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# Influence of bottom topography on an upwelling current : generation of long trapped filaments

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## Abstract

We investigate the influence of bottom topography on the formation and trapping of long upwelling filaments. They use a 2-layer shallow water model on the f-plane. A wind forced along-shore current, associated with coastal upwelling along a vertical wall, encounters a promontory of finite width and length, perpendicular to the coast.

In the lower layer, topographic eddies form, which are shown to drive the formation of a filament on the front. Indeed, as the upwelling current and front develop along the coast, the along shore flow crosses the promontory, re-arranging the potential vorticity structure and generating intense vortical structures : water columns with high potential vorticity initially localized upon the promontory are advected into the deep ocean, forming cyclonic eddies, while water columns from the deep ocean with low potential vorticity climb on the topography forming a trapped anti-cyclonic circulation. These topographic eddies interact with the upper layer upwelling front and form an elongated, trapped and narrow filament.

Sensitivity tests are then carried out and it is shown that :

- baroclinic instability of the front does not play a major role on the formation of long trapped filaments;
- increasing the duration of the wind forcing increases the upwelling current and limits the offshore growth of the filament;
- modifying the promontory characteristics (width, length, height and slopes) has strong impact on the filament evolution, sometimes leading to a multipolarisation of the potential vorticity anomaly structure which results in much more complicated patterns in the upper layer (numerous shorter and less coherent filaments). This shows that only specific promontory shapes can lead to the formation of well defined filaments;
- adding bottom friction introduces a slight generation of potential vorticity in the bottom layer over the promontory, but does not significantly alter significantly the formation of the filament along the outcropped front in the present configuration;
- modifying the stratification characteristics, in particular the density jump between the layers, has only a weak influence on the dynamics of topographic eddies and on filament formation;

- the influence of capes is also modest in our simulations, showing that topography plays the major role in the formation of long and trapped upwelling filaments.

*Key words:* Eastern boundary, Mesoscale dynamics, Upwelling/Downwelling, Topographic flows, Upwelling filaments, Potential vorticity

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## 1 Introduction

Long trapped filaments of cold water are ubiquitous features along upwelling fronts. They sometimes extend hundreds of kilometers offshore and have been shown to play a major role in the offshore transport of recently upwelled coastal water ([Kostianoy and Zatsepin, 1996]) and in feeding the oligotrophic offshore waters with nutrients and organic materials ([Alvarez-Salgado *et al.*, 2007] and more particularly over the Iberian peninsula upwelling [Alvarez-Salgado *et al.*, 2001]). They are thus important physical features for eastern boundary upwelling ecosystems.

Figure 1 shows a set of satellite images of the Iberian peninsula on the 09/05/2009 (column 1) and 09/05/2005 (column 2).. The images on the first row are the Sea Surface Temperature (SST) and on the second row the Chlorophyll-a concentration (Sea Surface Color). Long, trapped and recurrent filaments are observed on SST and chlorophyll maps off Cape Finisterre (43°N), São Vicente (37°N) and the Estremadura promontory (between 38.5 and 39.5°N). However on all maps, another tongue of cold upwelled water pointing offshore also clearly arises from the upwelling front just South of the Estremadura promontory.

In-situ and remotely sensed data have provided some descriptions of upwelling filaments (see for instance [Brink, 1983]; [Flament *et al.*, 1985]; [Washburn and Armi, 1988]; [Strub *et al.*, 1991]; [Navarro-Pérez and Barton, 1998]; [Barton, 2001]), and many numerical and theoretical studies have focused on their dynamics. [Haynes *et al.*, 1993] used satellite data of the West-Iberia upwelling system to show that large filaments were often closely related with capes, but noted the repeated occurrence of two large filaments at two different locations corresponding with a straight coastline. They assumed that those filaments were related with unstable meanders of the jet, but their stationarity remained unexplained.

One of the first process studies on upwelling filament dynamics was performed by [Haidvogel *et al.*, 1991]. Their model was able to reproduce cold filaments along the upwelling front. They concluded from a set of sensitivity cases, including removal of the coastline geometry and the bottom topography, that the presence of large capes along the coast, was necessary to the generation of upwelling filaments. [Marchesiello *et al.*, 2003] studied the equilibrium structure of the California current system and conducted different sensitivity tests. They showed that mesoscale variability was intrinsic to the current and not due to the variability of the forcing. Contrary to [Haidvogel *et al.*, 1991], they showed that even in the case of a straight coastline and a flat bottom, upwelling filaments and eddies still occurred, but with no preferential location, confirming the conclusions of [Roed and Shi, 1999], that instability induced filaments and eddies could happen in the absence of coastal irregularities. Removing only the topography,

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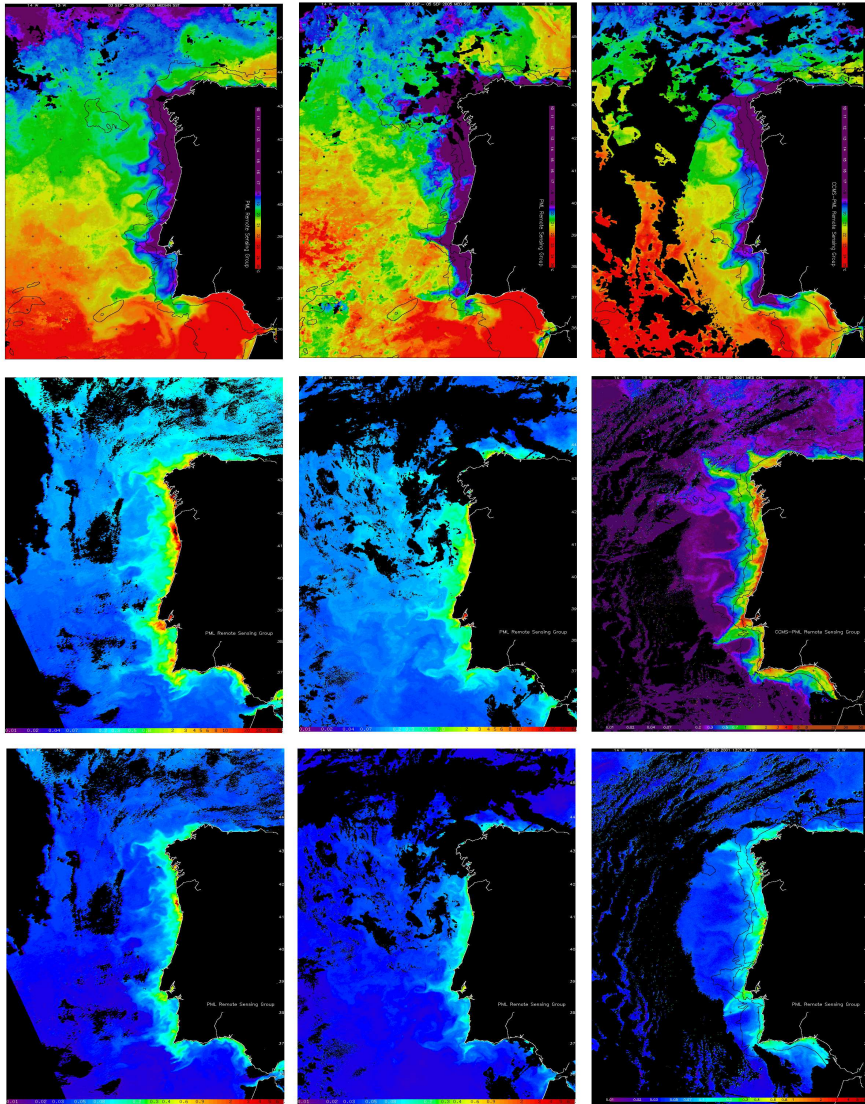


Figure 1: Sea Surface Temperature (first column) and Chlorophyll distribution (second column) situation off the West Iberian coast during three distinct upwelling episodes (04/09/2005 on the first row and 05/09/2009 on the second row.)

35 [Marchesiello *et al.*, 2003] concluded that a standing wave could be generated at the cape, in  
36 agreement with results from a numerical study of [Batteen, 1997] who noted anchoring of  
37 filaments at coastline irregularities, while [Roed and Shi, 1999] only noted a weak anchoring of  
38 the meander. [Ikeda, 1981] and [Ikeda, 1989] showed using two-layer quasigeostrophic models,  
39 that unstable meanders of a buoyancy driven coastal jet would move downstream by a combined  
40 effect of propagation and advection, so that some physical process had to be involved in the  
41 anchoring of filaments at the capes as observed by [Marchesiello *et al.*, 2003].

42 Recently [Batteen *et al.*, 2007] conducted a process oriented study of the Northern Canary  
43 Current System (NCCS) and revisited the upwelling instability and filamentation problem in this  
44 region. They modified the bottom topography, boundary conditions and wind forcing and found  
45 that :

- 46 • the flat bottom experiment shows many of the typically observed features of the NCCS;
- 47 • adding the bottom topography shows that topography has an important role in intensifying  
48 and trapping the equatorward current near the coast, in weakening the subsurface poleward  
49 current, and in intensifying eddies off the capes of Iberia;
- 50 • the flat bottom experiment produces anticyclonic meanders near cabo Roca and Cabo São  
51 Vincente, but not off Figueira da Foz;
- 52 • the beta effect plays an important role on the formation of the subsurface meander off cabo  
53 Roca.

54 To summarize, four main source have been identified to explain the generation of filaments  
55 along upwelling fronts :

- 56 • the frontal or baroclinic instability of the front;
- 57 • the effect of capes;
- 58 • the planetary beta effect;
- 59 • bottom topography.

60 Upwelling filaments have also been classified into different types (see [Strub *et al.*, 1991]), but  
61 one important distinction is whether they are trapped or not. Because the trapping of filaments  
62 always happens at the same locations, we believe that the observed long trapped upwelling fil-  
63 aments are associated with topographic features and we chose to focus on this aspect in the  
64 present study. Again, different studies ([Ikeda, 1989, Capet and Carton, 2004]) concluded that  
65 topographic irregularities were destabilization source for upwelling fronts, but the effect of bot-  
66 tom topography on the development of trapped filaments has not received a lot of attention, in  
67 particular the details of the mechanism is not clear and its sensitivity to different parameters  
68 remains to be studied.

69 [Stern and Chassignet, 2000] showed, using both a  $1_{1/2}$  and a three-layer isopycnic model,  
70 that intrinsic instability was not sufficient to generate detrainment of fluid and eddy-separation  
71 from the jet. They concluded that, to generate a blocking wave and detrain water, there was  
72 a need for a downstream variation in jet transport, and noted that this variation could happen  
73 in the case of alongshore varying topography, but did not investigate further on this point.  
74 [Viera and Grimshaw, 1994] studied the evolution of a potential vorticity front over an iso-  
75 lated topography, using a  $1_{1/2}$  layers quasi-geostrophic model, and showed, that a linearly

76 stable jet associated with a potential vorticity front could produce large and pinched off me-  
 77 anders when interacting with bottom topography. [Herbette et al., 2003] have shown that a  
 78 seamount could interact with a surface intensified eddy and generate filaments (or even split  
 79 the eddy). Finally, while studying the generation of secondary upwelling fronts along continen-  
 80 tal slopes [Rossi et al., 2009] found, in one of the experiment with a promontory (see fig. 22 in  
 81 [Rossi et al., 2009]), that a bottom topography could trap upwelled waters and even observed the  
 82 formation of a trapped filament extending offshore.

83 Therefore, we study the evolution of an upwelling front in the presence of an along shore  
 84 varying topography, in the form of cross-shore coastal promontories. We focus on the formation  
 85 and trapping of long filaments extending offshore and we base our approach on the potential  
 86 vorticity analysis used in the papers quoted in the previous paragraph.

87 The outline is :

- 88 • in section 2 (Model and tools) we describe the numerical model and recall some basic princi-  
 89 ples relating potential vorticity (thereafter *PV*) and potential vorticity anomalies (thereafter  
 90 *PVA*) to the dynamics;
- 91 • in section 3 (Reference experiment) we present a first simulation that illustrates the devel-  
 92 opment of a long filament. In particular we describe how the generation of *PVA* by the  
 93 displacement of water columns above the promontory can generate a permanent filament  
 94 trapped downstream of the promontory;
- 95 • in section 4 (Sensitivity study) we study the sensitivity of this mechanism to different  
 96 regimes and parameters : stable and unstable cases, wind forcing duration, promontory  
 97 height, width and length, stratification, bottom friction;
- 98 • in section 5 (Conclusion) we sum up and discuss our results.

## 99 2. Model and tools

### 100 2.1. Equations and model

101 The model used is an adiabatic version of MICOM (Miami Isopycnic Coordinate Ocean  
 102 Model) ([Bleck and Boudra, 1986]; [Bleck and Smith, 1990];[Bleck et al., 1992]) modified to  
 103 include a fourth order scheme in the non-linear advection terms and a biharmonic diffu-  
 104 sion operator to improve the PV dynamics ([Winther et al., 2007, Morel and McWilliams, 2001,  
 105 Herbette et al., 2003]). This model solves the shallow water equations which, for the two-layer  
 106 configurations considered here, can be expressed as :

$$\partial_t \mathbf{u}_k + (\mathbf{u}_k \cdot \nabla) \mathbf{u}_k + f_0 \mathbf{k} \times \mathbf{u}_k = -\nabla \mathcal{M}_k + F_k + T_k^w, \quad (1)$$

$$\partial_t h_k + \nabla \cdot (\mathbf{u}_k h_k) = 0, \quad (2)$$

107 where  $k$  is the layer number (here,  $k = 1$  for the top layer and  $k = 2$  for the bottom one),  
 108  $\mathbf{u}_k = (u_k, v_k)$  is the horizontal velocity,  $f_0 = 10^{-4} s^{-1}$  is the Coriolis parameter (considered  
 109 constant here),  $h_k$  is the thickness of the isopycnal layer  $k$ ,  $T_k^w$  represents the wind forcing, and  
 110  $F_k$  contains the frictional and viscosity terms (horizontal diffusion is associated with a bihar-  
 111 monic operator with a viscosity that depends on the velocity modulus and deformation tensor,

112 see [Winther *et al.*, 2007]). Finally,  $\mathcal{M}_k$  is the Montgomery potential (pressure along an isopyc-  
 113 nal surface), which can be written :

$$\mathcal{M}_k = \sum_{i=1}^2 g h_i + \sum_{i=1}^{k-1} \frac{\rho_i - \rho_k}{\rho_k} g h_i, \quad (3)$$

114 where  $\rho_i$  is the density of the isopycnic layer  $i$  and  $g$  is the gravity acceleration.

## 115 2.2. Configuration and parameters

116 The configuration is a periodic zonal channel on an  $f$ -plane, with vertical side walls on the  
 117 northern and southern boundaries. The bottom is flat except near the southern boundaries in the  
 118 middle of the domain where there exists a promontory. As shown in figure 2, the promontory is  
 119 composed of a flat plateau of variable height  $H_t$ , length  $L_y$  and width  $L_x$ , rounded at its offshore  
 120 edge, and surrounded by a Gaussian slope of a variable extension  $dL$  (figure 2). To represent a  
 121 mid latitude summer thermocline, the surface layer depth at rest  $H_1 = 50$  m and the bottom layer  
 122 depth away from the promontory is  $H_2^\infty = 2000$  m for most experiments. The upper layer density  
 123 is fixed to  $\rho_1 = 1000 \text{ kg/m}^3$ , the water column stratification is defined by the reduced gravity  
 124  $g' = g(\rho_2 - \rho_1)/\rho_1$ . Unless stated otherwise (when testing the sensitivity to the stratification  
 125 characteristics)  $g' \simeq 0.01$ .

