 In our simulations, the categorization of eddies is difficult because the main eddy and its  
294 companion eddies can be categorized as PVI-like vortices (they appear as –roughly constant–  
295 PV patches confined in a single layer, see Fig. 8), but also RVI-like vortices (the maximum  
296 of relative vorticity is confined in  $\sim 1000$  m deep layers, around which the relative vorticity  
297 either changes sign or is close to zero). Merging vortices are thus PVI/RVI hybrids. The  
298 background stratification should thus play a role in the merger efficiency. Our background  
299 stratification is similar to the RT case (Fig. 1(d)) and it has Brunt-Väisälä frequency maxima  
300 at the surface and near 750 m depth. Thus for RVI vortices, the merger is facilitated at these  
301 particular depths, while it is harder to merge for deeper eddies. This could explain in part  
302 why SCVs have more difficulties to merge with the main eddy in the dRTD simulation than  
303 in the other simulations.

304 One can observe a critical depth for the SCV shotgun, between 1000 and 750 m depth,  
305 for which the merging/alignment of eddies do not lead to the intensification of the main  
306 eddy’s core. If SCVs are close enough (in both vertical and horizontal directions), they  
307 eventually align with the main eddy, but this only results in a deepening of the eddy and/or  
308 a double-core eddy, with no influence on the original eddy core. Because of the complicated  
309 form of eddies and ambient stratification, it is here difficult to be more quantitative about  
310 the key parameters that influence the merging. A more extensive study in the parameter  
311 space would be necessary to discuss in details the alignment of vortices in a 3D primitive  
312 equation framework.

313 If no initial RT composite is present, the time evolution of  $R_{\max}$  (Fig. 6(b)) is roughly  
314 similar to cases described above. However, the main eddy vertical structure near the end  
315 of the simulation is strongly influenced by the SCVs generation depth (Fig. 7). In the  
316 dD simulation, SCVs hardly merge, because of the weak stratification below 1000 m depth,  
317 leading to a weak resulting eddy, intensified at depth. In the sD simulation, SCVs are not  
318 very intense, but they easily merge, leading to a weak surface-intensified anticyclone, that  
319 does not resemble the RT eddy either. In the mD simulation, the final shape of the main  
320 eddy is roughly similar to the mRTD simulation’s one. This show that the merging of SCVs  
321 between each other produces an eddy similar to the RT eddy only if SCVs are generated at  
322 a realistic depth, where merger is easier because of the strong stratification. This supports

323 the view of Smilenova et al. (2020), that from an ocean at rest, the RT eddy results from a  
 324 succession of merging events between SCVs generated at  $\sim 750$  m depth.

325 *3.3. On the importance of convection*

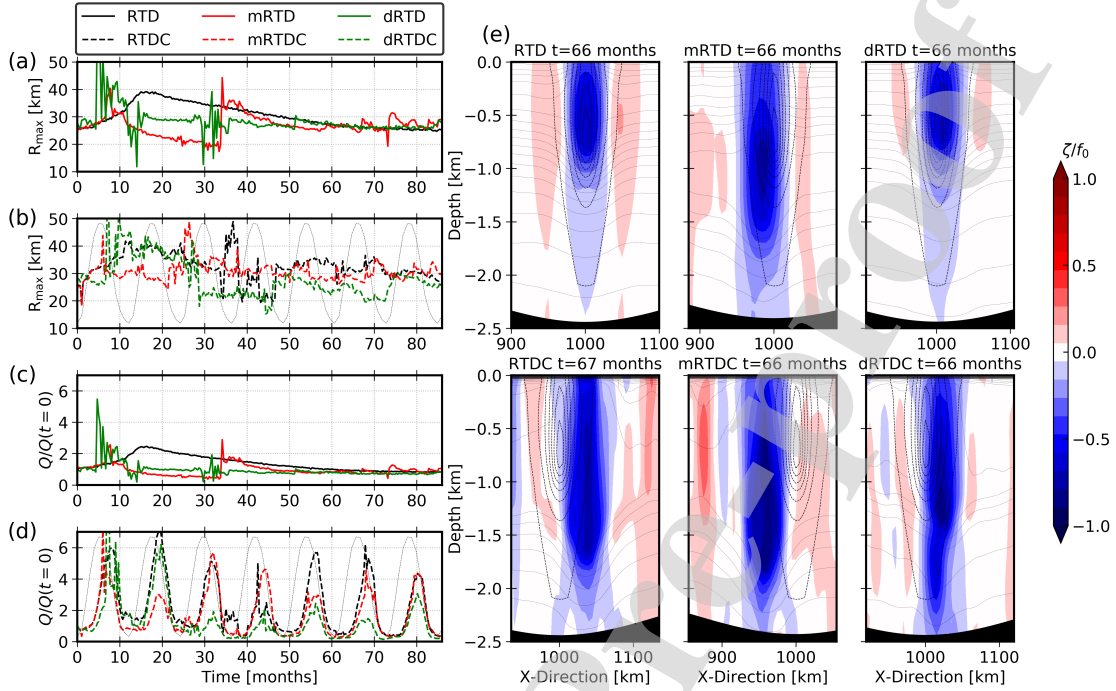


Figure 9: (a,b) (resp. (c,d)) Time evolution of  $R_{\max}$  (resp. ratio between PV and initial PV integrated over the main eddy) for some simulations without (a,c, solid line) or with (b,d, dashed lines) convection; the thin gray line shows the time evolution of the SNHF applied at the surface in the simulations with convection (see Fig. 4 for the values it reaches). (e) Same as Fig. 7 for the 6 simulations shown in (a,b).

