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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.

Keywords: mesoscale; vortex; merger; convection

The authors declare no competing interests.

1. Introduction

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

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

The RT eddy is located in the Rockall Trough, off Ireland in the North Atlantic. It has a clear signature at the sea surface (Heywood et al., 1994; White & Heywood, 1995; Volkov, 2005; Xu et al., 2015), but also at depth with high values of eddy available potential energy (Roulet et al., 2014). This eddy is less sampled than the LV and less known. However, 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.

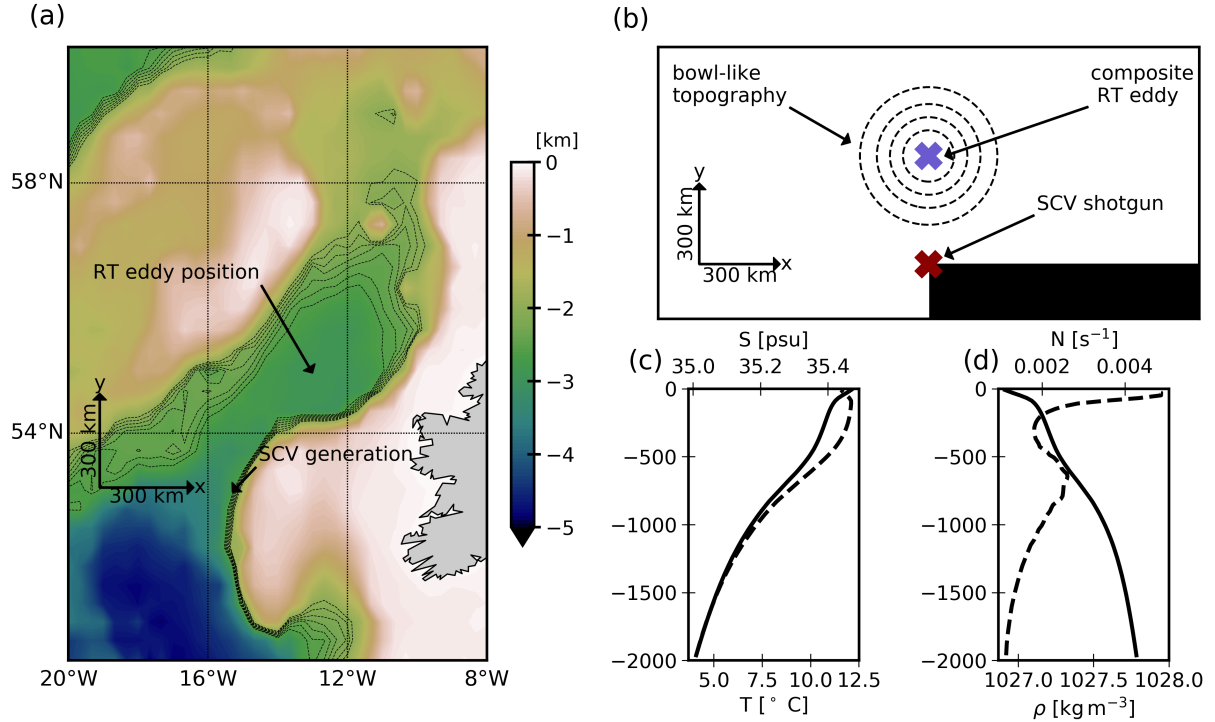


Figure 1: a) Rockall Trough bathymetry; dashed contours show isobaths 2500 to 2000 m depth with a 100 m interval. b) Scheme of idealized simulation setup; dashed contours show same isobaths as in a). c) Climatological background temperature (solid) and salinity (dashed) used in idealized simulations.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