126  $R_d = \sqrt{g' H_1 H_2 / (H_1 + H_2)} / f_0$  is the Rossby radius of deformation, and  $R_d \simeq 7 \text{ km}$  ( $H_1 =$   
 127  $50$  m,  $H_2^\infty = 2000$  m,  $f_0 = 10^{-4} \text{ s}^{-1}$  and  $g' = 1^\circ/oo$ ) for most of the experiments presented  
 128 below. !!!!!YM This value is smaller than the usual Rossby radius of deformation observed  
 129 in the deep ocean (around  $20 \text{ km}$ ). It is however consistent with the upper ocean stratification  
 130 in summer (when the seasonal pycnocline has been formed) and corresponds to mid-slope or  
 131 shelf characteristics. In addition, sensitivity experiments where the Rossby Radius is varied will  
 132 show that this parameter has a weak influence on the processes studied here (see section 4.2).  
 133 !!!!!YM

134 The parameters corresponding to the various simulations presented below can be found in  
 135 table 1 (fixed parameters) and 2 (variable parameters).

## 136 2.3. Potential vorticity and potential vorticity anomaly

137 For the shallow water model used here, the potential vorticity for each isopycnic layer is  
 138 defined as :

$$PV_k = \frac{f_0 + \zeta_k}{h_k}, \quad (4)$$

139 where  $\zeta_k = \nabla \times \mathbf{u}_k = \partial_x v_k - \partial_y u_k$  is relative vorticity in layer  $k$ , and  $h_k$  is the layer thick-  
 140 ness. In the absence of diabatic process, PV is conserved for each fluid particle. PV is  
 141 also related to the velocity field that can then be calculated by inverting the PV field un-  
 142 der the assumption of (cyclo-)geostrophic equilibrium. PV conservation and invertibility  
 143 are key properties which helped understand and interpret many geophysical fluid processes  
 144 ([McWilliams and Gent, 1980], [McIntyre and Norton, 1990], [Hoskins *et al.*, 1985], see also  
 145 [Morel *et al.*, 2006, Rossi *et al.*, 2009] for applications to upwelling dynamics).

146 PV is finite at rest and in order to invert it and to calculate the velocity, we use the potential  
 147 vorticity anomaly (PVA) which is defined in each layer  $k$  as the difference between the local PV

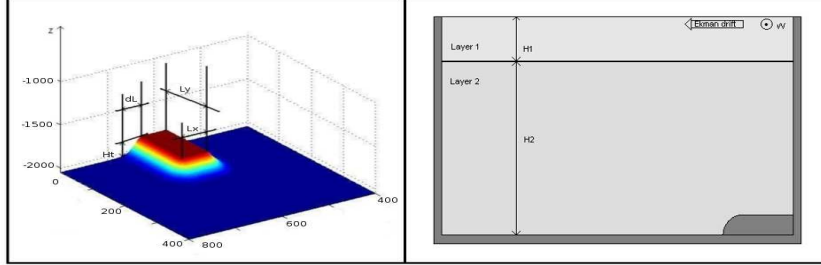


Figure 2: a : Details of the promontory : It is centered on  $x = 300\text{km}$ .  $L_x$  and  $L_y$  are respectively the length and width of its flat top,  $H_t$  its height, and  $dL$  is the typical length of its Gaussian sloping sides. b : Model configuration at rest : a shallow surface layer of depth  $H_1$  and density  $\rho_1$  lies over a bottom layer of depth  $H_2$  and density  $\rho_2$ . The numerical domain is a periodic zonal channel with vertical walls and a flat bottom.

148 and a reference  $PV$  for a state at rest (no current and flat isopycnals) (see [*Herbette et al.*, 2003,  
149 *Herbette et al.*, 2005, *Rossi et al.*, 2009]) :

$$PVA_k = H_k (PV_k - PV_k^{ref}) = H_k \left( \frac{f_0 + \zeta_k}{h_k} - \frac{f_0}{H_k} \right) = \frac{H_k}{h_k} (\zeta_k - f_0 \frac{\delta h_k}{H_k}), \quad (5)$$

150 where  $H_k(x, y)$  is the layer thickness at rest, and  $\delta h_k = h_k - H_k$ . Notice that we have also multiplied  
151 the  $PV$  difference by the layer thickness at rest so that  $PVA$  is proportional to the vorticity, which  
152 makes it easier to analyze.  $PVA$  contains the dynamical signal associated with the  $PV$  field and  
153 the geostrophic velocity field can be inferred from the  $PVA$  field. Notice that, as  $H_k$  is a function  
154 of position, contrary to  $PV$ ,  $PVA$  is not conserved for each particle in the presence of a variable  
155 bottom topography. It is however directly related to the circulation. The presence of a  $PVA$  pole  
156 in a layer  $k$  is indeed associated with a circulation extending to all layers but intensified in layer  $k$   
157 ([*Hoskins et al.*, 1985, *Rossi et al.*, 2009]) : a positive  $PVA$  pole being associated with a cyclonic  
158 circulation, a negative one with an anticyclonic circulation.

159 As shown in [*Verron and Le Provost*, 1985, *Herbette et al.*, 2003], when a current develops  
160 above a seamount, two opposite sign eddies appear : an anticyclone trapped above the topo-  
161 graphic feature associated with the displacement of low  $PV$  water columns from the deep ocean  
162 upon the seamount and a cyclone associated with the advection of high  $PV$  water columns off  
163 the topography. Figure 3) describes this process which is adiabatic and relies on the advection  
164 of  $PV$  and the formation of  $PVA$  poles. It also shows that between the two opposite sign eddies  
165 a strong jetlike current is formed.

#### 166 2.4. Previous results and general upwelling characteristics

167 [*Morel et al.*, 2006] found an exact analytical solution for the geostrophic circulation of a 2-D  
168 configuration with a flat bottom and a constant wind forcing  $T^w$ . In practice,  $T^w = \tau^w / (\rho_1 h_1)$



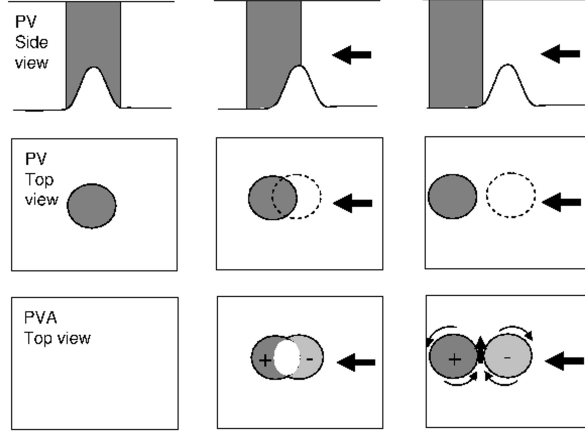


Figure 3: Schematic view of the development of a topographic dipole when a current develops above a seamount. The initial PV structure (top panels for the side view and middle panels for the top view) is associated with high PV above the seamount and lower PV in the deeper ocean; the fluid is initially at rest and the PVA (lower panels) is null. The current exchanges the position of low and high PV water columns which forms positive and negative PVA. An anticyclonic circulation develops above the topography whereas a cyclonic circulation is associated with the high PV water columns detaching from the seamount. The current is intensified between the opposite sign eddies.

169 (where  $\tau^w$  is the surface wind stress) is not constant but, with the approximation  $T^w = \tau^w / (\rho_1 H_1)$ ,  
 170 the following formulas provide a good evaluation for the position of the outcropping front and  
 171 velocity field as a function of the wind stress intensity and the duration of the wind forcing. It is  
 172 then possible to obtain a stationary basic state current with the desired characteristics by limiting  
 173 the forcing to a chosen period for a given wind stress intensity.

174 The distance of the outcropping front from the coast is :

$$Y(t) = \min\{0, -\frac{T^w}{f(1 + \delta)}(t - t_0)\} \quad (6)$$

$$t_0 = \frac{fR_d(1 + \delta)}{T^w} \quad (7)$$

175 and the alongshore velocity field in both layers is given by :

176 if  $y < Y(t)$

$$U_1 = U_c \exp \frac{y}{R_d} + U_b(t), \quad (8)$$

$$U_2 = -\delta U_c(t) \exp \frac{y}{R_d} + U_b(t), \quad (9)$$

177 if  $y \geq Y(t)$

$$U_1 \quad \text{undefined}, \quad (10)$$

$$U_2 = -f\delta(y - Y(t)) - \delta U_c(t) \exp \frac{Y}{R_d} + U_b(t). \quad (11)$$

178 where  $\delta = H_1/H_2$ ,  $t_0$  is the time necessary for the lower layer to outcrop at the coast ( $y = 0$ ).  $U_b(t)$

179 and  $U_c(t)$  are the temporal evolution of the barotropic and baroclinic components respectively  
 180 and can be written as :  
 181 if  $t < t_0$

$$U_b = \frac{\delta T^w}{1 + \delta} t \quad (12)$$

$$U_c = \frac{T^w}{1 + \delta} t \quad (13)$$

182 if  $t \geq t_0$

$$U_b = \frac{\delta T^w}{1 + \delta} t \quad (14)$$

$$U_c = f R_d \exp \frac{-Y(t)}{R_d} \quad (15)$$

183 Notice that the maximum current is reached at the outcropping front (for  $t \geq t_0$ ) and is given  
 184 by:

$$\begin{aligned} U_1^{max} &= f R_d + \frac{\delta T^w}{1 + \delta} t \\ &= (1 + \delta) f R_d + \delta f |Y(t)| \end{aligned} \quad (16)$$

185 Notice that the barotropic mode is spatially constant and only the cross shore spatial structure  
 186 of the baroclinic mode varies as  $\exp \frac{y}{R_d}$ . In addition, the amplitude of the baroclinic component  
 187 of the velocity field is limited whereas the barotropic one grows linearly with time (until other  
 188 processes such as bottom friction become non-negligible).

189 The wind stress corresponding to a 30 knots wind (15 m/s) is  $\tau^w \simeq 0.2 N/m^2$  and thus we get  
 190  $T^w \simeq 4 \cdot 10^{-6} m/s^2$  (for  $H_1 = 50 m$ ). Then, the previous formulas show that it takes about  
 191  $t_o \simeq 2$  days for the outcropping front to be generated and after 10 days of wind forcing, the front  
 192 is located  $Y \simeq 35 km$  offshore and the maximum velocity at the front is about  $U_1^{max} \simeq 70 cm/s$ .  
 193 The barotropic velocity, and the velocity field in the deep layer over most of the domain, are  
 194  $U_b \simeq 8 cm/s$ . The characteristics of the upwelling found in the numerical simulations presented  
 195 below are in very good agreement with these analytical results.

### 196 3. Reference experiment

197 The wind forcing was kept constant  $T^w = \tau^w / (\rho_1 H_1)$  in the reference simulation that we  
 198 present here.

199 Figure 4 shows the evolution of the *PVA* in the upper layer superimposed on the shape of the  
 200 promontory for the reference experiment (see table 2). During the upwelling development, the  
 201 upper layer vanishes close to the coast and is replaced by deep waters that reach the sea surface.  
 202 This area is associated with an infinite *PVA* in the upper layer (see [Bretherton, 1966]) and is thus  
 203 delimited by a strong *PVA* gradient that we use to trace the upwelling front and the development  
 204 of the upwelling filaments (it is qualitatively comparable to the sea surface temperature front).

205 The strong *PVA* gradient associated with the upwelling front becomes evident on the third  
 206 day of the experiment. It is accompanied by an intense baroclinic surface intensified jet super-  
 207 imposed on a spatially constant barotropic westward flow. The influence of bottom topography

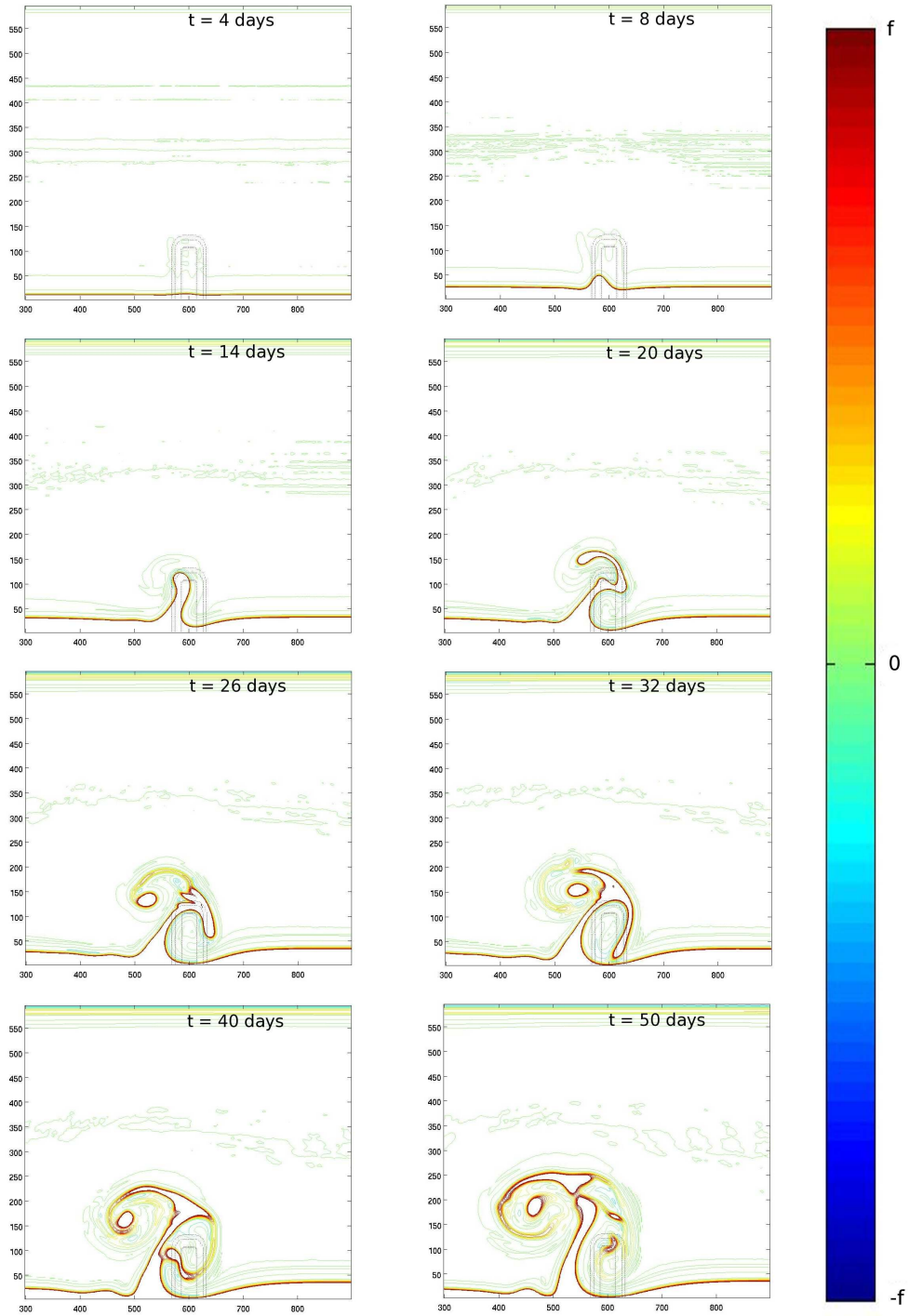


Figure 4: Evolution of the PVA in the upper layer for the reference experiment at  $t = 4, 8, 14, 20, 26, 32, 40, 50$  days. The thick red line represents the  $PVA = +f$  contour and is a good marker of the upwelling front. The axis are labelled in km.