326 If we add a negative heat flux at the surface (as in RTDC, mRTDC, dRTDC or DC  
 327 simulations), convection appears. As a result, isopycnals deepen during about 6 months  
 328 each year, following the imposed seasonal cycle (Fig. 4). The Ertel potential vorticity  $Q$   
 329 defined as

$$Q = (f_0 + \zeta)\partial_z b - (\partial_z v)(\partial_x b) + (\partial_z u)(\partial_y b), \quad (3)$$

330 with  $f_0$  the Coriolis frequency,  $b$  the buoyancy, and  $\zeta$  the relative vorticity, undergoes a  
 331 seasonal increase in the main eddy's core (see Fig. 9(b)). The shallower the SCV shotgun is,  
 332 the larger this intensification. This mechanism of intensification by wintertime convection

333 is rather multifaceted. The deepening of the core intensifies radial density gradients and  
334 subsequently increases the azimuthal velocity of the eddy. During these convective periods  
335 the main eddy is thus intensified, and its radius increases (see Fig. 9(a)). On the other  
336 hand, when density gradients are increased the eddy is no longer in thermal wind balance,  
337 and therefore an adjustment mediated by a secondary circulation is required. We refer the  
338 reader to Legg et al. (1998); Legg & McWilliams (2001) for a full explanation of this process.

339 If we compare the RTD and the RTDC simulations, one can see that even if no SCV  
340 shotgun is present, the evolution of the main eddy's radius experiences sharp increases in  
341 the presence of convection. This is due to the fact that during convective periods, the  
342 surface mixed-layer deepens in the whole domain, and SCVs are spontaneously generated  
343 by mixed-layer baroclinic instabilities (Callies et al., 2015). This leads to vortex thinning  
344 events between convectively-generated SCVs and the main eddy, that subsequently increase  
345 the main eddy's radius (Schubert et al., 2020). These events appear throughout the whole  
346 simulation involving convection, and they are difficult to characterize because they can be  
347 generated above the main eddy and merge with it within a few days. However, it can  
348 be noticed that in the simulation with both a SCV shotgun at mid-depth and convection  
349 (mRTDC), we observe that the main eddy's radius continuously oscillates between a mean  
350 value of about 30 km. In this simulation, the SCVs are more numerous than in the mRTD  
351 simulation. The number of merging events of small SCVs with the main eddy is larger when  
352 convection is present. This leads to a lot of small increases of radius rather than decrease  
353 periods followed by a large increase.

354 Convection and merging events with convectively-generated SCVs act together to in-  
355 crease the horizontal shape of the main eddy, but also its intensity at depth. Indeed, with  
356 convection, the main eddy is more intensified at depth than in simulation without convection,  
357 see *e.g.*, Fig. 9(c) for the dRTDC simulation. The main eddy is intensified at a depth where  
358 no merging with neither the convectively-generated SCVs in the mixed-layer nor the SCV  
359 shotgun SCVs occurs. This reflects the importance of the direct convectively driven mode.  
360 As mentioned in the previous section, if the SCVs are generated below a critical depth (*e.g.*  
361 in dRTD simulation), little merging events are observed. If we add the convection (dRTDC  
362 simulation), the main eddy's core deepens. Subsequently, the vertical distance between the

363 main eddy's core and SCVs decreases, and merger/alignment is eased. This leads to a deep  
364 intensification of the main eddy, see Fig. 9(c). Convection thus allows to intensify the main  
365 eddy's core by 1) deepening the core during wintertime, 2) generating small SCVs –by mixed-  
366 layer instabilities– that can eventually merge with the main eddy, and 3) helping alignment  
367 by reducing the distance between the main eddy's core and deeply-generated SCVs. It should  
368 be noticed however that from our analysis, the relative importance of these three mechanisms  
369 cannot be precisely gauged. This quantification should be the aim of further investigations.

370 As a sensitivity test, we also ran a simulation with only convection (DC simulation, not  
371 shown). In this simulation, anticyclonic convectively-generated SCVs are attracted by the  
372 bowl-shaped topography. They subsequently merge between each other, and eventually form  
373 a  $\sim 50$  km radius anticyclone in the bowl, *i.e.*, a main eddy. This eddy appears after about  
374 33 months of simulation and is intensified between 1000 and 1500 m depth. This test further  
375 shows that the convection by itself is sufficient to lead to a single long-lived anticyclonic  
376 eddy in a bowl topography.

#### 377 3.4. Mechanisms of eddy decay

378 As can be seen in the time evolution of  $R_{\max}$  (see *e.g.*, Fig. 6), after the main eddy's  
379 radius increases due to either merging or convection, periods of decay that can last for  
380 several years are observed. During these periods, little merging occur, and some physical  
381 mechanisms lead to the erosion of the eddy. We describe those in the following section.