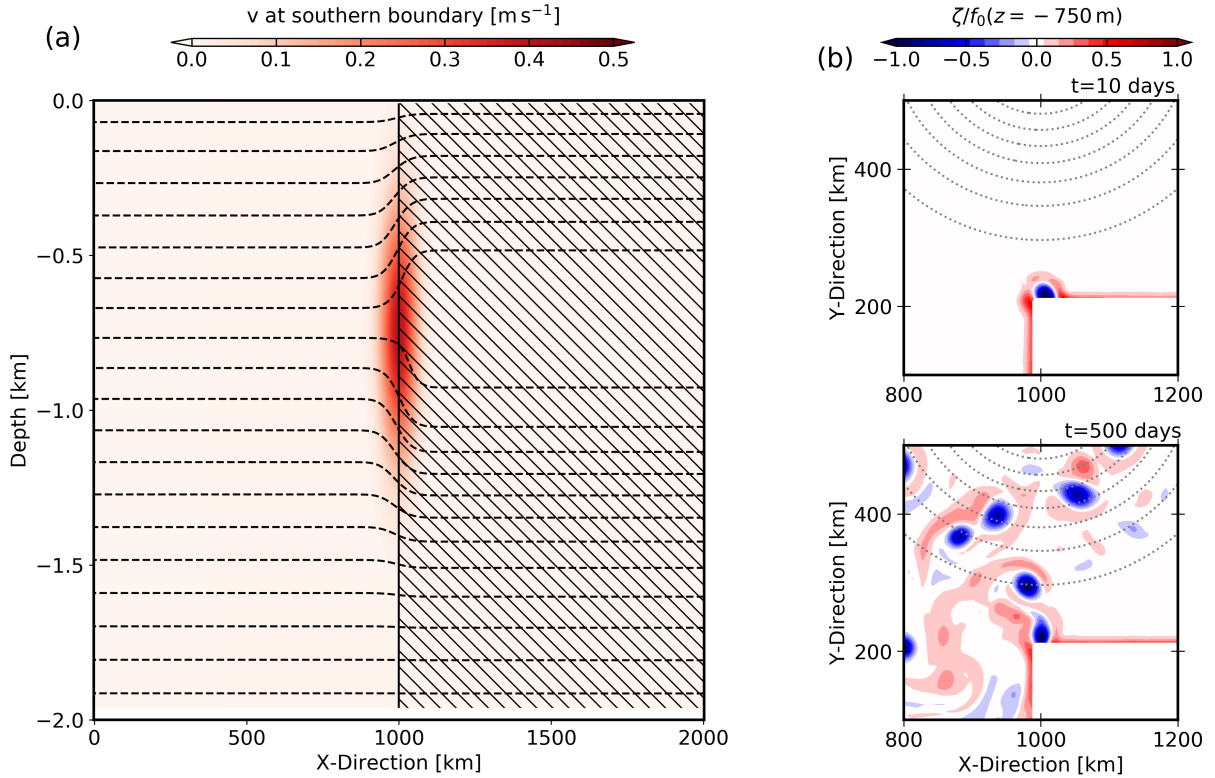


Figure 2: a) Meridional velocity at the southern boundary for the "middle" case; black lines indicate isopycnals with a 0.5 kg m^{-3} spacing; hatched area indicate the position of the mask. b) Snapshots of normalized relative vorticity at $t=10$ and $t=500$ days, at 750 m depth, showing the SCV generation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