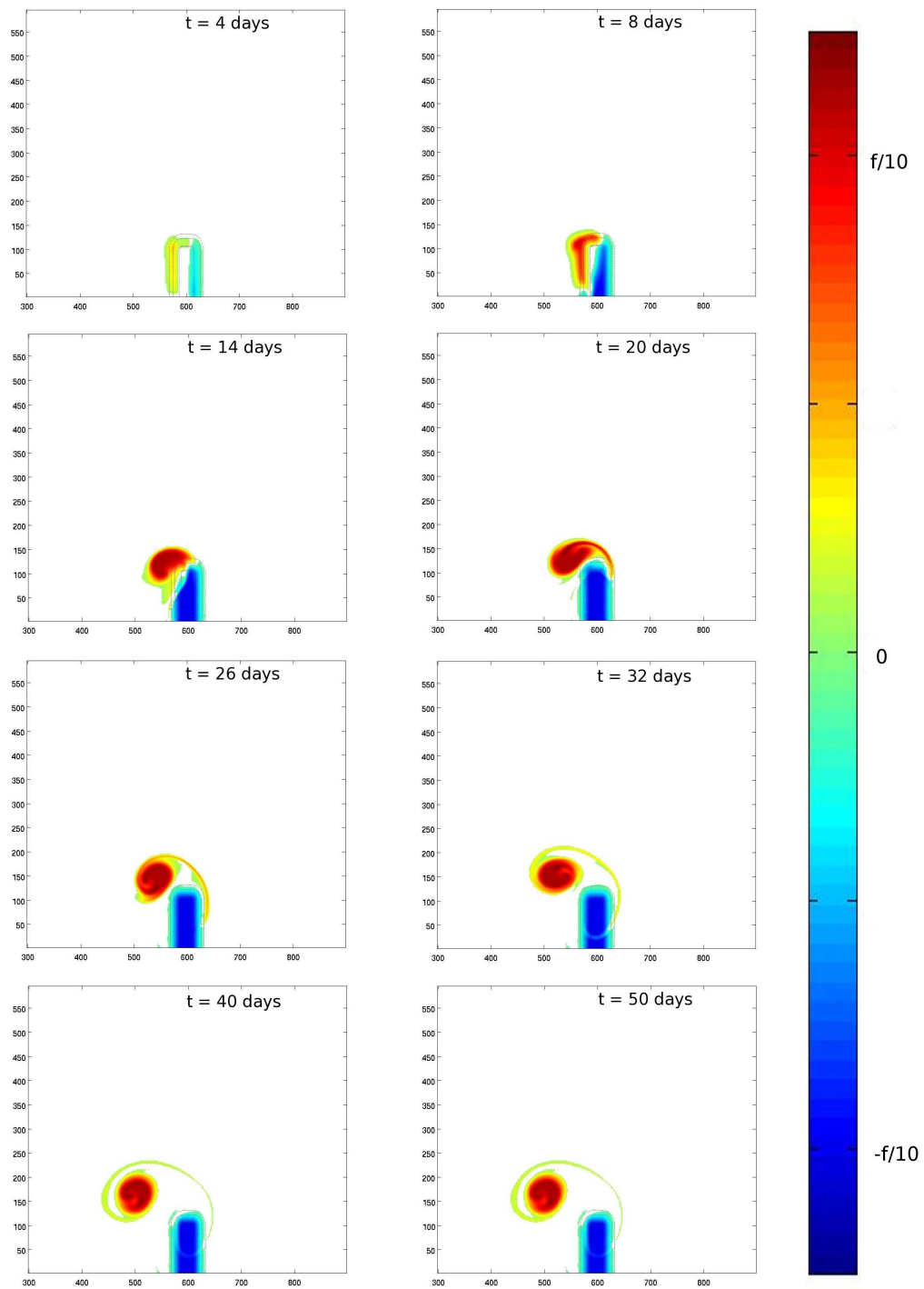


Figure 5: Evolution of the PVA in the bottom layer for the reference experiment at  $t = 4, 8, 14, 20, 26, 32, 40, 50$  days. The generation of PVA is visible in the first 10 days. It is then advected from day 10 to day 50.

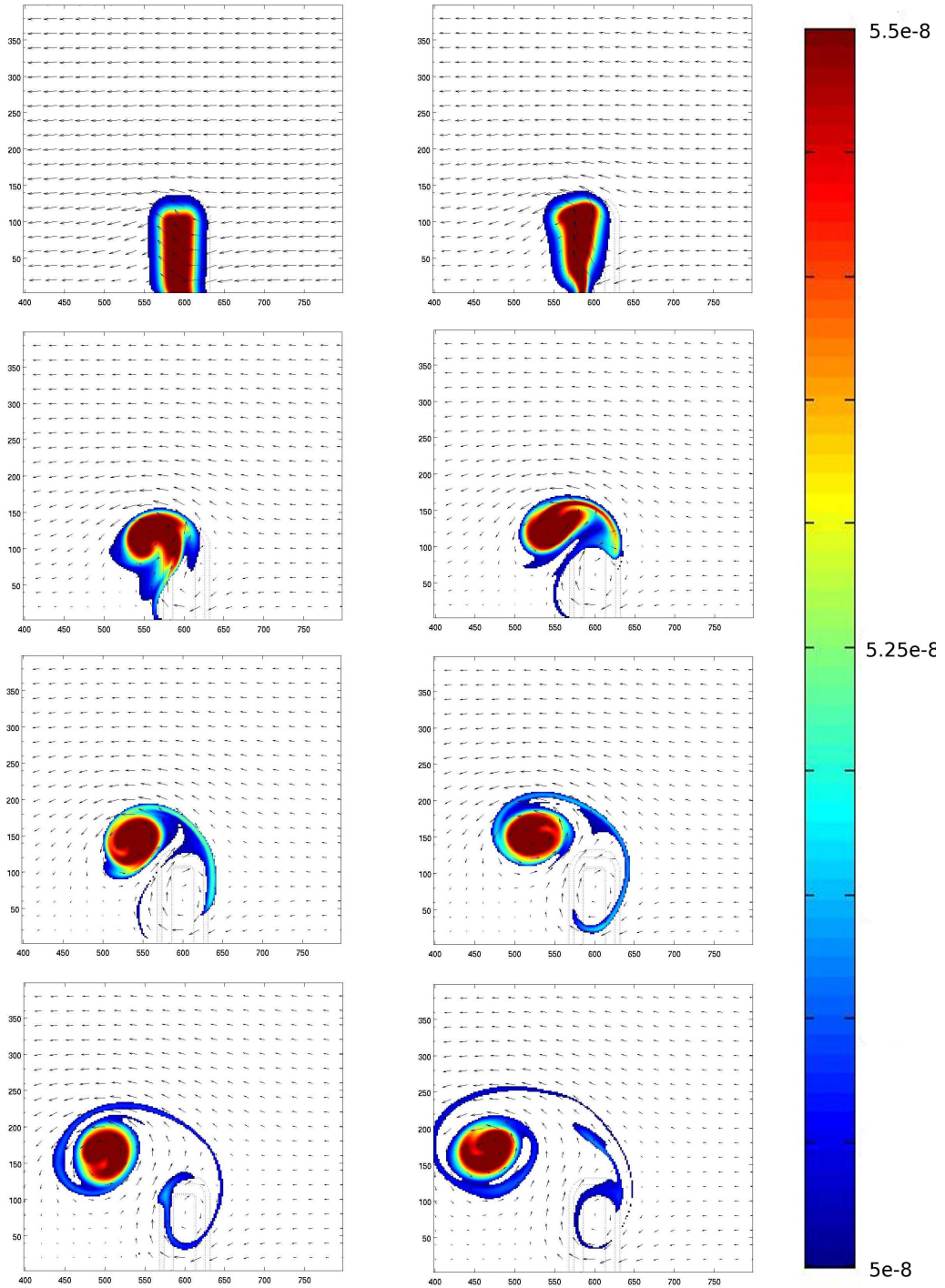


Figure 6: Evolution of the bottom layer absolute potential vorticity for the reference experiment superimposed on the velocity field. The high PV pool is visible on the promontory. Note that potential vorticity is strictly conserved and is only advected by the current. Comparison with Fig. 5 also shows that the anticyclonic and cyclonic circulations are associated with the negative and positive PVA poles that develop because of bottom topography, as expected.

208 is noticeable after 4 days, when the front begins to form a small meander on the western (down-  
 209 stream) side of the promontory. The initial topographical meander keeps on growing offshore  
 210 downstream of the promontory. After 12 days of simulation, it starts to roll up anticyclonically  
 211 around the promontory. At day 16, another branch appears, growing offshore but partly rolling  
 212 up cyclonically. The meander has then become a thin and elongated filament, surrounded by  
 213 two counter-rotating recirculations. Its length is about 200 km and its width about 100 km near  
 214 the coast and 20 km near the tip. It is similar to the 'squirts', observed in all major upwelling  
 215 systems. After 28 days, the filament is still growing offshore and is about 220 km long, but its  
 216 offshore edge has rolled up cyclonically. As shown on Fig. 4, the filament continues to grow  
 217 until the end of the simulation (it is about 230 km long after 50 days), corresponding to a mean  
 218 growth rate of about 5 km per day for the whole life cycle, consistent with the observed mean  
 219 value found by [Kostianoy and Zatsepin, 1996]. But the growth rate of the filament is variable  
 220 during the life cycle, with higher initial growth rates of up to  $12 \text{ km day}^{-1}$  (i.e.  $14 \text{ cm s}^{-1}$ ) after  
 221 10 days, then decreasing until the end of the experiment.

222 The formation of the filament is very similar to the erosion process of a surface vortex by  
 223 a seamount studied in [Herbette et al., 2003]. As already explained above, the origin of these  
 224 counter-rotating eddies can indeed be inferred from the  $PV$  and  $PVA$  evolution in the lower  
 225 layer. Indeed, as shown in [Morel et al., 2006, Rossi et al., 2009] a barotropic westward current  
 226 is generated during the upwelling development. The baroclinic circulation opposes this current  
 227 in the upper layer, but its extension is of the order of the first internal radius of deformation  
 228 (about 7km here) which is quite small. As a result, water columns move westward over most of  
 229 the lower layer. As the dynamics is adiabatic here, the initial  $PV$  field is simply advected and  
 230 the positive anomaly associated with the promontory moves downstream (see Fig. 6) replaced  
 231 by lower  $PV$  water columns coming from deeper region. Figure 5 shows that it creates opposite  
 232 sign  $PVA$  (see also [Herbette et al., 2003]): high  $PV$  water columns coming from the promontory  
 233 move in regions with lower  $PV$  at rest, forming a positive  $PVA$  downstream of the promontory,  
 234 while low  $PV$  water columns move upon the promontory, which is associated with high  $PV$  at  
 235 rest, forming a negative  $PVA$  that is being trapped above the promontory. As shown in Fig. 6,  
 236 this topographic  $PVA$  dipole is associated with cyclonic and anticyclonic circulations extending  
 237 over the whole water column. An offshore jetlike current develops between the two opposite sign  
 238  $PVA$  poles, which forms the filament.

239 After 7 days both positive and negative  $PVA$  reach a maximum modulus of  $\pm 0.1f$ , the nega-  
 240 tive  $PVA$  obviously remains trapped above the topography, maintaining offshore currents on the  
 241 western side of the promontory which reach about  $40 \text{ cms}^{-1}$ . But the high  $PVA$  pole is strongly  
 242 deformed and propagates offshore and westward under the combined effect of advection and in-  
 243 teraction with bottom topography. The topographic  $\beta$ -drift of a  $PVA$  pole along the slope of the  
 244 promontory scales -in the quasi-geostrophic approximation- as  $U_{drift} = \beta_t R_d^2$ , with  $\beta_t = f_0 \alpha / H_2$   
 245 where  $\alpha = H_t / dL$  is the characteristic slope of the promontory, and  $R_d^2$  is the square of the Rossby  
 246 radius. For the reference experiment,  $U_{drift}$  is found to be about  $2.5 \text{ cms}^{-1}$ . The barotropic ve-  
 247 locity field associated with a  $PVA$  pole in the lower layer can be scaled using the circulation  
 248 theorem :

$$U^- \approx C / 2\pi l \quad (17)$$

$$C = \int \int_P PVA \frac{h_2}{H_{tot}} dS \quad (18)$$

$$\simeq \int \int_P PVA \, dS \quad (19)$$

249 where  $C$  is the circulation or total PVA reservoir inside a domain  $P$ ,  $l$  is the distance from the  
 250 center of the PVA pole,  $H_{tot} = H_1 + H_2$  and  $dS$  is the surface element. If we assume that all water  
 251 columns above the promontory have been replaced by water columns coming from the deeper  
 252 ocean, the negative PVA forming above the promontory is given by :

$$PVA = -\frac{f H_p(x, y)}{H_2^\infty} \quad (20)$$

253 where  $H_p(x, y)$  is the the promontory height.

254 The total circulation associated with the negative PVA of the promontory, over the domain  $P$   
 255 is then

$$C = -\frac{f}{H_2} \int \int_P H_p(x, y) dS \quad (21)$$

$$\simeq -\frac{f}{H_2} H_t [L_x L_y + dL (L_x + L_y/2)] \quad (22)$$

256 For the positive pole, the calculations are similar : the PVA reservoir, and thus circulation, is  
 257 exactly the opposite of the negative one above the promontory (water columns are exchanged  
 258 between the deep ocean and the promontory). As the effect of both PVA poles superimposes, the  
 259 maximum barotropic current between both poles is thus roughly given by :

$$U_{max}^{jet} \simeq 2C/2\pi l \quad (23)$$

$$C \simeq \frac{f}{H_2} H_t [L_x L_y + dL (L_x + L_y/2)] \quad (24)$$

260 where  $l$  is the mid distance between both pole centers.

261 When both PVA poles are well developed,  $l \simeq 30$  km and  $U_{max}^{jet} \simeq 36$  cm/s, which is the  
 262 order of magnitude of the maximum offshore current observed downstream of the promontory  
 263 (40 cm/s). These modelled velocities are in very good agreement with what have been observed  
 264 in-situ in the IPUS area and also in other upwelling regions (see [Sanchez *et al.*, 2008]).

265 Notice that the estimation of  $U_{max}^{jet}$  or  $U^-$  is only correct in the case of circular PVA structures,  
 266 or far enough from the structure so that shape effects become negligible. Here, this is obviously  
 267 not verified, but, using this simple scaling can give us a good insight of the order of magnitude  
 268 of the velocity associated with the topographic PVA pole development and its sensitivity to the  
 269 promontory characteristics.

270 In summary, an anticyclonic circulation is generated and trapped above the promontory by  
 271 advection of low PV over the topography, forming a negative PVA pole. A cyclonic circulation  
 272 also forms because of advection of high PV from the promontory into a deeper environment.  
 273 This forms a trapped topographic dipole associated with a strong offshore current that generates  
 274 the filament and its well known 'squirt' or 'mushroom' shapes (see [Strub *et al.*, 1991]). The  
 275 strength of the current depends on the total PVA reservoir of the promontory.

276 Finally, notice that, even though the initial topographic cyclone slowly separates from the  
 277 trapped anticyclone, because of the outcropping and vanishing of the upper layer, the meander  
 278 and filament are themselves also associated with an equivalent high PVA (see [Bretherton, 1966])  
 279 reinforcing and maintaining a cyclonic circulation on the downstream side of the negative PVA  
 280 pole.

## 281 4. Sensitivity experiments

282 To strengthen the physical relevance of the mechanism described above and to assess the  
283 respective importance of the various parameters and characteristics of the configuration, a set of  
284 sensitivity tests was performed. Here we focused on the stability of the front, the forcing duration  
285 time, of the promontory characteristics (width, length, height and slopes), of the stratification and  
286 of bottom friction. For comparison of the various model output, we take as a reference time the  
287  $t = 42$  days (6 weeks) output, and use the *PVA* maps as a qualitative indicator of the efficiency  
288 of the model configuration to produce long, coherent and trapped filaments.