## 382 3.4.1. Bottom drag

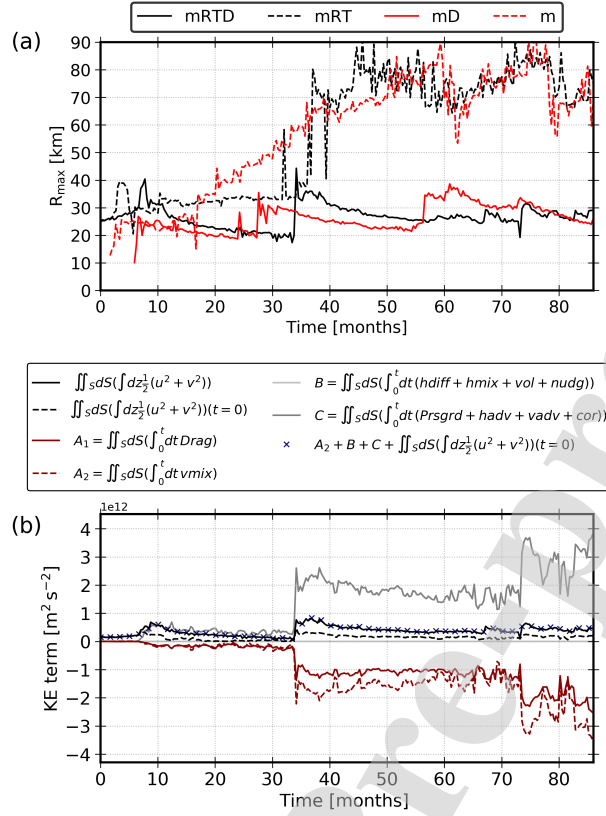


Figure 10: (a) Evolution of  $R_{max}$  during simulations, with (solid) and without (dashed) bottom drag parameterization. (b) Evolution of kinetic energy terms in the mRTD simulation. Each term is integrated in time and in the contour of the main eddy. Note that the superposition of blue crosses with the black solid lines shows that the KE energy budget is closed, with respect to eq. (2).

383 The principal mechanism responsible for the eddy decay in our simulations is the bottom  
 384 drag. We observe that no radius decay period are seen in simulations without bottom drag  
 385 parameterization (Fig. 10(a)). This leads to a main eddy being too intense, and too large  
 386 in comparison with the RT eddy. Also, without drag, the main eddy becomes anomalously  
 387 barotropic (see for instance Fig. 11(c)). It thus has a 3D shape very different from the RT  
 388 eddy.

389 The bottom drag seems to be the major limiting factor for the eddy growth due to merging  
 390 with SCVs. In the KE equation budget (Fig. 10(b)) the pressure gradient, the advection  
 391 and the Coriolis terms dominate the main eddy's KE gain (C in Fig. 10(b)). They are the



392 result of merging with other coherent structures (SCVs), and conversion from potential to  
393 kinetic energy (not shown). The horizontal diffusion and mixing, as well as volume change  
394 and nudging have a neglectable contribution to the KE budget (B). The bottom drag ( $A_2$ ),  
395 included in the vertical mixing ( $A_1$ ), appears to dominate the main eddy's KE loss. It  
396 compensates the other terms, and increases in amplitude each time the eddy gains KE by  
397 merging with other vortices. The bottom drag contribution is intensified when the main eddy  
398 drifts away from the center of the bowl, because the water depth is smaller. It is thus greater  
399 just before symmetric merger events, because the main eddy co-rotates with its companion,  
400 and subsequently drifts away from the center of the bowl. This can be seen in Fig. 10(b),  
401 with Drag KE term peaking just before merging related steps (see at *e.g.*  $t = 35$  months).