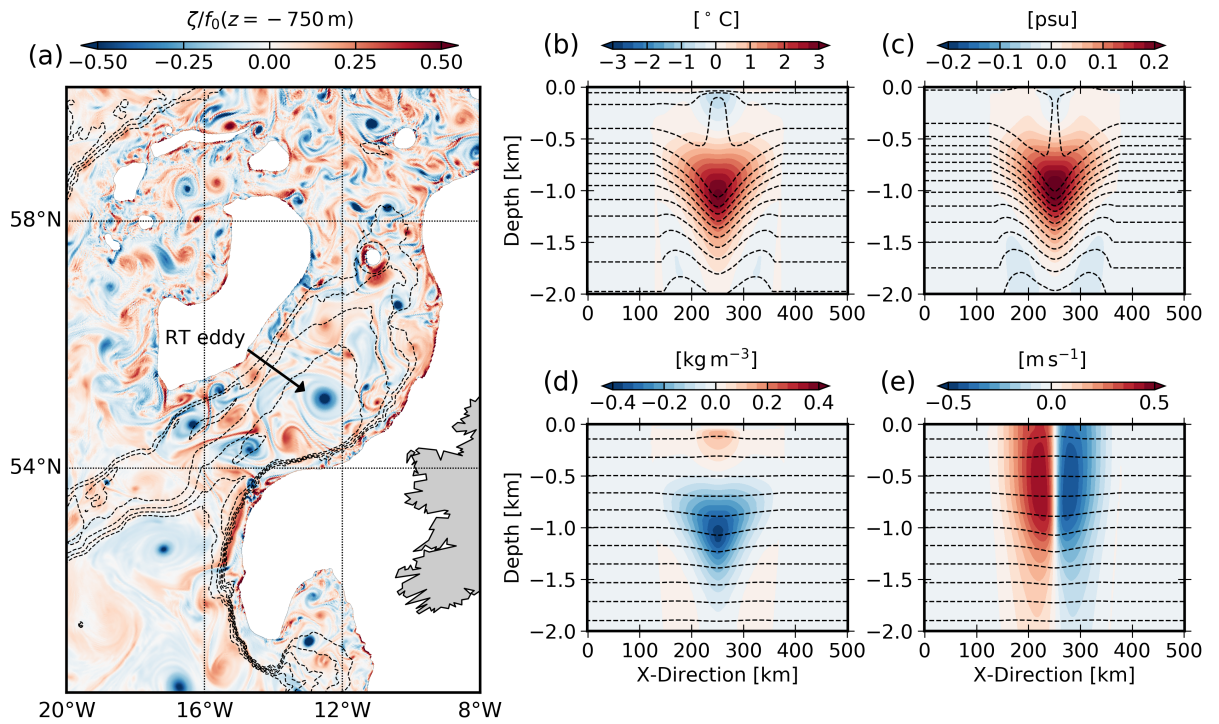


Figure 3: a) Snapshot of normalized relative vorticity at 750 m depth in the Rockall Trough area, from the realistic simulation (Le Corre et al., 2020) in which the composite anticyclone was extracted; dashed contours show isobaths from 3500 to 2000 m depth with a 250 m interval. (b,c,d,e) Temperature 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°N and
134 57°N , and 15°W and 11°W . The annual variation of SNHF imposed in the simulations with
135 convection is shown in Fig. 4.

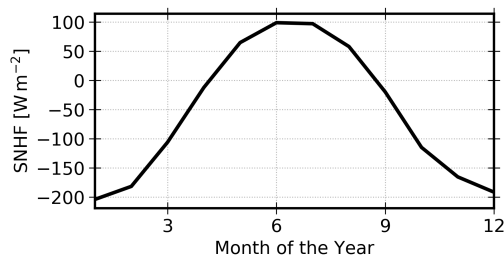


Figure 4: Surface net heat flux imposed in idealized simulations with convection.

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:

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

175 with summation convention, $i = 1, 2$, and $j = 1, 2$, u_i are the horizontal component of
 176 velocity, ∂_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 • $\text{hadv} = \int dz u_j \partial_j (\frac{1}{2}u_i^2)$,
- 181 • $\text{vadv} = \int dz w \partial_z (\frac{1}{2}u_i^2)$,
- 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 $\int_0^t dt$ 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.