### 289 4.1. The stability of the front

290 The reference run showed that the sole presence of the topography allowed the de-  
291 velopment of a long filamentary structure reaching as far as 230 km offshore. How-  
292 ever, mixed barotropic-baroclinic instability is a well known feature of upwelling currents  
293 [Shi and Roed, 1999] and has sometimes also been referred to as the main process for filament  
294 formation [Haidvogel et al., 1991]. It is thus important to evaluate the relative influence of to-  
295 pography and intrinsic instability on the development of the long filament.

296 Baroclinic instability can only develop when there exists opposite sign potential vorticity gra-  
297 dients or *PVA* (see [Charney and Stern, 1962]). The outcropping front is associated with positive  
298 *PVA*. As shown by [Morel et al., 2006], negative *PVA* is generated along the upwelling front  
299 (see Fig. 7 below) because as isopycnic surfaces bend upward they enter the region influenced  
300 by the wind stress. A wind stress curl then exists along isopycnic levels which has been shown to  
301 necessarily form negative *PVA* (the formation of negative *PVA* by the wind has also been studied  
302 in [Thomas, 2005]). In the simple 2-layer configuration used here, this effect is associated with  
303 the fact that  $T^w = \tau^w / (\rho_1 h_1)$  varies with the layer depth  $h_1$ .

304 In the reference experiment, the possibility of the flow to become baroclinically unstable has  
305 thus been suppressed by modifying the distribution of the wind forcing :  $T^w = \tau^w / (\rho_1 H_1)$  pro-  
306 vides a constant wind stress so that the upwelling still develops but the dynamics remains adia-  
307 batic and the PV field is conserved. As a result, no negative *PVA* is formed along the front and  
308 no baroclinic instability can develop.

309 In the present test, we use the actual wind forcing  $T^w = \tau^w / (\rho_1 h_1)$ . Figures 7 and 8 show  
310 the evolution of the *PVA* in the upper layer and in the lower layer respectively. In compari-  
311 son with the reference experiment (see Fig. 4 and 5), negative *PVA* is develops along the up-  
312 welling front. This negative *PVA* strip interacts with the positive *PVA* associated with the out-  
313 cropping forming new small-scale meanders, with wavelengths of 30 km after 10 days (notice  
314 the association of the small upwelling front meanders with small negative *PVA* poles). These  
315 small-scale meanders are associated with baroclinic (or sometimes called frontal) instabilities  
316 ([Barth, 1989 a, Barth, 1989 b, Morel et al., 2006, Capet and Carton, 2004, Killworth, 1980])  
317 but are neither trapped nor forming long filaments. They indeed propagate along the upwelling  
318 front, re-enter the domain on the eastern side and only very slowly develop after their initial  
319 growing. After 5 weeks of experiment, their offshore extension is less than 50 km (from the  
320 front).

321 The impact on the main filament is also minor: the positive and negative *PVA* poles still  
322 develop in the bottom layer and their time evolution is not significantly modified. The surface  
323 filament is very similar to the one observed in the stable case and it extends as far offshore. The  
324 only noticeable difference, apart from the absence of the small amplitude meanders along the  
325 front, is that the topographic filament is here truncated by the small scale eddies and also appears



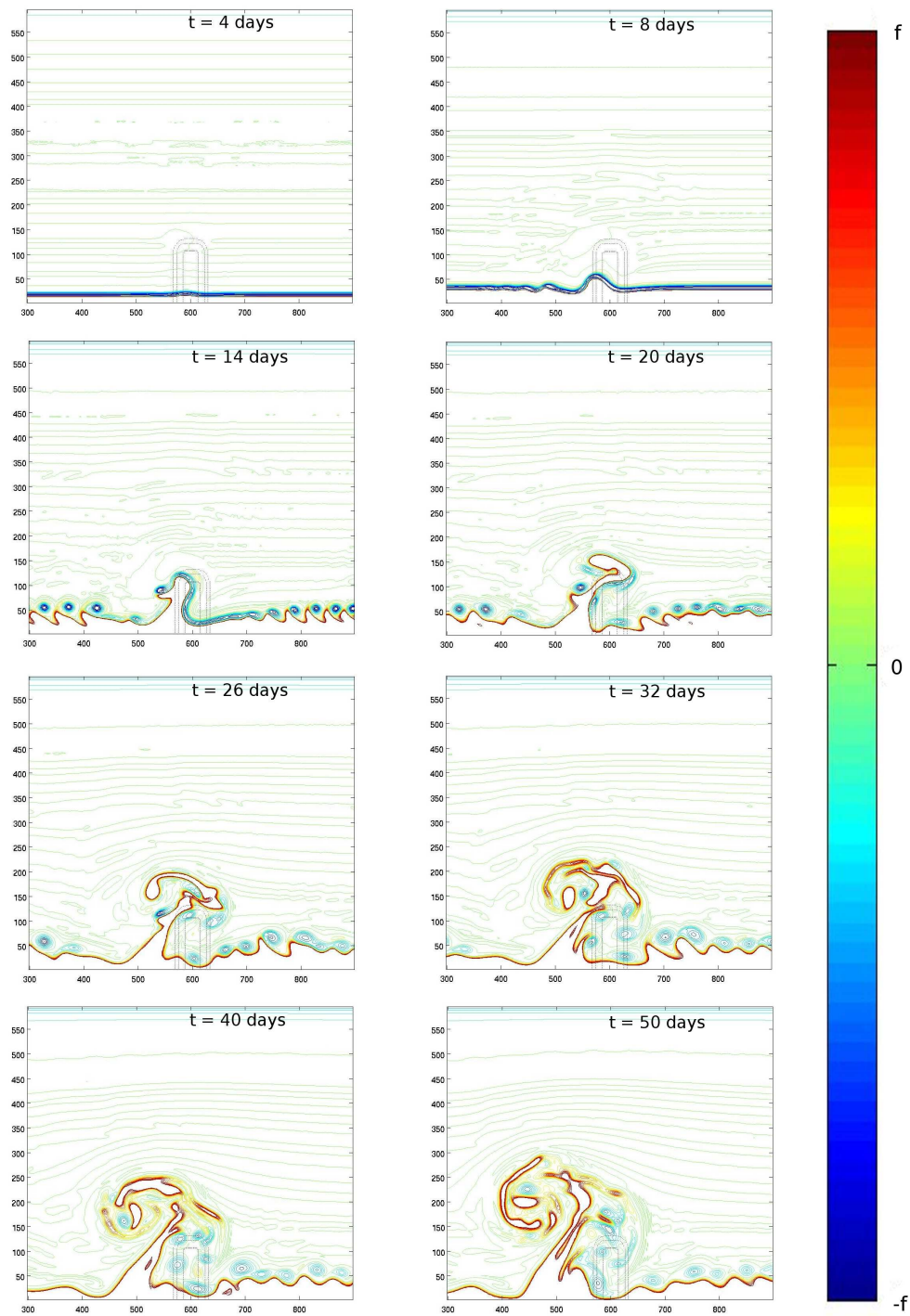


Figure 7: Evolution of the PVA in the upper layer for the unstable experiment at  $t = 4, 8, 14, 20, 26, 32, 40, 50$  days. Notice the additional smaller meanders, but the formation of the large filament is the same as in Fig. 4.

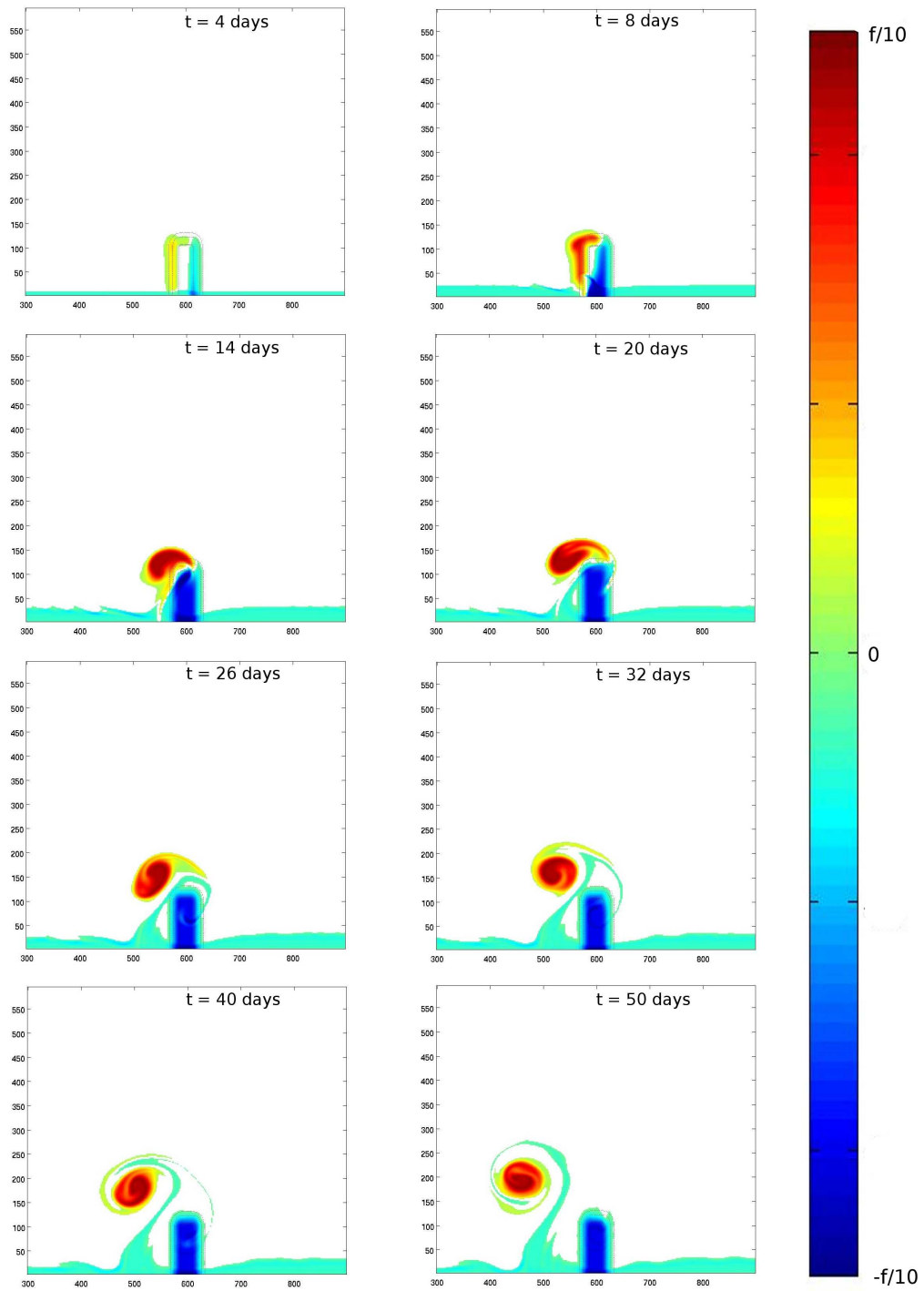


Figure 8: Evolution of the PVA in the bottom layer for the unstable experiment at  $t = 4, 8, 14, 20, 26, 32, 40, 50$  days.

326 slightly larger at its base. This proves that, at least in our simplified configuration, intrinsic  
327 baroclinic instability has little influence on the formation of the long filament and that the main  
328 mechanism is associated with topography, as described above.

329 The stable front configuration is also of particular interest because the dynamics is adiabatic  
330 and  $PV$  is conserved following fluid parcels in all layers and can be used as a tracer. For this  
331 reason, and because we have shown that no substantial difference existed in the formation of the  
332 filament, we keep the stable front configuration as our reference experiment for the following  
333 sensitivity tests that will thus be carried with the modified and constant wind forcing.

#### 334 4.2. The influence of the stratification characteristics

335 In addition to the reference experiment ( $g' = 0.01$ ,  $H_1 = 50$  m), six additional experiments were  
336 performed to evaluate the influence of the stratification characteristics on the dynamics of the  
337 topographic filament : 3 experiments varying  $g'$  (0.005, 0.02 and  $0.03 \text{ ms}^{-2}$ ), and 3 experiments  
338 varying  $H_1$  (25, 100, and 200 m). It may seem redundant to vary both parameters (as they  
339 both influence the Rossby radius) but we finally found out that their respective influence on the  
340 upwelling front evolution is quite different.

341 Figure 9 represent the upper layer PVA after 42 days for different values of  $g'$  and shows  
342 only modest modification of the filament. This is not entirely surprising as the density jump  
343 mostly influences the baroclinic currents in the vicinity of the front via the Rossby radius of  
344 deformation. Topographic eddies are formed and influence the dynamics through the barotropic  
345 circulation, which is not modified. In addition, the position of the upwelling front is also only  
346 slightly affected by a modification of  $g'$  : the offshore displacement is not modified and only the  
347 initial time at which the outcropping front forms depends on this parameter.

348 Varying  $H_1$  (Fig. 10) does not modify the lower layer dynamics either (see the similarities  
349 of the PVA structure in the lower layer after 42 days on the right panels of Fig. 10). However,  
350 since it also plays a role in the strength of the wind forcing ( $T^w = \tau^w / (\rho_1 H_1)$ ), it strongly mod-  
351 ifies the position of the upwelling front, which forms later and extends more slowly for deeper  
352 thermoclines (larger  $H_1$ ). The differences in the filament evolution with different  $H_1$  is thus the  
353 result of the time lag between the upwelling front evolution associated with  $H_1$  and the distribu-  
354 tion of the topographic eddies when the outcropping first forms. As a result, the advective effect  
355 of the topographic eddies on the upwelling front is in general simply delayed. The time period  
356 necessary for the upwelling front to be formed is  $t_o \approx 0.7, 2, 6$  and 16 days for  $H_1 = 25, 50,$   
357 100 and 200 m respectively. As a result, for the duration of the wind forcing considered here  
358 (10 days), varying  $H_1$  does not strongly modify the filament except for the deepest thermocline  
359 (here associated with the case  $H_1 = 200\text{m}$ ) for which the upwelling front is not formed and no  
360 filament is then visible (see Fig. 10 lower panel). Interestingly, the final offshore extent has close  
361 values for the all other experiments.

362 Finally notice that the experiment where  $g'$  is varied and the experiment where  $H_1$  is varied  
363 have different Rossby radius of deformation :  $R_d = 5, 7, 10, 14$  km, for  $H_1 = 25, 50, 100, 200$   
364 m respectively (or  $g' = 0.005, 0.01, 0.02$  and  $0.03$  respectively). This underlines again that  
365 the important mechanisms for the filament development is the barotropic circulation and the  
366 formation of the upwelling front. In our configuration the development of the filament is mainly  
367 controlled by the bottom layer PVA evolution which is almost insensitive to  $R_d$ .