## 402 3.4.2. Centrifugal instability

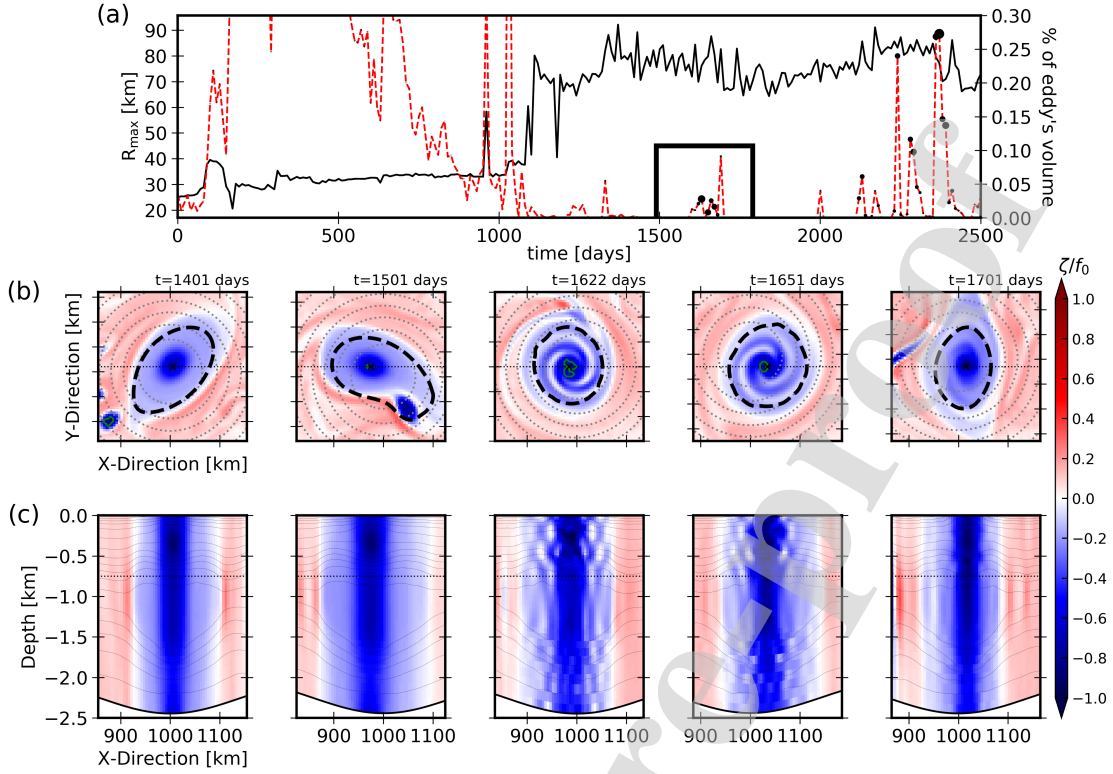


Figure 11: (a) Time evolution of  $R_{\max}$  (black line) and % of eddy's volume with negative PV (dashed red) for the mRT simulation. The size of black dots indicates the amplitude (in absolute value) of the minimum PV in the eddy's contour. (b) Horizontal sections of normalized relative vorticity at 750 m depth; green contours indicate the places where the PV is negative. (c) Vertical sections of normalized relative vorticity passing through the center of the main eddy; the thin dashed lines indicate the depth of horizontal sections shown in (b).

403 In cases without drag, it can be seen that the main eddy still experiences abrupt radius  
 404 decrease events, see *e.g.*, between  $t = 1500$  and  $1700$  days in mRT simulation (Fig. 11(a)).  
 405 At this time, a rapid radius decrease is seen just after a vortex thinning event with a small  
 406 SCV. Horizontal sections of relative vorticity (Fig. 11(b)) show that after the main eddy  
 407 absorbs the SCV, a spiral-like pattern appears in the eddy's core. This pattern is seen in  
 408 the whole water column. There, the normalized relative vorticity reaches  $\zeta/f_0 \sim -1$ .

409 This pattern is typical of centrifugal instability (Cushman-Roisin & Beckers, 2011). This  
 410 diagnostic is confirmed by the following facts. (1) The PV in the eddy's core is negative

411 near its center (green contours in Fig. 11(b)), which is the necessary condition ( $fQ < 0$ )  
 412 for centrifugal instability. (2) The horizontal shear terms are responsible for the extreme  
 413 decrease of PV in the eddy's core (not shown).

414 Negative PV patches are strongly unstable, and the nearly materially conserved nature  
 415 of PV implies that negative PV does not occur spontaneously inside the fluid. Thus, the  
 416 generation of negative PV in the fluid must be forced, for instance by appropriate frictional  
 417 interactions with nearby boundaries or interactions with the wind. Here, no such mechanism  
 418 is present. The decrease of PV is due to the abrupt change of horizontal velocity gradients  
 419 resulting from the vortex thinning of a small SCV around the main eddy (Fig. 11(b)). The  
 420 change of PV occurs where density fronts are sharp and parameterized diapycnal mixing  
 421 occurs (see Appendix C in de Marez et al. (2020a)). The centrifugal instability is thus  
 422 triggered by the interaction of the main eddy with the SCV, and eventually leads to an  
 423 abrupt erosion of the eddy. About five major centrifugal instability events occur at  $t > 1000$   
 424 days, see the peaks of red dashed curve in Fig. 11(a), that show times when negative PV is  
 425 seen in the eddy's core. Note that before  $t = 1000$  days, negative PV is found in the eddy's  
 426 core, but with values very close to zero. The eddy is thus at this moment not intense enough  
 427 to be subject to the instability.

428 Such centrifugal instabilities are preferably seen in simulations without drag. In those,  
 429 the main eddy is more intense, and it reaches very low PV values that are suitable for  
 430 instabilities. Nevertheless, such instabilities can still be seen in *e.g.*, mRTD simulation, with  
 431 a smaller signature than in mRT (not shown).

#### 432 4. Summary and discussion

433 We studied the lifecycle of an anticyclonic eddy trapped in a bowl-like topography, which  
 434 is subject to the interaction with like-signed SCVs and/or convection. From the analysis  
 435 of 16 simulations with varying parameters, we show that *the balance between merger and*  
 436 *bottom drag* allows the eddy to have a roughly constant 3D shape throughout several years.  
 437 On the one hand the vortex merger with small SCVs allows the eddy to grow in size, and  
 438 intensify at depth. As merger events occur at the SCV generation depth, the final main eddy  
 439 is intensified at this particular depth. These mergers are enhanced when SCVs are generated

440 at a depth where the stratification is large, and when convection is at work. Indeed, the  
441 convection (1) deepens the main eddy and increases the merger efficiency at depth, and (2)  
442 generates other SCVs in the mixed-layer that eventually merge with the main eddy. On the  
443 other hand the bottom drag erodes the eddy. The bottom drag is the main contribution to  
444 the eddy's KE loss. When it is not included in simulation, the eddy becomes barotropic, and  
445 centrifugal instabilities triggered by the merger with SCVs erode the eddy over the whole  
446 water column.

447 Merging and convection both contribute to the maintaining of anticyclonic eddies trapped  
448 in bowl-shaped topographies. Our study thus show that it is difficult to disentangle the two  
449 mechanisms, in particular if we replace this in a more realistic context. However, the final  
450 shape of the main eddy can give keys about the mechanisms responsible for the long lifetime  
451 of such eddies. If the main eddy is intensified at multiple depth (it has *e.g.*, a double-core  
452 eddy), the eddy has certainly experienced one or several merging with other vortices. Also, if  
453 the depth of intensification of the main eddy is correlated with the depth of a SCV generation  
454 site nearby, merger between these vortices probably happened. One can therefore state that  
455 in these cases, merging played a major role in the maintaining of the eddy.