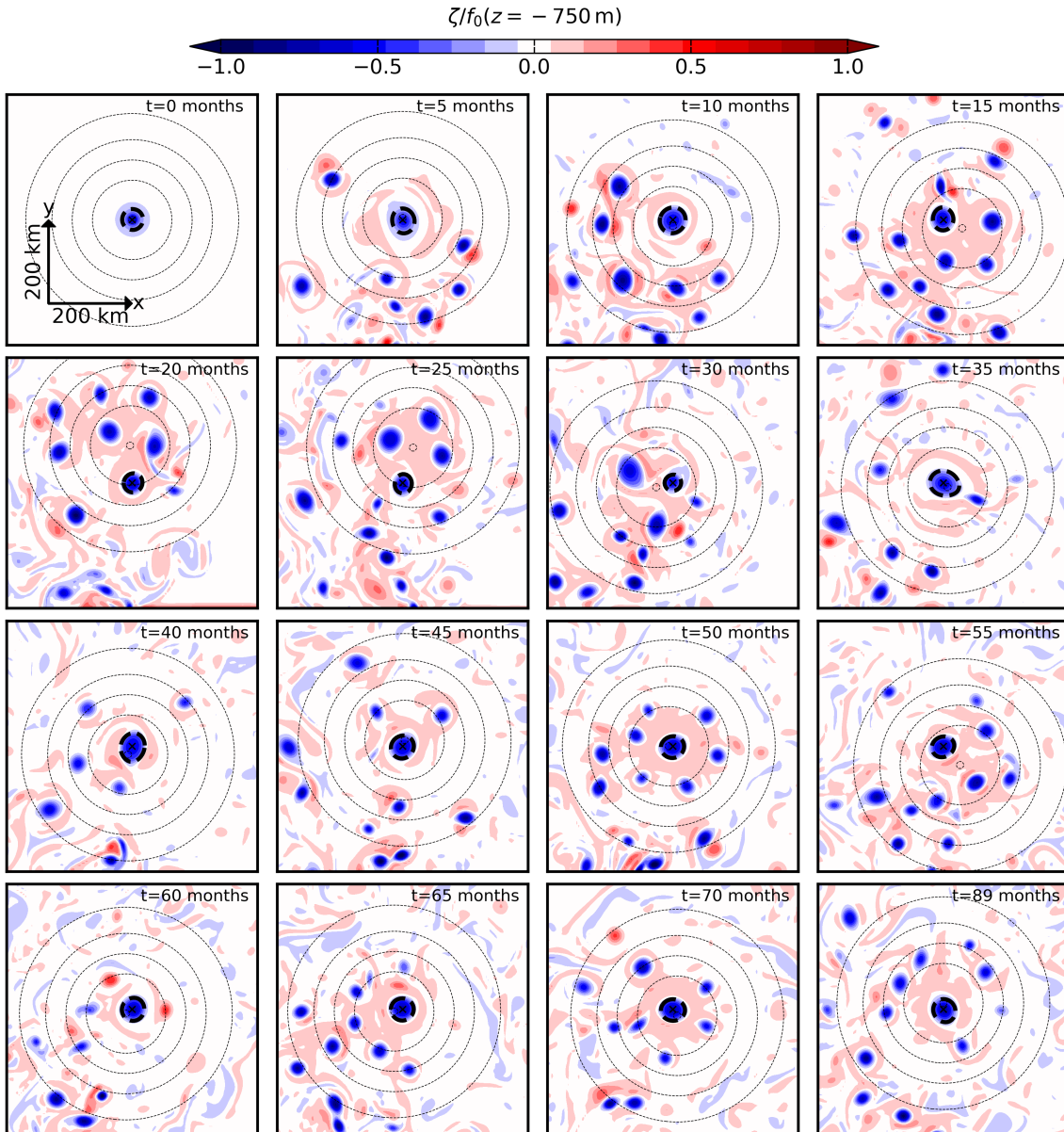


Figure 5: Snapshots of normalized relative vorticity at 750 m depth, in the mRTD simulation. Each panel is 600 km large, and is centered around the main eddy. Bold dashed contours indicate the contour of maximum velocity of the main eddy. Thin dashed contours show isobath from 2500 to 2000 m depth with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

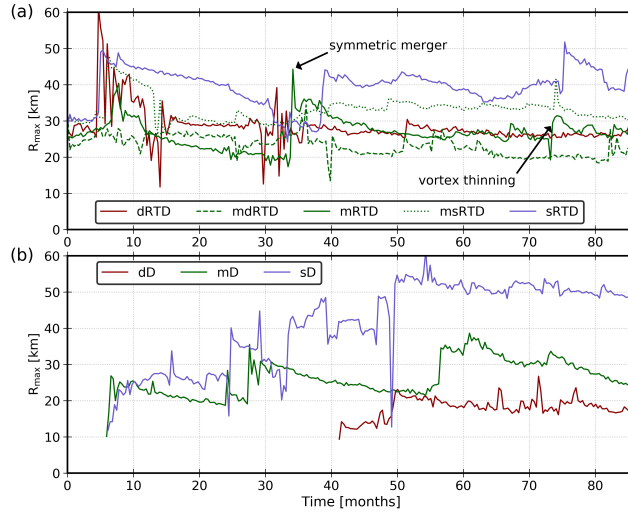


Figure 6: Evolution of R_{max} during simulations, for different initialization depth of the SCV shotgun (a) with and (b) without the RT composite at the center of the bowl. All simulations include bottom drag parameterization.