#### 368 4.3. The forcing duration time

369 We here study the effect of a variation of the wind forcing duration time. As seen above, the  
370 wind forcing acts both on the offshore front position and the velocity strength, especially the one

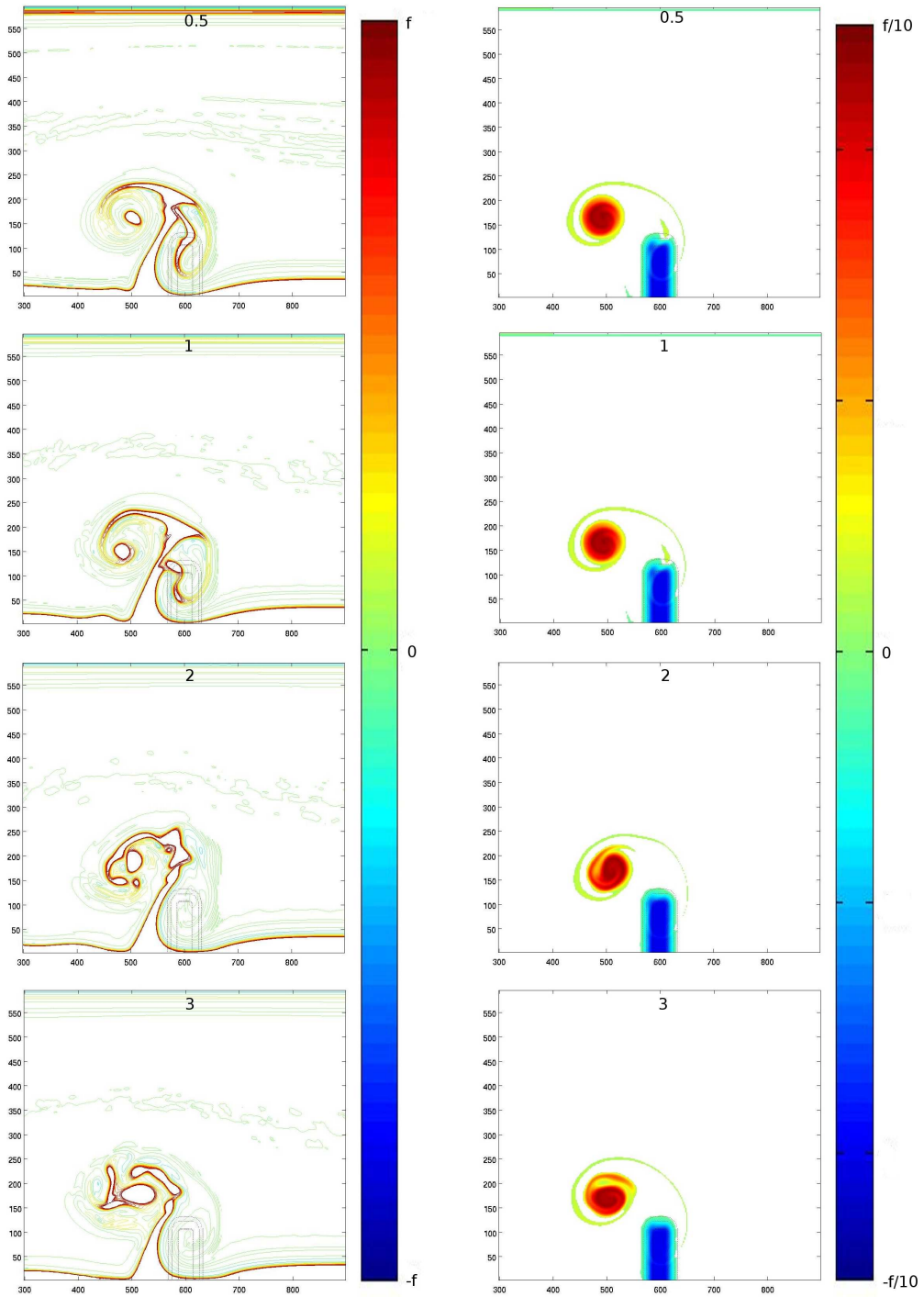


Figure 9: Maps of *PVA* in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $g' = 0.005, 0.01, 0.02$  and  $0.03 \text{ms}^{-2}$  experiments. !!!

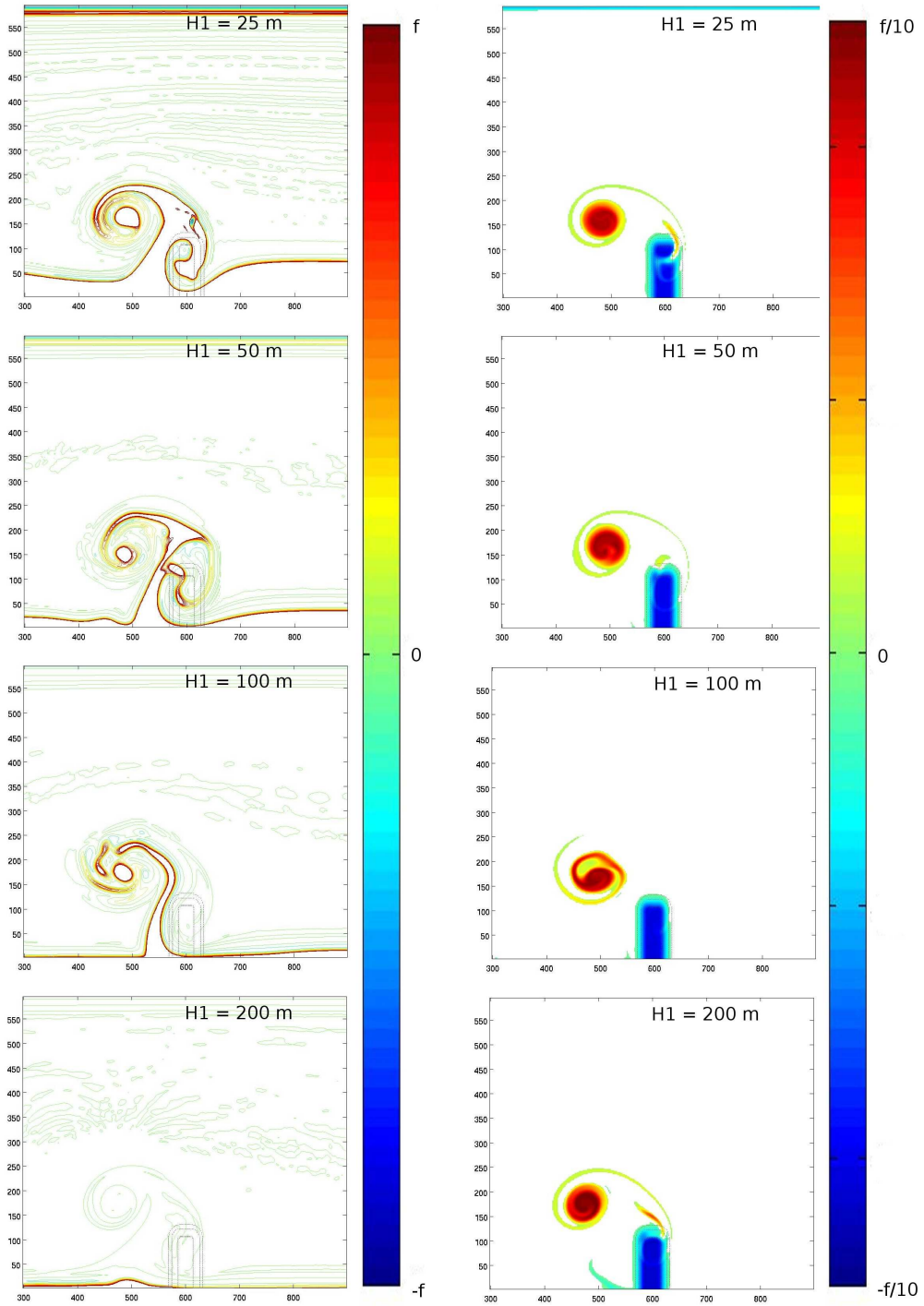


Figure 10: Maps of *PVA* in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $H_1 = 25, 50, 100, 200$ m experiments.

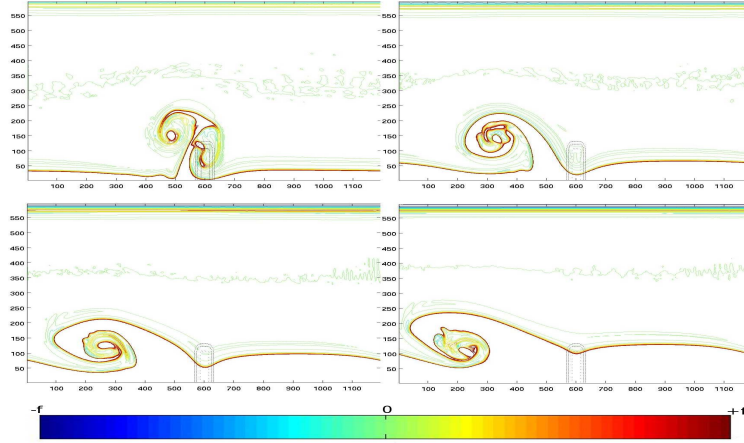


Figure 11: Maps of PVA in the upper layer at  $t = 42$  days for the 10, 20, 30 and 40 days of wind forcing cases.

371 of the barotropic current.

372 Figure 11 shows the *PVA* front after 42 days of experiments in the upper layer for forcing  
 373 durations of 10 (reference experiment), 20, 30, and 40 days. There still exists a trapped filament  
 374 that extends far offshore downstream of the promontory, but its characteristics drastically depend  
 375 on the forcing duration : it becomes thicker and bends downstream when the wind blows for a  
 376 long time.

377 In fact, increasing the wind forcing duration does not substantially modify the formation of  
 378 the *PVA* in the bottom layer but induces stronger barotropic current directed downstream. This  
 379 intensified upwelling can mask the topographic circulation and the offshore jet. In addition, as  
 380 the barotropic current increases, the positive pole becomes quickly advected downstream and  
 381 only shortly interacts with the negative *PVA* pole on the topography. As a result, the upper  
 382 layer *PVA* front rolls up cyclonically around the bottom layer positive *PVA* pole and is entrained  
 383 downstream, giving it a breaking wave like shape. When increasing the forcing duration time, the  
 384 distance from the front and jet to the coastal wall increases, while the offshore distortion of the  
 385 front is less obvious, since its initial position almost reaches the offshore edge of the promontory  
 386 (see the 40 day forcing case).

387 Notice that according to Eq. 13 and 15 the maximum barotropic velocity is roughly given  
 388 by  $U_b^{max} \simeq 8,6 \cdot 10^{-3} t_d$ , where  $t_d$  is the duration time of the wind forcing in days. We thus get  
 389  $U_b^{max} \simeq 8,6 \text{ cm/s}$  for 10 days and  $U_b^{max} \simeq 34,4 \text{ cm/s}$  for 40 days, which is stronger than the  
 390 offshore advection associated with the topographic eddies. Notice such barotropic currents are  
 391 far beyond what is observed, at least offshore the continental shelf, and that in practice, bottom  
 392 friction keeps the barotropic velocity from reaching such values.

393 4.4. The promontory height

394 As the main process proposed here for the development of an upwelling filament is the gen-  
395 eration of topographic *PVA* in the bottom layer associated with the existence of a promontory,  
396 it is important to detail how the shape and size of the latter can affect *PVA* generation and thus  
397 filamentation. The maximum *PVA* and the strength of the topographic eddies are proportional to  
398 the height of the promontory which is thus a key parameter.

399 Six experiments were carried out with different promontory heights :  $H_t = 50, 100, 300, 500,$   
400  $1000$  and  $1500$  m, to be compared with the  $200$  m of the reference experiment. Figure 12 shows  
401 the upper and lower layer *PVA* field after 42 days for the  $50, 200, 500,$  and  $1500$  m experiments.  
402 For small topographies ( $H_t = 50$  m), the filament forming in the upper layer along the front  
403 has a much smaller offshore extension, is less pinched off and its tip is advected downstream.  
404 In fact, the effect is the same as for the influence of the forcing duration discussed above: the  
405 topographic circulation becomes much smaller than the upwelling current (the offshore current  
406 is about  $10 \text{ cm s}^{-1}$  for  $H_t = 50$  m, to be compared with the reference experiment where it is  
407 about  $40 \text{ cm s}^{-1}$ ). The positive *PV* pole is quickly advected downstream and the offshore current  
408 is masked by the upwelling current giving the filament a breaking wave shape and limiting its  
409 offshore extension.

410 The  $H_t = 500$  and  $H_t = 1500$  m experiments (see Fig. 12 left panels) show that after 42 days  
411 of experiment, the filament is also much reduced in comparison with the reference experiment  
412 ( $100$  km for  $H_t = 1500$  m, and  $140$  km for  $H_t = 500$  m). The limiting factor for large  $H_t$  is  
413 associated with the difficulty for water columns to climb on or leave the topography. Indeed,  
414 most of the positive and negative *PVA* in the bottom layer remains trapped on the slope (see  
415 Fig. 12 right panels). As a result, instead of forming two strong opposite sign *PVA* poles that  
416 locally reinforce the offshore circulation, *PVA* of both signs mix on the promontory evolving  
417 into a complex pattern of multiple small poles with few coherence. The overall integrated *PVA*  
418 and circulation associated with the topographic eddies is then much reduced. In fact, as already  
419 found by [Herbette et al., 2003] in the case of a vortex interaction with a seamount, the topo-  
420 graphic circulation can not be much stronger than the background velocity and the *PVA* creation  
421 is limited. Figure 13 shows the maximum extent of the filament for the 7 experiments. Similarly  
422 to [Herbette et al., 2003]'s optimum value of seamount height for vortex erosion, there exists an  
423 optimal promontory height for the filament extension which corresponds here to the reference  
424 experiment :  $H_t = 200$  m.

425 4.5. The promontory width

426 Three experiments were carried out to test the sensitivity of the filament formation to the width  
427 of the promontory (parameter  $L_x$ ). We tested  $L_x = 0$  km (Gaussian ridge),  $20$  km (reference  
428 experiment),  $50,$  and  $100$  km (see Fig. 14). This parameter mostly affects the *PVA* reservoir and  
429 strength of the circulation associated with the topographic eddies. The evolution of *PVA* in the  
430 bottom layer for small  $L_x$  exhibits similarities with the reference experiment, with a generation  
431 of negative *PVA* on the upstream part of the promontory in the first week of experiment, fully  
432 invading it after 10 days, and a generation of positive *PVA* downstream of the ridge, quickly  
433 evolving into a cyclonic vortex detaching from the offshore edge of the ridge. In fact, in the case  
434 of small  $L_x$ , most of the *PVA* reservoir is contained along the promontory slope, not above the  
435 plateau, and the results are then obviously not sensitive to this parameter in this case.