456 In the real ocean, other processes can affect the shape of such an eddy. For instance,  
457 internal waves and fine-scale ( $\mathcal{O}(1)$  m) processes can lead to the dissipation of long-lived  
458 mesoscale eddies. In the LV case, Fer et al. (2018) showed through high-resolution turbu-  
459 lence measurements that the background shear as well as near-inertial waves trapped by the  
460 negative vorticity of the LV are the dominant sources of kinetic energy loss. More generally,  
461 internal waves are suspected to drain a significant part of the energy of such mesoscale eddies  
462 (Barkan et al., 2021). These mechanisms are hardly resolved in the simulations discussed  
463 in the present paper, and are mainly controlled by the numerical parameterization (*i.e.*, the  
464 vertical mixing induced by the KPP scheme). Furthermore, the simulation lacks realistic lev-  
465 els of internal waves. Simulations with higher resolution and fully realistic atmospheric and  
466 tidal forcings should thus be required in order to determine the relative importance of these  
467 other processes compared to the bottom drag. At larger scales opposite-signed mesoscale  
468 coherent structures can travel to the eddy's location. This could modify the behavior of  
469 the eddy by dipolar effect, and affect the merging efficiency with SCVs (Rodríguez-Marroyo

470 et al., 2011). Furthermore, the presence of a mean current due to large-scale circulation or  
471 local coastal current can erode the eddy because of the presence of an ambient horizontal  
472 shear (Perrot & Carton, 2010). If the eddy moves toward the coast, the interaction with  
473 coastal Kelvin waves can also affect its trajectory and shape (Dewar & Hogg, 2010; Gula &  
474 Zeitlin, 2010; Hogg et al., 2011; de Marez et al., 2020b).

475 Despite this, in the Rockall Trough (as well as in *e.g.*, the Lofoten Basin) the semi-  
476 permanent anticyclonic eddies are rather isolated from the coast and other currents. Our  
477 study can thus support the view of Smilenova et al. (2020) or Trodahl et al. (2020) that such  
478 semi-permanent anticyclonic eddies are mainly maintained by the merger (or alignment) with  
479 smaller-scale vortices. In this high latitude regions, convection is large, and indeed deepens  
480 isopycnal and subsequently increases the eddy's core potential vorticity, as discussed in  
481 *e.g.*, Bosse et al. (2019). However, we show here that convection principally enhances the  
482 number of merger with small eddies, either at the surface or at depth with SCVs. The  
483 merging/alignment with SCVs is thus likely to be the more important mechanism to sustain  
484 mesoscale anticyclones trapped in a bowl, as stated in the LV case by Trodahl et al. (2020)  
485 and by Smilenova et al. (2020) in the RT eddy case.

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492 **References**

- 493 Barkan, R., Srinivasan, K., Yang, L., McWilliams, J. C., Gula, J., & Vic, C. (2021). Oceanic  
494 mesoscale eddy depletion catalyzed by internal waves. *Earth and Space Science Open*  
495 *Archive*, (p. 14). URL: <https://doi.org/10.1002/essoar.10507068.1>. doi:10.1002/  
496 [essoar.10507068.1](https://doi.org/10.1002/essoar.10507068.1).
- 497 Bosse, A., Fer, I., Lilly, J. M., & Soiland, H. (2019). Dynamical controls on the longevity  
498 of a non-linear vortex : The case of the Lofoten Basin Eddy. *Scientific Reports*, *9*.  
499 doi:10.1038/s41598-019-49599-8.
- 500 Bosse, A., Testor, P., Houpert, L., Damien, P., Prieur, L., Hayes, D., Taillandier, V., Dur-  
501 rieu de Madron, X., d'Ortenzio, F., Coppola, L., Karstensen, J., & Mortier, L. (2016).  
502 Scales and dynamics of Submesoscale Coherent Vortices formed by deep convection in  
503 the northwestern Mediterranean Sea: Vortices in the NW Mediterranean Sea. *Journal of*  
504 *Geophysical Research: Oceans*, *121*, 7716–7742. doi:10.1002/2016JC012144.
- 505 Callies, J., Ferrari, R., Klymak, J. M., & Gula, J. (2015). Seasonality in submesoscale  
506 turbulence. *Nature Communications*, *6*. doi:10.1038/ncomms7862.
- 507 Carnevale, G. F., Kloosterziel, R. C., & Van Heijst, G. J. F. (1991). Propagation of barotropic  
508 vortices over topography in a rotating tank. *Journal of Fluid Mechanics*, *233*, 119–139.  
509 doi:10.1017/S0022112091000411.
- 510 Carton, J. A., Chepurin, G. A., Chen, L., & Grodsky, S. A. (2018). Improved Global Net  
511 Surface Heat Flux. *Journal of Geophysical Research: Oceans*, *123*, 3144–3163. doi:10.  
512 1002/2017JC013137.
- 513 Chelton, D. B., Gaube, P., Schlax, M. G., Early, J. J., & Samelson, R. M. (2011). The  
514 Influence of Nonlinear Mesoscale Eddies on Near-Surface Oceanic Chlorophyll. *Science*,  
515 *334*, 6054. doi:10.1126/science.1208897.
- 516 Corr ard, S. M., & Carton, X. (1999). Vertical alignment of geostrophic vortices. In *IUTAM*  
517 *Symposium on Simulation and Identification of Organized Structures in Flows* (pp. 191–  
518 200). Springer.