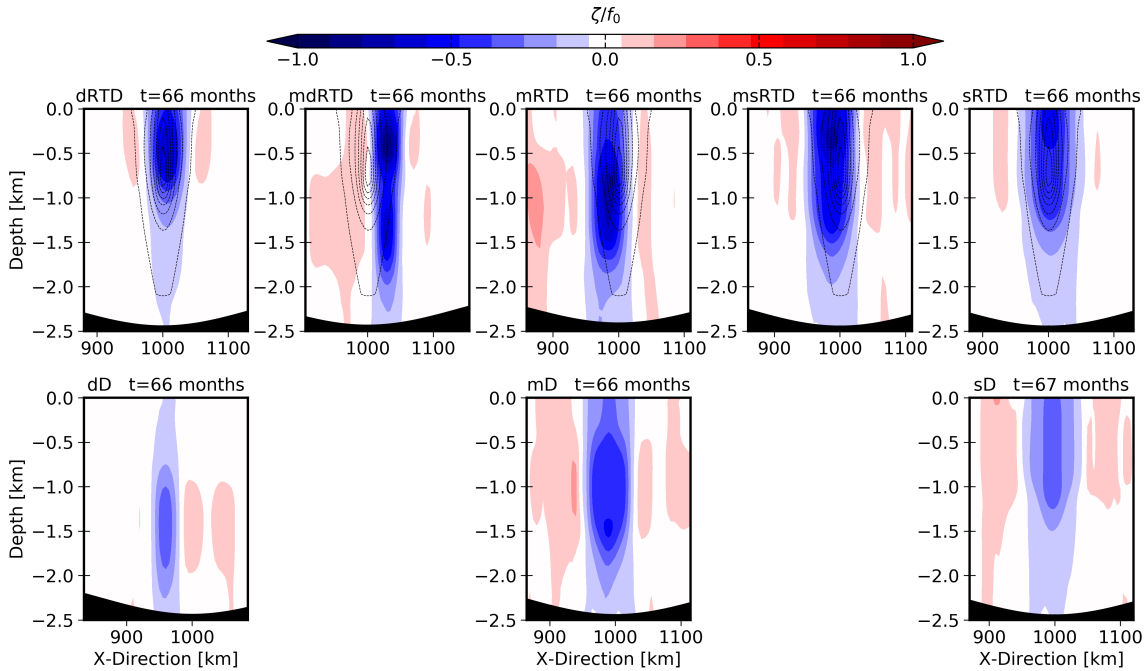


Figure 7: Vertical sections of normalized relative vorticity passing through the center of the main eddy after ~ 5 and a half years of simulation, for simulations shown in Fig. 6. Thin contours in the top row show the contours of normalized vorticity through the center of the RT composite at initialization; 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 ~ 500 m depth and ~ 1300 m depth.

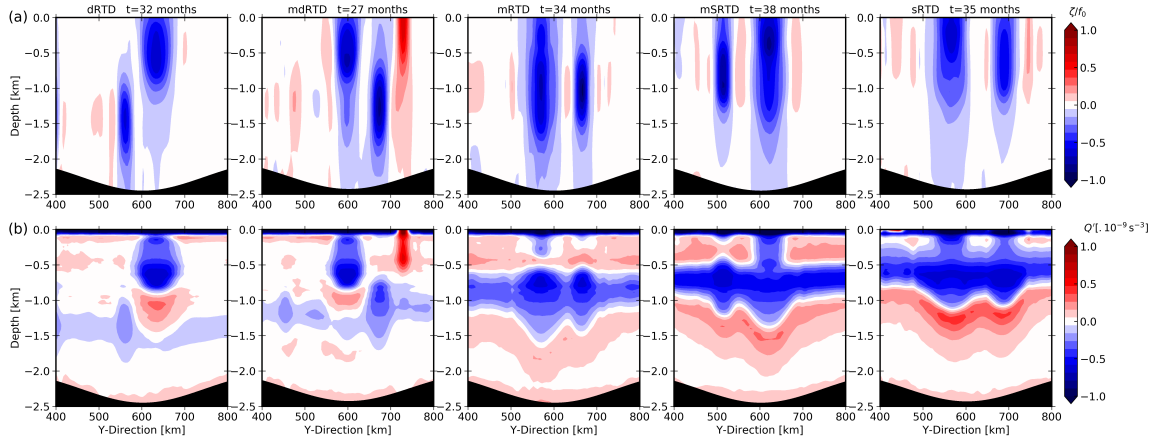


Figure 8: Vertical section of (a) normalized relative vorticity and (b) associated PV anomaly, at times just before a merger of the main eddy with a companion eddy, in dRTD, mdRTD, mRTD, msRTD, and sRTD simulations. All sections pass through the center of both eddies.