436 Increasing  $L_x$  increases the *PVA* reservoir and the potential strength of the topographic eddies.  
437 The  $L_x = 50$  km and  $100$  km cases show a strong rolling up of the positive *PVA* and of the filament



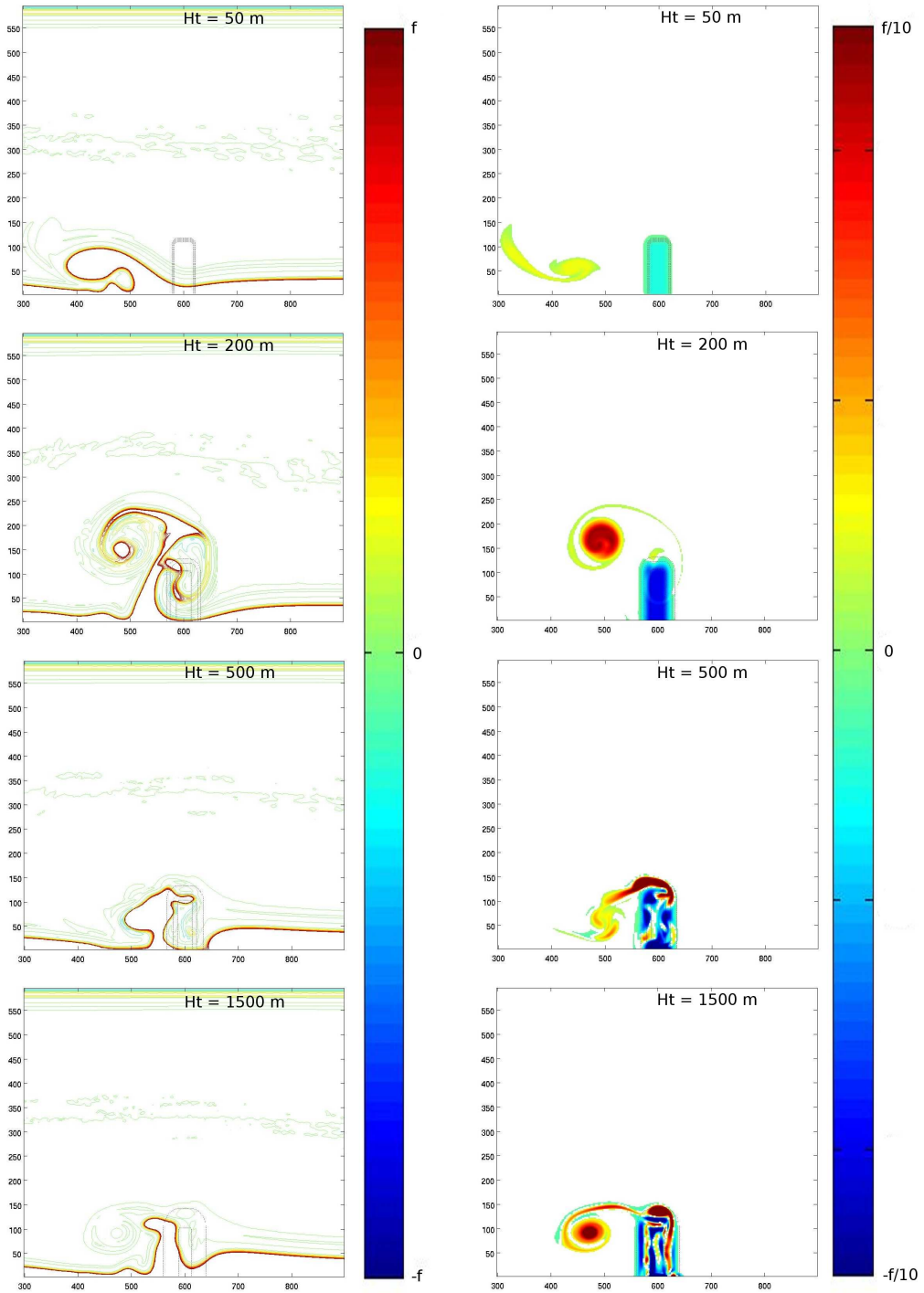


Figure 12: Maps of *PVA* in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $H_t = 50, 200, 500$  and  $1500m$  experiments.



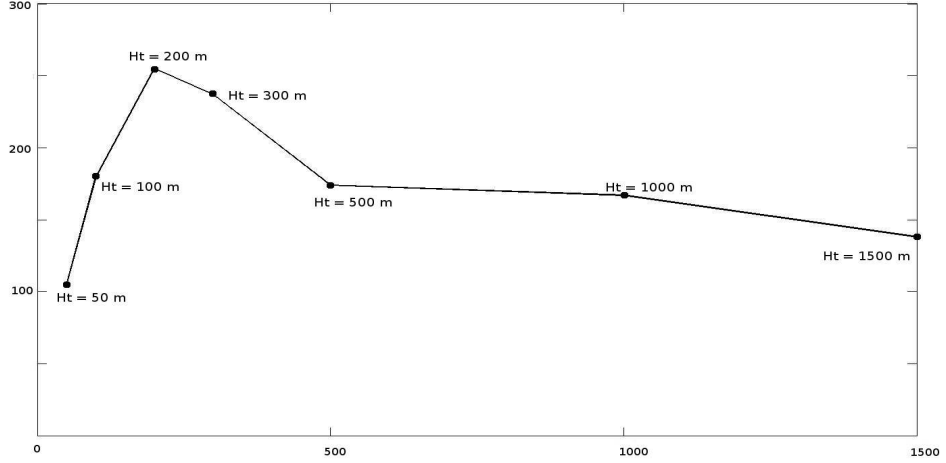


Figure 13: Offshore extent of the filament after 50 days of experiment for the  $H_t = 50, 100, 200, 300, 500, 1000$  and  $1500m$  cases. The x axis is  $H_t$  (kms) and the y axis is offshore distance (kms).

438 around the promontory. The *PVA* structure also exhibits a multipolar structure and the filament  
 439 has multiple branches that do not extend very far offshore. In fact, strong topographic eddies  
 440 leads to complex non linear interactions between the opposite sign *PVA* poles. The position  
 441 and shape of the negative *PVA* is fixed and remains trapped above the promontory, whereas  
 442 the positive one is advected and deformed by the total velocity field that develops in the lower  
 443 layer. The latter effect is a combination between the barotropic circulation associated with the  
 444 upwelling development, which is spatially constant, and the anticyclonic eddy, which varies  
 445 spatially and can induce strong deformations. When the *PVA* reservoir increases, the effect of the  
 446 negative *PVA* pole dominates the positive *PVA* pole and filament dynamics which are advected  
 447 anticyclonically around the promontory and deformed. This greatly reduces the total length of  
 448 the filament.

449 As a result, the width of the promontory also plays an important role in the development of  
 450 a coherent filament structure in the upper layer and again, there exists an optimum value for  
 451 the width of the promontory. This is shown in Fig. 15 where the offshore extension of the  
 452 topographic filament is plotted for various choices of  $L_x$ . The optimum value is  $L_x = 20$  km  
 453 (reference experiment) for the present configuration.

#### 454 4.6. The side slopes

455 In order to evaluate the importance of topographic  $\beta$ -effect in the offshore displacement of the  
 456 positive *PVA*, three experiments were performed with different margin lengths for the promon-  
 457 tory :  $dL = 0, 10, 20$  (reference experiment) and  $40$  km. This parameter acts on the *PVA*  
 458 reservoir (with close similarities with  $L_x$ ) but also on the topographic slope and  $\beta$ -effect. The  
 459 previous choices for  $dL$  corresponds to slopes  $\tan\alpha = \infty, 2 \cdot 10^{-2}, 10^{-2}, 5 \cdot 10^{-3}$  respectively.

460 Figure 16 shows upper and lower layer *PVA* maps at  $t = 42$  days for the different margin  
 461 lengths. As could be expected, the influence of  $dL$  is similar to  $L_x$  : above a critical value,

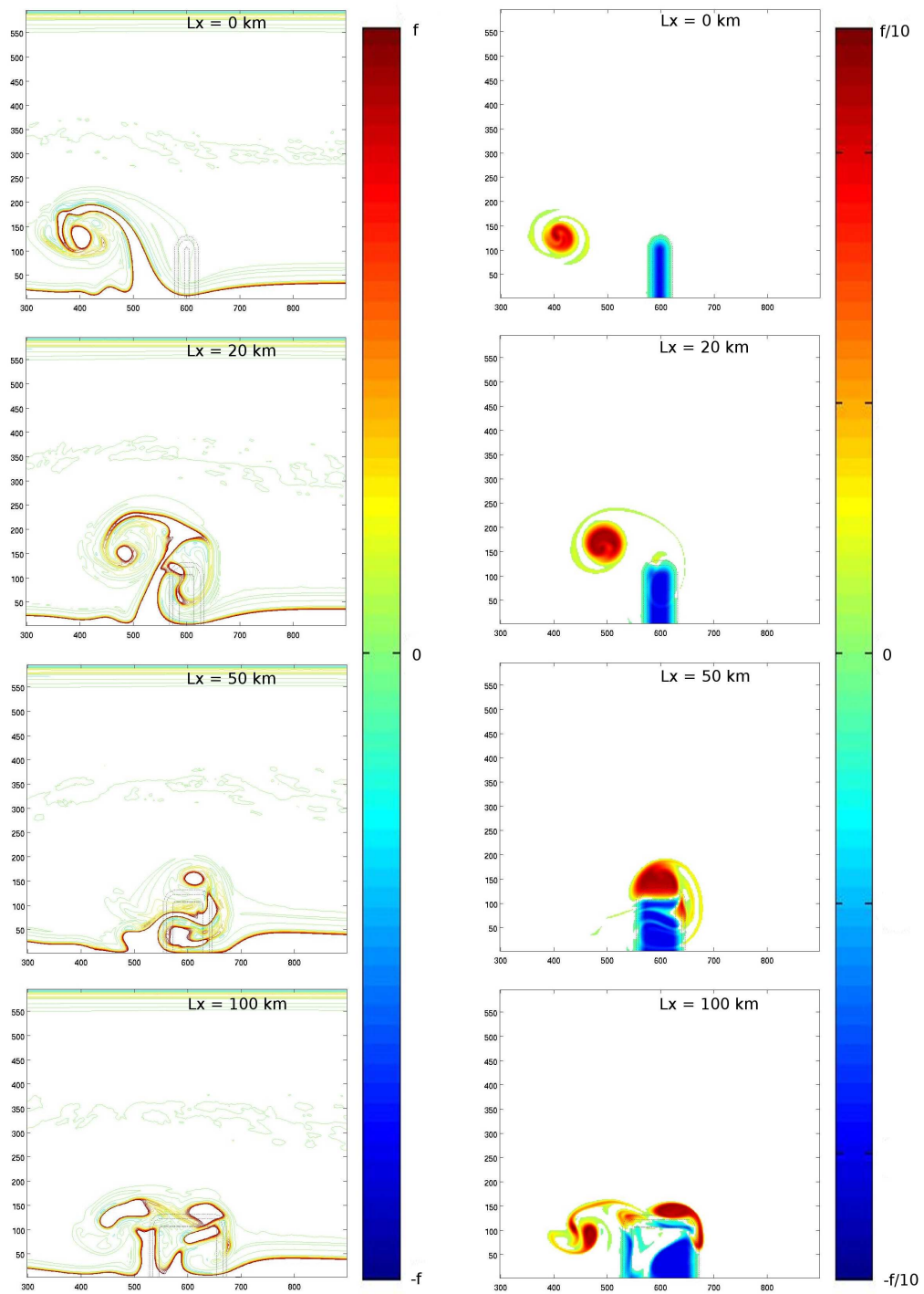


Figure 14: Maps of  $PVA$  in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $L_x = 0, 20, 50$  and  $100\text{km}$  experiments.

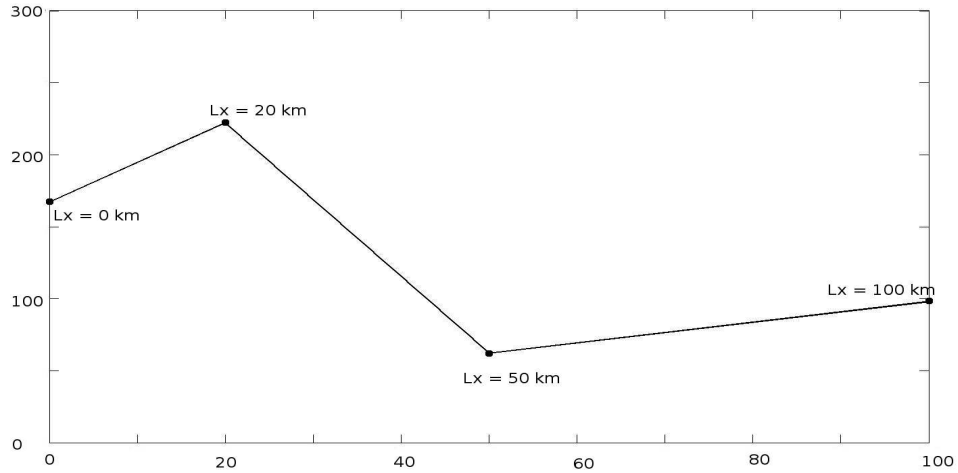


Figure 15: Offshore extent of the filament after 50 days of experiment for the  $L_x = 0, 20, 50$  and  $100\text{km}$  cases. The x axis is  $L_x$  (kms) and the y axis is offshore distance (kms).

462 the total circulation is dominated by the negative PVA pole and the filament wraps around the  
 463 promontory. Despite very different values for the topographic slope and  $\beta$ -effect, the similarities  
 464 between Fig. 16 and 14 are striking.

465 This confirms that the topographic  $\beta$ -effect has a minor impact on the dynamics, and that the  
 466 important parameter in the generation of coherent and trapped filaments is the total amount of  
 467 PVA over the promontory.

#### 468 4.7. The promontory length

469 As discussed above, anisotropy in the shape of the promontory can also modify the structure  
 470 and strength of the topographic circulation. The sensitivity of the results to the promontory length  
 471  $L_y$  was thus studied with  $L_y = 50$  km,  $L_y = 100$  km (reference experiment),  $L_y = 150$  km and  $L_y$   
 472  $= 200$  km.

473 Figure 17 shows the structure of the PVA after 42 days in both layers and for the different  
 474  $L_y$ . The offshore extension of the bottom layer negative PVA pole obviously follows  $L_y$  and  
 475 also drives the length of the filament which always extends further than the promontory. The  
 476  $L_y = 200\text{km}$  case shows that there exists a maximum length of the filament over which it breaks,  
 477 so that very long promontories are not necessarily the most efficient ones. This is underlined in  
 478 Fig. 18 which shows the maximum offshore extension of the filament as a function of  $L_y$ . The  
 479 optimal value is here around 150 km. Another particular feature for long promontories ( $L_y = 200$   
 480 km), is that the filament no longer rolls up around the positive PVA pole as it drifts far offshore,  
 481 its base is much wider and its offshore shape much thinner.

#### 482 4.8. Bottom friction

483 Two experiments were performed adding a bottom friction term to the reference configuration.  
 484 Figure 19 shows the PVA in the upper layer (left hand panel) and the PV in the bottom layer at

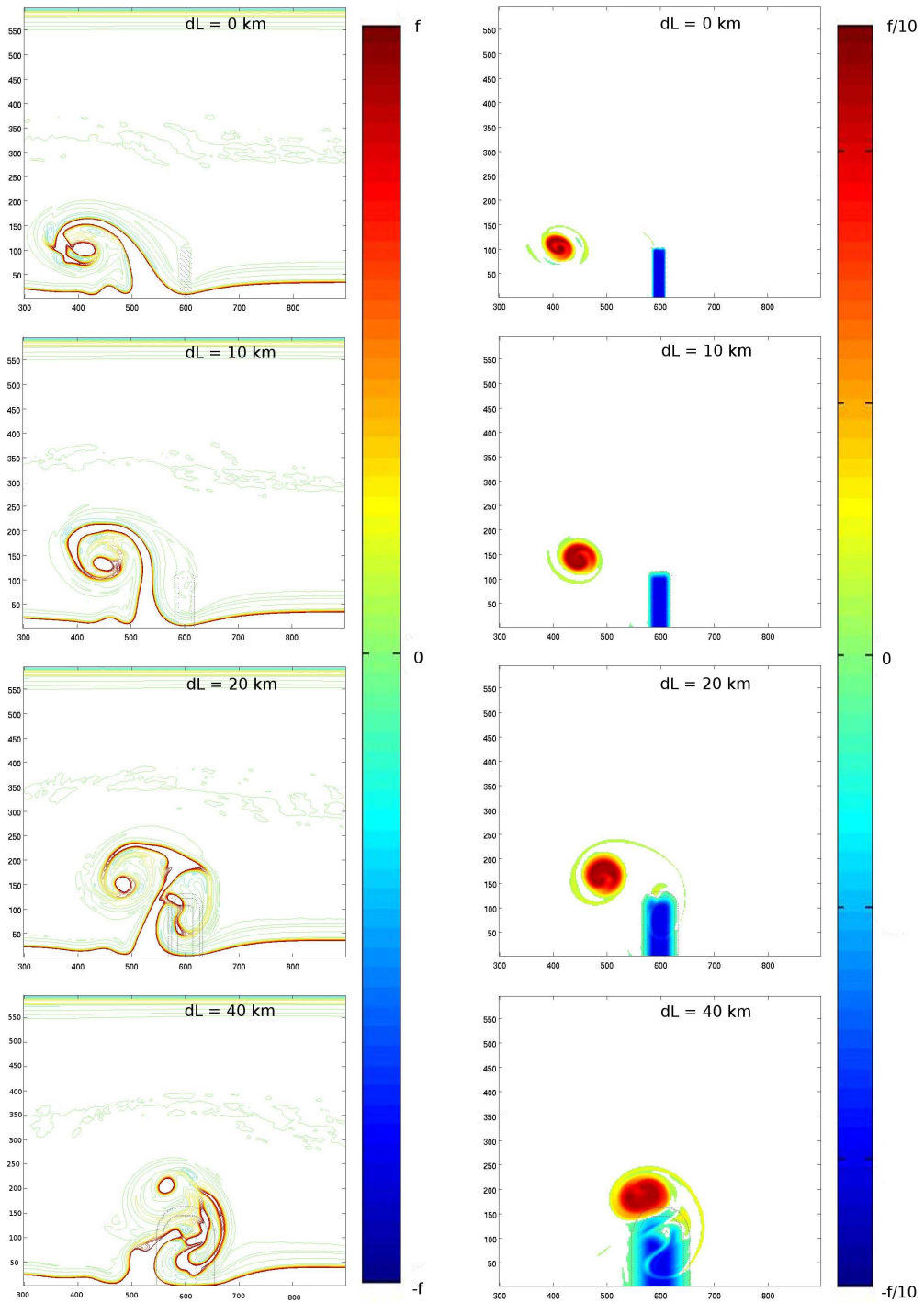


Figure 16: Maps of *PVA* in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $dL = 0, 20, 50$  and  $100\text{km}$  experiments.