- 519 Cushman-Roisin, B., & Beckers, J. (2011). *Introduction to geophysical fluid dynamics: physical*  
520 *and numerical aspects*. Academic press.
- 521 de Marez, C., Carton, X., L'Hégaret, P., Meunier, T., Stegner, A., Le Vu, B., & Morvan,  
522 M. (2020). Oceanic vortex mergers are not isolated but influenced by the  $\beta$ -effect and  
523 surrounding eddies. *Scientific Reports*, *10*. doi:10.1038/s41598-020-59800-y.
- 524 Deremble, B., Dewar, W. K., & Chassignet, E. P. (2016). Vorticity dynamics near  
525 sharp topographic features. *Journal of Marine Research*, *74*, 249–276. doi:10.1357/  
526 002224016821744142.
- 527 Dewar, W. K., & Hogg, A. M. (2010). Topographic inviscid dissipation of balanced flow.  
528 *Ocean Modell.*, *32*, 1–13. doi:10.1016/j.ocemod.2009.03.007.
- 529 Dong, C., McWilliams, J. C., Liu, Y., & Chen, D. (2014). Global heat and salt transports  
530 by eddy movement. *Nature Communications*, *5*, 3294. doi:10.1038/ncomms4294.
- 531 Fer, I., Bosse, A., Ferron, B., & Bouruet-Aubertot, P. (2018). The Dissipation of Kinetic  
532 Energy in the Lofoten Basin Eddy. *Journal of Physical Oceanography*, *48*, 1299–1316.  
533 doi:10.1175/JPO-D-17-0244.1.
- 534 Garreau, P., Dumas, F., Louazel, S., Stegner, A., & Le Vu, B. (2018). High-Resolution Obser-  
535 vations and Tracking of a Dual-Core Anticyclonic Eddy in the Algerian Basin. *Journal of*  
536 *Geophysical Research: Oceans*, *123*, 9320–9339. URL: [http://doi.wiley.com/10.1029/  
537 2017JC013667](http://doi.wiley.com/10.1029/2017JC013667). doi:10.1029/2017JC013667.
- 538 Gelderloos, R., Katsman, C. A., & Drijfhout, S. S. (2011). Assessing the Roles of Three  
539 Eddy Types in Restratifying the Labrador Sea after Deep Convection. *Journal of Physical*  
540 *Oceanography*, *41*, 2102–2119. doi:10.1175/JPO-D-11-054.1.
- 541 Gula, J., Molemaker, J., & McWilliams, J. (2015). Gulf Stream dynamics along the south-  
542 eastern u.s seaboard. *Journal of Physical Oceanography*, *45*, 690–715. doi:10.1175/  
543 JPO-D-14-0154.1.

- 544 Gula, J., Molemaker, M. J., & McWilliams, J. C. (2016). Topographic generation of sub-  
545 mesoscale centrifugal instability and energy dissipation. *Nature Communications*, *7*.  
546 doi:10.1038/ncomms12811.
- 547 Gula, J., & Zeitlin, V. (2010). Instabilities of buoyancy-driven coastal currents and their  
548 nonlinear evolution in the two-layer rotating shallow-water model. Part 1. Passive lower  
549 layer. *J. Fluid Mech.*, *659*, 69–93. doi:10.1017/S0022112010002405.
- 550 Heywood, K. J., McDonagh, E. L., & White, M. A. (1994). Eddy kinetic energy of the  
551 North Atlantic subpolar gyre from satellite altimetry. *Journal of Geophysical Research*,  
552 *99*, 22525. doi:10.1029/94JC01740.
- 553 Hogg, A. M., Dewar, W. K., Berloff, P., & Ward, M. L. (2011). Kelvin wave hydraulic control  
554 induced by interactions between vortices and topography. *J. Fluid Mech.*, *687*, 194–208.  
555 doi:10.1017/jfm.2011.344.
- 556 Ioannou, A., Stegner, A., Le Vu, B., Taupier-Letage, I., & Speich, S. (2017). Dynamical Evo-  
557 lution of Intense Ierapetra Eddies on a 22 Year Long Period. *Journal of Geophysical Re-*  
558 *search: Oceans*, *122*, 9276–9298. URL: <http://doi.wiley.com/10.1002/2017JC013158>.  
559 doi:10.1002/2017JC013158.
- 560 Ivanov, Y., & Korablev, A. (1995). Formation and regeneration of the pycnocline lens in the  
561 norwegian sea. *Russian Meteorology and Hydrology*, (pp. 62–69).
- 562 Klein, P., Hua, B. L., Lapeyre, G., Capet, X., Le Gentil, S., & Sasaki, H. (2008). Upper  
563 ocean turbulence from high-resolution 3D simulations. *J. Phys. Oceanogr.*, *38*, 1748–1763.
- 564 Köhl, A. (2007). Generation and Stability of a Quasi-Permanent Vortex in the Lofoten Basin.  
565 *Journal of Physical Oceanography*, *37*, 2637–2651. doi:10.1175/2007JP03694.1.
- 566 LaCasce, J. H. (1998). A geostrophic vortex over a slope. *Journal of physical Oceanography*,  
567 *28*, 2362–2381.
- 568 Lam, J., & Dritschel, D. G. (2001). On the beta-drift of an initially circular vortex patch.  
569 *Journal of Fluid Mechanics*, *436*, 107–129. doi:10.1017/S0022112001003974.