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); Corréard & 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. Corréard & 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 ~ 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 ~ 750 m depth.

325 3.3. On the importance of convection

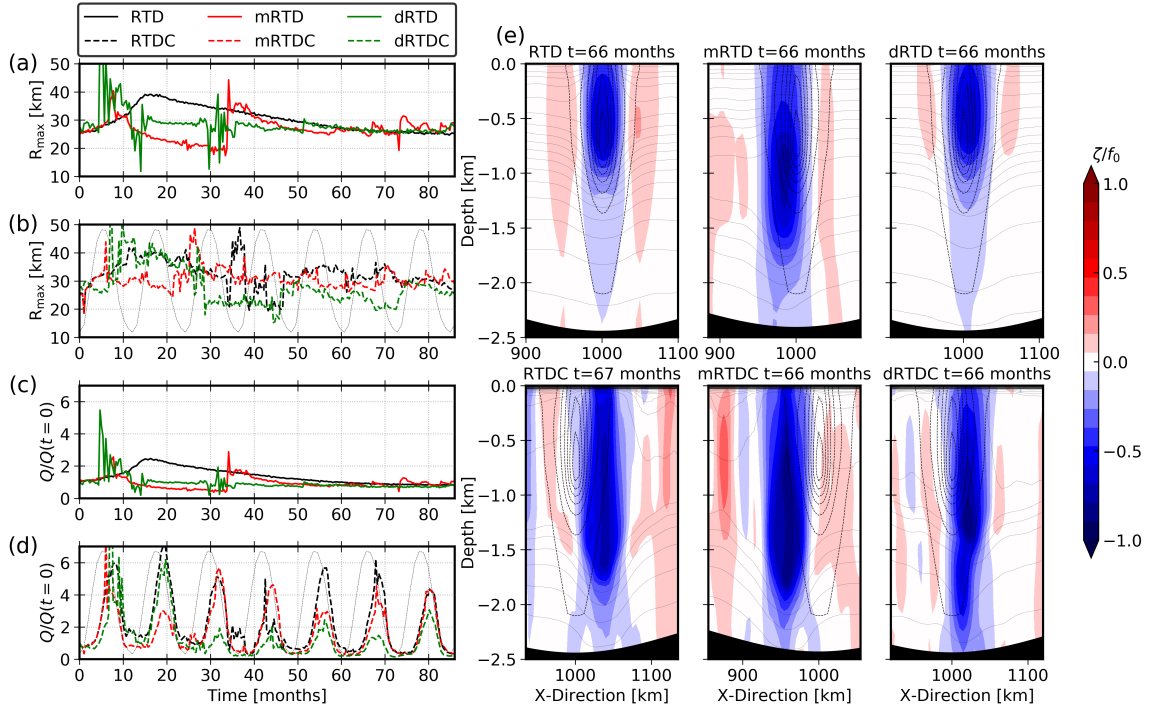


Figure 9: (a,b) (resp. (c,d)) Time evolution of R_{\max} (resp. ratio between PV and initial PV integrated over the main eddy) for some simulations without (a,c, solid line) or with (b,d, dashed lines) convection; the thin gray line shows the time evolution of the SNHF applied at the surface in the simulations 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 ζ 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 ~ 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.