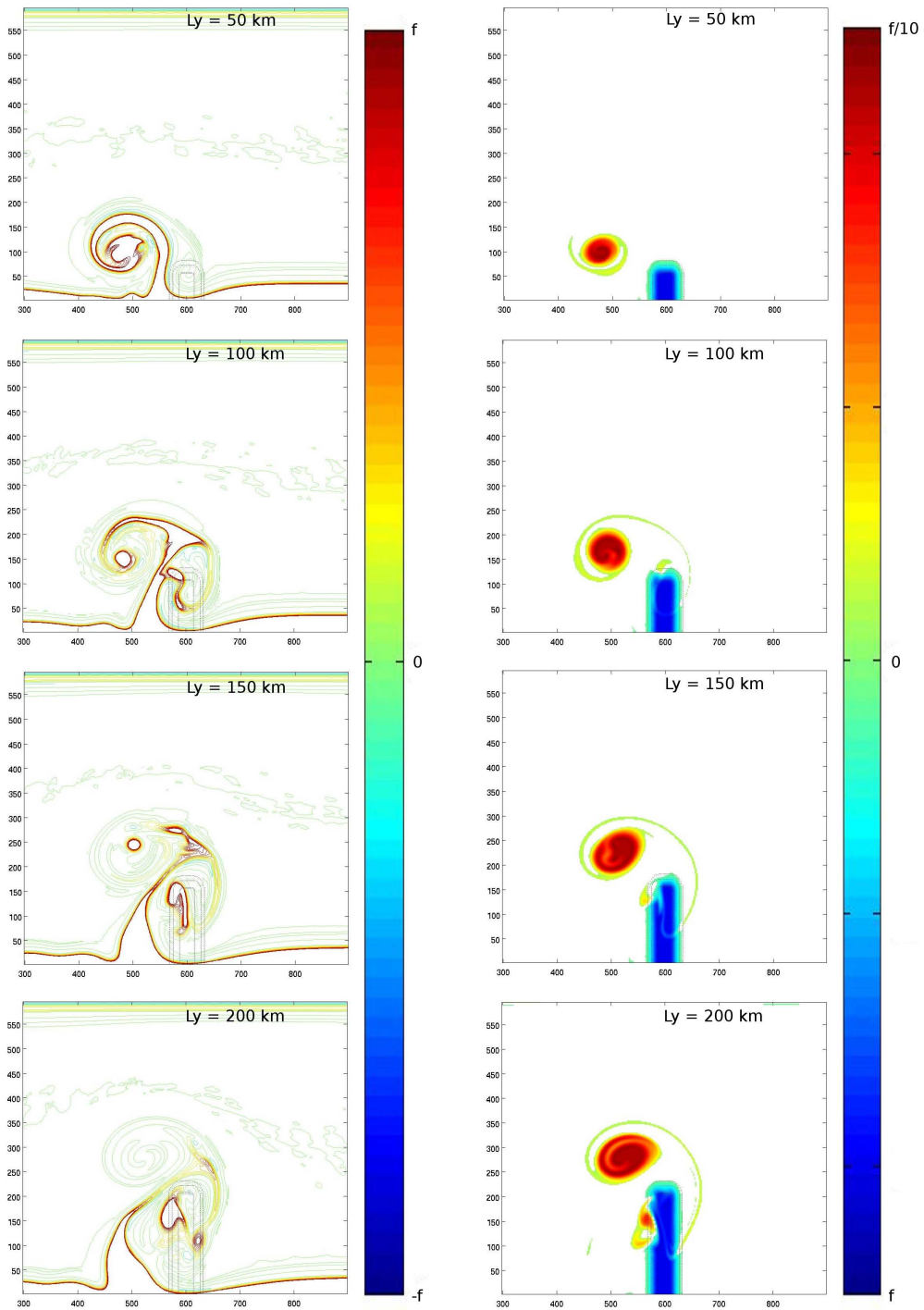


Figure 17: Maps of *PVA* in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $Ly = 50, 100, 150$  and  $200\text{km}$  experiments.

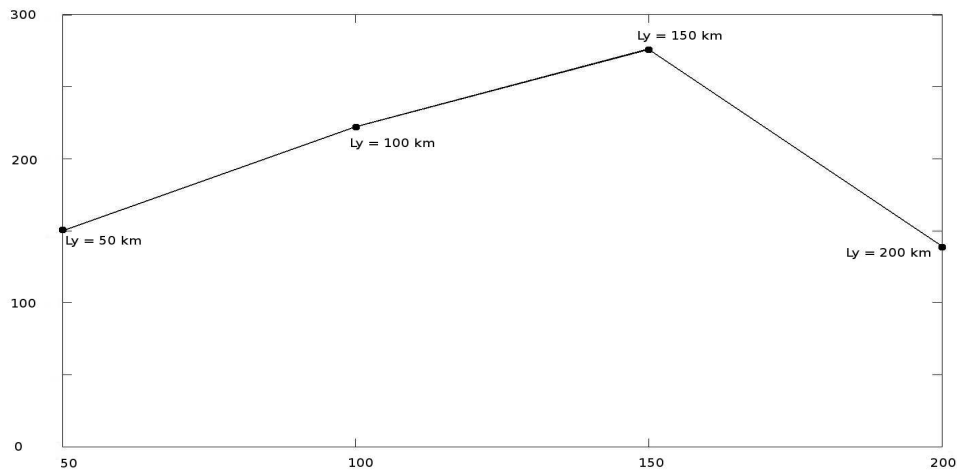


Figure 18: Offshore extent of the filament after 50 days of experiment for the  $L_y = 50, 100, 150$  and  $200\text{km}$  cases cases. The x axis is  $L_y$  (kms) and the y axis is offshore distance (kms).

485  $t = 42\text{days}$  (right hand panel) for the  $C_d = 0, 3 \cdot 10^{-3}$ , and  $5 \cdot 10^{-3} \text{Nm}^{-2}$  cases. The formation  
 486 and evolution of the filament in the upper layer is unchanged by the addition of the bottom  
 487 friction. The effect of friction is only visible in the  $PV$  of the bottom layer: As the positive  $PV$   
 488 reservoir that has left the promontory to form the positive  $PVA$  pole gets eroded, new higher  $PV$   
 489 is generated over the promontory, resulting in weakening the negative  $PVA$  pole.

490 As a result, bottom friction leads to a relaxation of the circulation and to a new state of rest,  
 491 which allows new topographic eddies and front filamentation to occur if the wind starts blowing  
 492 again.

#### 493 4.9. Influence of other topographic features

#### 494 4.10. Capes

495 Complementary experiments including a cape or a cape superimposed on a larger promon-  
 496 tory and a canyon where performed in order to compare the impact of the coastline geometry  
 497 with the topographic process proposed here. Upwelling filament dynamics have sometimes been  
 498 associated with capes triggering ([Strub et al., 1991]), but most capes have large promontory-  
 499 like undersea extends like Cabo Roca, Cabo Finisterre and the Estremadura promontory on the  
 500 western Iberian coast.

501 The upper two panels in Fig. 20 show the  $PVA$  in the upper (left hand side) and bottom  
 502 (right hand side) layer for two sizes of capes, both having the same Gaussian shape. Viscosity  
 503 induces increased diabatic effects near boundaries which results here in the development of a  
 504 layer of positive  $PVA$ . The presence of a cape allows this  $PVA$  to detach from the coast and  
 505 wrap into a positive  $PVA$  pole. This process generates a cyclonic vortex downstream of the cape  
 506 that in turn induces an offshore displacement of the outcropped front for small capes (cape 1  
 507 is 50 km long and 25 km wide), and in the generation of a thin filament parallel to the coast

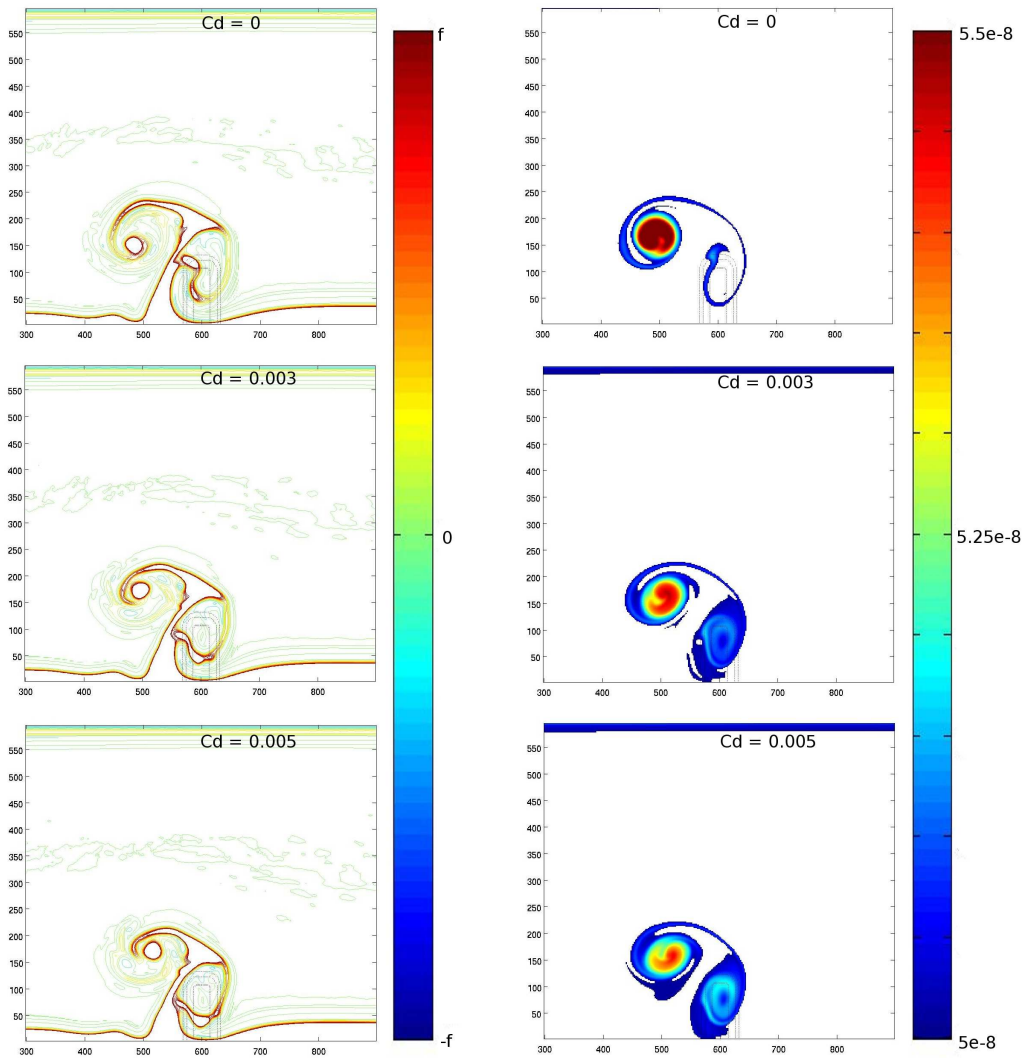


Figure 19: Maps of  $PVA$  in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the  $C_d = 0, 3 \cdot 10^{-3}$  and  $5 \cdot 10^{-3} Nm^{-2}$  experiments.

508 for the larger capes (cape 2 is 100 km long and 100 km wide). Superimposing the first cape  
509 with the reference promontory (third panel) leads to a long and thin filament pointing offshore  
510 and developing downstream of the cape. This combination of cape and promontory seems to be  
511 particularly efficient to generate the filament, because the cape alters the anticyclonic circulation  
512 on the promontory, so that the filament keeps on growing offshore instead of rolling up around  
513 the promontory.

#### 514 4.11. Canyons

515 The lower panels in Fig. 20 show the effect of a canyon having the same shape as the reference  
516 promontory. Such a configuration generates a cyclonic circulation above the promontory and an  
517 anticyclonic one downstream. As a result, the jetlike current between both eddies is now directed  
518 shoreward and can not generate a filament. However, the cyclonic circulation induces an offshore  
519 current upstream of the canyon that, even though associated with a much weaker current than the  
520 promontory case, can advect the upwelling front offshore. This is observed on Fig. 20 where a  
521 filament forms in the upper layer upstream of the canyon. Also notice that the upwelling front  
522 disappeared downstream of the promontory. There probably also exists an optimal shape of the  
523 canyon for the development of the filament, but this is beyond the scope of the present study and  
524 we simply conclude that canyons can also play an important role in the formation of trapped and  
525 long upwelling filaments.

#### 526 4.12. Combination of canyons and promontories

527 In many upwelling systems, the topography is not as simple as a promontory or a  
528 canyon surrounded by a large flat bottom area. The bottom topography off the Rias Baixas, North  
529 of the Iberian upwelling system is a good example of a succession of canyons and promontories.  
530 In that case, a more complex dynamics is expected. To explore the impact of such a combi-  
531 nation on the filament development and trapping, we set up two configurations : One with a  
532 canyon downstream of the promontory and the other with a canyon upstream of the promon-  
533 tory. Both are tested with two values of promontory/canyon height/depth. Figure 21 shows the  
534 *PVA* in layer 1 (left hand side panels) and in layer 2 (right hand side panels) for a  $H_t = 200m$   
535 promontory/canyon combination (panel 1),  $H_t = 100m$  promontory/canyon combination (panel  
536 2),  $H_t = 200m$  canyon/promontory combination (panel 3) and  $H_t = 100m$  canyon/promontory  
537 combination (panel 4). In all 4 cases, the bottom layer *PVA* dynamics is slightly complicated be-  
538 cause of the generation of 4 *PVA* poles, but the main effects associated with the effect of isolated  
539 topographic features are still visible : as the water initially situated over the flat bottom crosses  
540 the promontory (canyon), a negative (positive) *PVA* is generated, while a positive (negative)  
541 *PVA* is generated as the water from the promontory(canyon) crosses the canyon (promontory).  
542 Another negative (positive) *PVA* is formed when the water originally located above the canyon  
543 reaches the flat bottom area. Once the water columns originally situated upstream of the promon-  
544 tory (canyon) have completely crossed the topography and reached the downstream flat bottom  
545 area, the *PVA* distribution is as follow : 2 trapped opposite sign *PVA* over the promontory and  
546 the canyon, and two free opposite sign *PVA* downstream of the topography, evolving as a dipole.  
547 Note that in the  $H_t = 200m$  cases, there is a slight multipolarisation because of stronger non  
548 linear interaction between the higher *PVA* poles, while in the  $H_t = 100m$  case, there is a stronger  
549 advection of the free *PVA* poles.