- 570 Large, W. G., McWilliams, J. C., & Doney, S. C. (1994). Oceanic vertical mixing: A review  
571 and a model with a nonlocal boundary layer parameterization. *Reviews of Geophysics*,  
572 *32*, 363–403. doi:10.1029/94RG01872.
- 573 Le Corre, M., Gula, J., Smilenova, A., & Houpert, L. (2019). On the dynamics of a  
574 deep quasi-permanent anticyclonic eddy in the rockall trough. *Association Française de*  
575 *Mécanique, Brest, France*, (p. 12).
- 576 Le Corre, M., Gula, J., & Tréguier, A.-M. (2020). Barotropic vorticity balance of the  
577 North Atlantic subpolar gyre in an eddy-resolving model. *Ocean Science*, *16*, 451–468.  
578 doi:10.5194/os-16-451-2020.
- 579 Le Vu, B., Stegner, A., & Arsouze, T. (2018). Angular Momentum Eddy Detection and  
580 Tracking Algorithm (AMEDA) and Its Application to Coastal Eddy Formation. *Journal*  
581 *of Atmospheric and Oceanic Technology*, *35*, 739–762. doi:10.1175/JTECH-D-17-0010.1.
- 582 Legg, S., McWilliams, J., & Gao, J. (1998). Localization of deep ocean convection by a  
583 mesoscale eddy. *Journal of Physical Oceanography*, *28*, 944–970.
- 584 Legg, S., & McWilliams, J. C. (2001). Convective modifications of a geostrophic eddy field.  
585 *Journal of physical oceanography*, *31*, 874–891.
- 586 Lilly, J. M., Rhines, P. B., Schott, F., Lavender, K., Lazier, J., Send, U., & D’Asaro, E.  
587 (2003). Observations of the Labrador Sea eddy field. *Progress in Oceanography*, *59*,  
588 75–176. doi:10.1016/j.pocean.2003.08.013.
- 589 de Marez, C., L’Hégaret, P., Morvan, M., & Carton, X. (2019). On the 3D structure of  
590 eddies in the Arabian Sea. *Deep Sea Research Part I: Oceanographic Research Papers*, .  
591 URL: <https://linkinghub.elsevier.com/retrieve/pii/S0967063718303650>. doi:10.  
592 1016/j.dsr.2019.06.003.
- 593 de Marez, C., Meunier, T., Morvan, M., L’Hégaret, P., & Carton, X. (2020a). Study of the  
594 stability of a large realistic cyclonic eddy. *Ocean Modelling*, *146*, 101540. doi:10.1016/j.  
595 ocemod.2019.101540.

- 596 de Marez, C., Meunier, T., Tedesco, P., L'Hégaret, P., & Carton, X. (2020b). Vortex-wall  
597 interaction on the  $\sigma$ -plane and the generation of deep submesoscale cyclones by internal  
598 Kelvin Waves-current interactions. *Geophysical & Astrophysical Fluid Dynamics*, *114*,  
599 588–606. doi:10.1080/03091929.2020.1772779.
- 600 McWilliams, J. C. (1985). Submesoscale, coherent vortices in the ocean. *Reviews of Geo-*  
601 *physics*, *23*, 165. doi:10.1029/RG023i002p00165.
- 602 Ménesguen, C., Le Gentil, S., Marchesiello, P., & Ducouso, N. (2018). Destabilization of  
603 an oceanic meddy-like vortex: energy transfers and significance of numerical settings. *J.*  
604 *Phys. Oceanogr.*, *48*, 1151–1168. doi:10.1175/jpo-d-17-0126.1.
- 605 Mkhinini, N., Coimbra, A. L. S., Stegner, A., Arsouze, T., Taupier-Letage, I., & Béranger,  
606 K. (2014). Long-lived mesoscale eddies in the eastern Mediterranean Sea: Analysis of  
607 20 years of AVISO geostrophic velocities. *Journal of Geophysical Research: Oceans*,  
608 *119*, 8603–8626. URL: <http://doi.wiley.com/10.1002/2014JC010176>. doi:10.1002/  
609 2014JC010176.
- 610 Nof, D., & Dewar, W. (1994). Alignment of lenses: laboratory and numerical experiments.  
611 *Deep Sea Research Part I: Oceanographic Research Papers*, *41*, 1207–1229. doi:10.1016/  
612 0967-0637(94)90041-8.
- 613 Perrot, X., & Carton, X. (2010). 2D vortex interaction in a non-uniform flow. *Theoretical*  
614 *and Computational Fluid Dynamics*, *24*, 95–100. doi:10.1007/s00162-009-0127-4.
- 615 Rodríguez-Marroyo, R., Viúdez, A., & Ruiz, S. (2011). Vortex Merger in Oceanic Tripoles.  
616 *Journal of Physical Oceanography*, *41*, 1239–1251. doi:10.1175/2011JP04582.1.
- 617 Roulet, G., Capet, X., & Maze, G. (2014). Global interior eddy available potential energy  
618 diagnosed from Argo floats. *Geophysical Research Letters*, *41*, 1651–1656. doi:10.1002/  
619 2013GL059004.
- 620 Schubert, R., Gula, J., Greatbatch, R. J., Baschek, B., & Biastoch, A. (2020). The Sub-  
621 mesoscale Kinetic Energy Cascade: Mesoscale Absorption of Submesoscale Mixed Layer