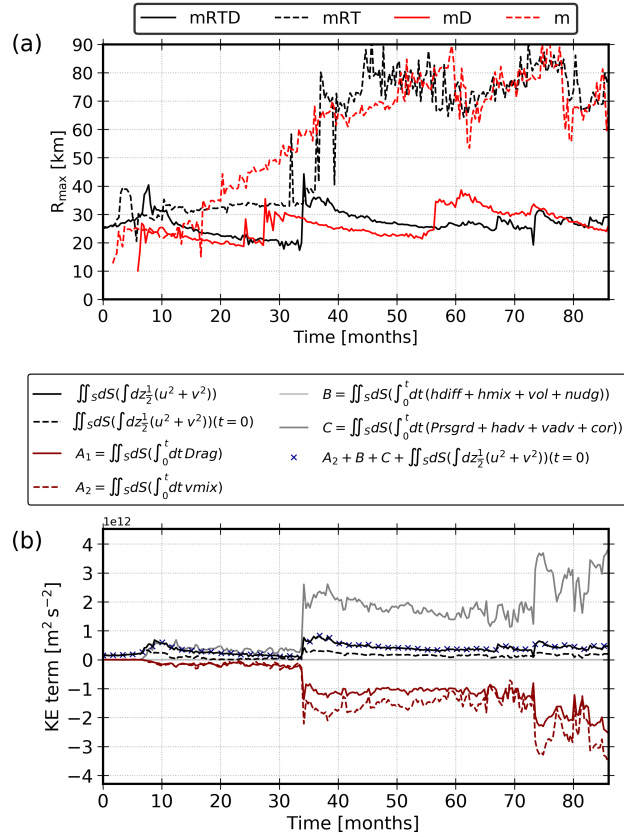


Figure 10: (a) Evolution of R_{max} during simulations, with (solid) and without (dashed) bottom drag parameterization. (b) Evolution of kinetic energy terms in the mRTD simulation. Each term is integrated in time and in the contour of the main eddy. Note that the superposition of blue crosses with the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

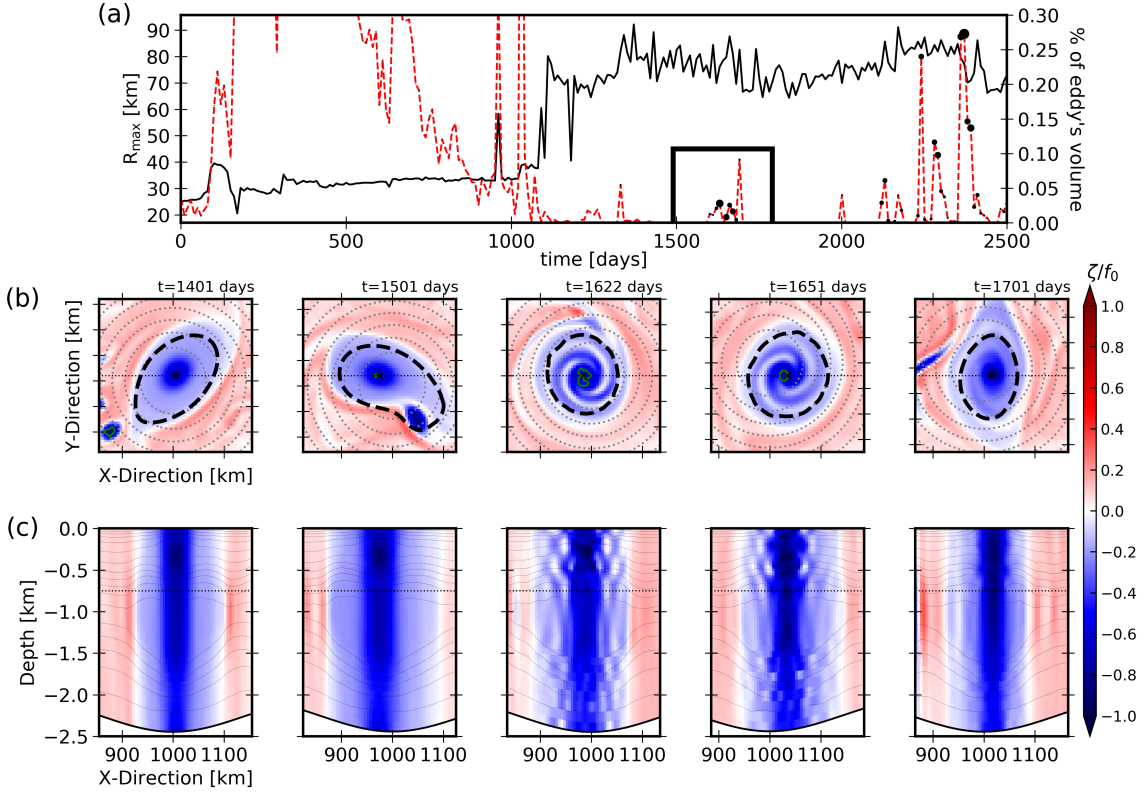


Figure 11: (a) Time evolution of R_{\max} (black line) and % of eddy's volume with negative PV (dashed red) for the mRT simulation. The size of black dots indicates the amplitude (in absolute value) of the minimum PV in the eddy's contour. (b) Horizontal sections of normalized relative vorticity at 750 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 preferly 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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