- 622 Eddies and Frontal Downscale Fluxes. *Journal of Physical Oceanography*, 50, 2573–2589.  
623 URL: <https://doi.org/10.1175/JPO-D-19-0311.1>. doi:10.1175/JPO-D-19-0311.1.  
624 arXiv:<https://journals.ametsoc.org/jpo/article-pdf/50/9/2573/4992078/jpod190311.pdf>.
- 625 Shchepetkin, A. F., & McWilliams, J. C. (2005). The regional  
626 oceanic modeling system (ROMS): a split-explicit, free-surface,  
627 topography-following-coordinate oceanic model. *Ocean Modell.*, 9, 347–404.
- 628 Shchepetkin, A. F., & McWilliams, J. C. (2011). Accurate Boussinesq oceanic  
629 modeling with a practical, "stiffened" equation of state. *Ocean Modell.*,  
630 38, 41–70. doi:10.1016/j.ocemod.2011.01.010.
- 631 Smilenova, A., Gula, J., Le Corre, M., Houpert, L., & Reecht, Y. (2020).  
632 A Persistent Deep Anticyclonic Vortex in the Rockall Trough Sustained  
633 by Anticyclonic Vortices Shed From the Slope Current and Wintertime  
634 Convection. *Journal of Geophysical Research: Oceans*, 125. doi:10.1029/  
635 2019JC015905.
- 636 Solodoch, A., Stewart, A. L., & McWilliams, J. C. (2021). Formation of  
637 Anticyclones above Topographic Depressions. *Journal of Physical Oceanography*,  
638 51, 207–228. doi:10.1175/JPO-D-20-0150.1.
- 639 Sutyrin, G. G. (2019). On vortex intensification due to stretching out of  
640 weak satellites. *Physics of Fluids*, 31, 075103. doi:10.1063/1.5098068.
- 641 Sutyrin, G. G., & Radko, T. (2019). On the peripheral intensification of  
642 two-dimensional vortices in smaller-scale randomly forcing flow. *Physics of*  
643 *Fluids*, 31, 101701. doi:10.1063/1.5118752.
- 644 Trodahl, M., Isachsen, P. E., Lilly, J. M., Nilsson, J., & Kristensen,  
645 N. M. (2020). The Regeneration of the Lofoten Vortex through Vertical  
646 Alignment. *Journal of Physical Oceanography*, 50, 2689–2711. URL:  
647 <https://journals.ametsoc.org/view/journals/phoc/50/9/jpoD200029.xml>.  
648 doi:10.1175/JPO-D-20-0029.1.

- 649 Verron, J., Hopfinger, E. J., & McWilliams, J. C. (1990). Sensitivity  
650 to initial conditions in the merging of two-layer baroclinic vortices.  
651 *Physics of Fluids A: Fluid Dynamics*, 2, 886--889. doi:10.1063/1.857647.
- 652 Verron, J., & Valcke, S. (1994). Scale-dependent merging of baroclinic  
653 vortices. *Journal of Fluid Mechanics*, 264, 81--106. doi:10.1017/  
654 S0022112094000595.
- 655 Volkov, D. L. (2005). Interannual Variability of the Altimetry-Derived  
656 Eddy Field and Surface Circulation in the Extratropical North Atlantic  
657 Ocean in 1993{2001. *Journal of Physical Oceanography*, 35, 405--426.  
658 doi:10.1175/JP02683.1.
- 659 White, M. A., & Heywood, K. J. (1995). Seasonal and interannual changes in  
660 the North Atlantic subpolar gyre from Geosat and TOPEX/POSEIDON altimetry.  
661 *Journal of Geophysical Research*, 100, 24931. doi:10.1029/95JC02123.
- 662 Xu, W., Miller, P. I., Quartly, G. D., & Pingree, R. D. (2015). Seasonality  
663 and interannual variability of the European Slope Current from 20years  
664 of altimeter data compared with in situ measurements. *Remote Sensing of  
665 Environment*, 162, 196--207. doi:10.1016/j.rse.2015.02.008.
- 666 Yu, L., Bosse, A., Fer, I., Orvik, K. A., Bruvik, E. M., Hessevik, I., &  
667 Kvalsund, K. (2017). The Lofoten Basin eddy: Three years of evolution  
668 as observed by Seagliders: THE LOFOTEN BASIN EDDY. *Journal of Geophysical  
669 Research: Oceans*, 122, 6814--6834. doi:10.1002/2017JC012982.
- 670 Zhang, Z., Wang, W., & Qiu, B. (2014). Oceanic mass transport by mesoscale  
671 eddies. *Science*, 345, 322--324. doi:10.1126/science.1252418.

Highlights for 'The influence of merger and convection on an anticyclonic eddy trapped in a bowl':

- Anticyclonic eddies lying in a topographic depression live for several years
- Small SCVs generated nearby the anticyclonic eddy intensify the latter at their depth of generation
- Wintertime convection is efficient in intensifying the anticyclonic eddy
- The sustaining for several years of such eddies is due to the balance between merger, convection, and bottom drag

**Charly de Marez:** Conceptualization, Methodology, Software, Writing- Original draft.

**Mathieu Le Corre:** Software, Writing- Reviewing and Editing.

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*Journal Pre-proof*

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