550 Even though both canyon/promontory and promontory/canyon combinations produce long fil-  
551 aments, there is a dynamical difference between them : In the promontory/canyon case, the most



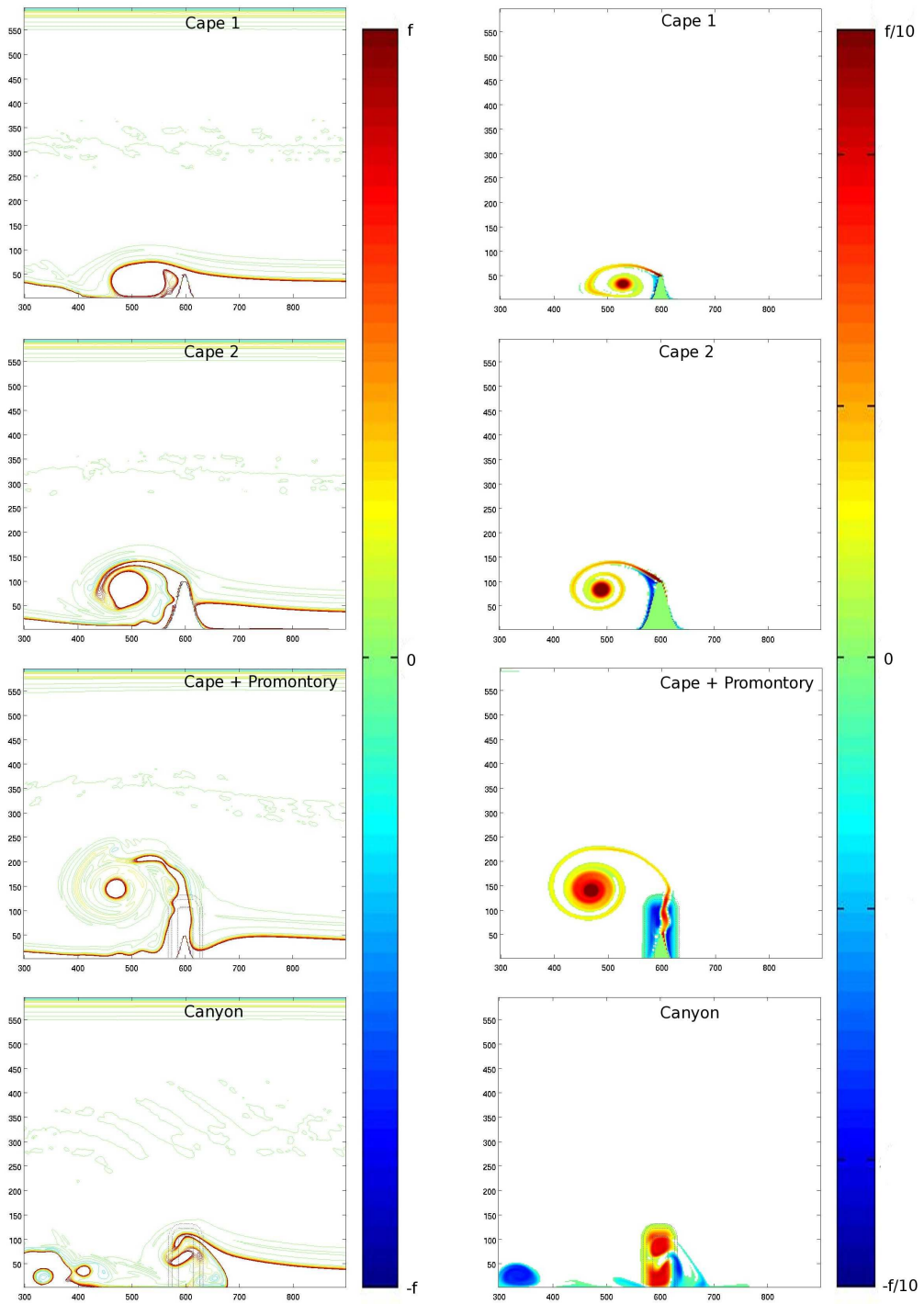


Figure 20: Maps of  $PVA$  in the upper (left hand panel) and bottom (right hand panel) layers at  $t = 42$  days for the two capes experiments (panels 1 and 2), the cape over the promontory experiment (panel 3) and the canyon experiment (panel 4).

552 upstream *PVA* pole is the trapped negative one associated with the promontory which results  
553 in an offshore deviation of the front, while in the canyon/promontory combination, the trapped  
554 positive *PVA* associated with the canyon entrain the front inshore, inhibiting a proper filamen-  
555 tation over the topography. This results in a major difference in the nature of the observed  
556 filaments : while the filament of the promontory/canyon combination is generated by the effect  
557 of the trapped dipole resulting in a permanently trapped filament between the promontory and  
558 the canyon, the filament generated by the canyon/promontory combination is generated by the  
559 effect of the free dipole resulting in a free filament, moving downstream as it gets advected by  
560 the barotropic current.

561 As shown by these simple examples, the combinations of canyons and promontories with  
562 complex shapes, as existing in nature, may lead to a more drastic selection of regions for the  
563 generation of long and trapped filaments.!!!!?YM

## 564 **5. Conclusion**

### 565 *5.1. Generation of long trapped filament by topographic effect*

566 In this paper, we have studied the formation of long trapped upwelling filaments which are  
567 ubiquitous features in all major upwelling systems. We focused on the effect of bottom topogra-  
568 phy using an idealized two-layer configuration with a wind forcing that generates an upwelling  
569 front along a vertical coastal wall in the presence of a transverse promontory or ridge. The phys-  
570 ical process studied is based on the formation of *PVA* when the upwelling current interacts with  
571 the bottom topography. At rest, in the bottom layer, the promontory is a positive potential vortic-  
572 ity anomaly pool, because the ambient potential vorticity is stronger over the promontory than in  
573 the rest of the domain. As the upwelling current sets up, this high potential vorticity gets advected  
574 downstream of the promontory by the alongshore current, and is replaced by lower potential vortic-  
575 ity water, generating a negative potential vorticity anomaly (*PVA*) over the promontory, and a  
576 positive potential vorticity anomaly downstream. The positive *PVA* is advected offshore by the  
577 trapped negative pole current field, and evolves into a cyclonic vortex eventually advected down-  
578 stream. The negative *PVA* induces an anticyclonic circulation anomaly that remains trapped  
579 above the topography and, if it is strong enough, will modify the mean upwelling current. This  
580 produces a geostrophic offshore flow on the downstream side of the promontory which is able to  
581 distort the upwelling front and then forms a meander. The latter finally evolves into a thin fila-  
582 ments that grows offshore, and that may be sometimes slightly rolling up around the topographic  
583 eddies. The barotropic part of the circulation plays the main role in the latter process.

584 Then, we evaluated the influence of baroclinic instability on the formation of the meanders  
585 generated by topographic effects using stable and unstable configurations. The modelled stable  
586 current is able to generate a very large and stationary filament when interacting with topog-  
587 raphy whereas baroclinic instabilities of the jet produce additional numerous shorter me-  
588 anders quickly propagating downstream that have a weak influence on the topographic fila-  
589 ment development. The characteristics of the meanders and filaments associated with baro-  
590 clinic instability are sensitive to the stratification and, as found in some previous studies (see  
591 [*Haidvogel et al.*, 1991, *Strub et al.*, 1991]), long filament can also emerge in this case, but the  
592 trapping of the filament and their development at identical locations can only be explained by  
593 topographic effects.

594 We have also found that the deviation of the upwelling current by a cape and the generation of  
595 vorticity by the viscous boundary layer generate a cyclonic pool of cold water downstream of the

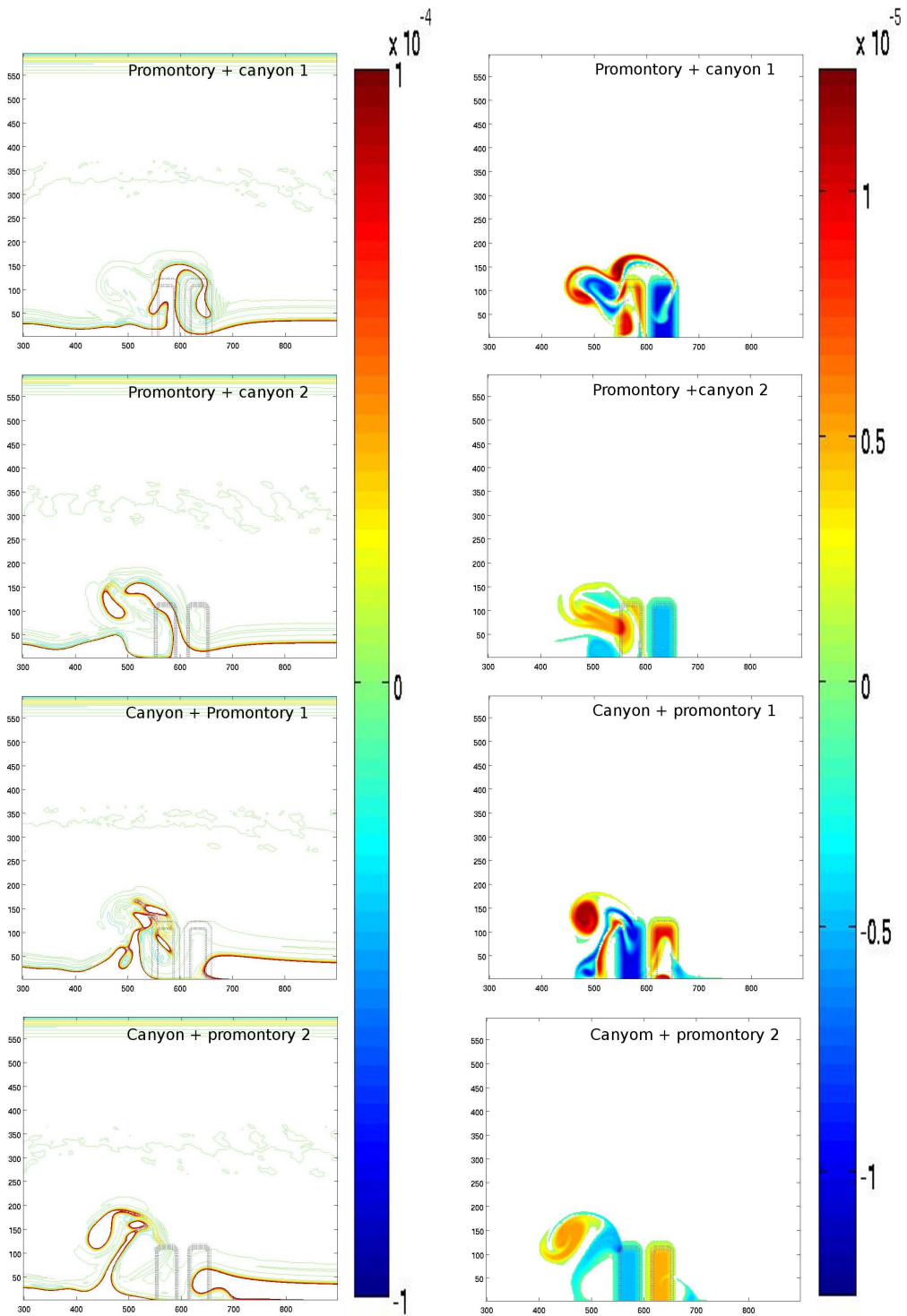


Figure 21: Maps of  $PVA$  in layer 1 (left hand side panels) and in layer 2 (right hand side panels) for a  $H_t = 200m$  promontory/canyon combination (panel 1),  $H_t = 100m$  promontory/canyon combination (panel 2),  $H_t = 200m$  canyon/promontory combination (panel 3) and  $H_t = 100m$  canyon/promontory combination (panel 4).

596 cape rather than a long offshore filament. Thus, at least for the simplified configurations used in  
597 the present paper, the process involved in the generation of upwelling filaments at capes is rather  
598 associated with the submarine promontories that generally exist in their continuity, modifying  
599 the potential vorticity structure of the current as explained above.

600 The topographic eddies are generated by the barotropic circulation associated with the up-  
601 welling development. Their advective effect on the upwelling front as well as the formation of  
602 filaments are again associated with their barotropic vorticity. The process thus mainly involves  
603 the barotropic circulation. As a consequence, modifying the stratification does not strongly mod-  
604 ify the mechanism we have identified, nor the generation of topographic filaments as long as the  
605 upwelling front is formed.

606 Increasing the duration of the wind forcing induces a further extension of the upwelling front,  
607 and intensifies the strength of the upwelling barotropic current but not of the topographic ed-  
608 dies. The cyclonic eddy detaching from the topography is thus advected further downstream  
609 which, together with the change in the upwelling front extension, modify the characteristics of  
610 the filament. In our case the filament width increases and bends downstream when applying  
611 wind forcing for a longer time. Let us note that increasing the duration of the wind forcing also  
612 increases the source of diabatic *PV* generation (see [Morel *et al.*, 2006]) and thus the instability  
613 of the current. This was not investigated in the present paper, but we expect that, as a result, the  
614 relative importance of the unstable structure increases as the forcing lasts longer, so that a greater  
615 part of the offshore transport can be attributed to the instability of the jet in this case. This should  
616 however not change our conclusions on the trapping of long upwelling filaments.

617 Sensitivity tests to the shape of the topographic feature have shown that the width, height,  
618 length and slope of the topography are important parameters for the process. Multipolarisation  
619 of the *PVA*, decreasing the coherence and the length of the surface filaments, can occur in the  
620 case of a too wide or too tall promontory. The height of the promontory controls the available  
621 *PVA* pool, so that a too small promontory can not produce a large stationary filament, but neither  
622 can a very tall promontory above which flows and *PVA* can barely form. There exists optimal  
623 values of the topography characteristics to maximize the offshore extension of the filament. In  
624 other words, the formation of long trapped upwelling filaments by topographic features is a  
625 selective process and is restricted to some height, width, slope and length ranges.

## 626 5.2. Discussion and perspectives

627 !!!!!YM Even though our model is of the simplest, it allows to clearly identify a dynamical  
628 process which seems relevant to explain the stationarity and the repeatability of upwelling fila-  
629 ments at some particular locations. This simplicity however makes a thorough comparison with  
630 observed current structures a difficult task, but a qualitative discussion on some general patterns  
631 remains possible.

632 It is important to note that the various upwelling systems develop over very different topogra-  
633 phies : if the Iberian or the Chilean margins show complicated patterns with a succession of  
634 ridges and canyons in the continuation of the rias, the North African topography is a combination  
635 of a smooth margin with a few well defined promontories (Cape Ghir, Cape Blanc). As shown  
636 in section 4.4 and 4.9, the generation of trapped filaments is highly sensitive to the topographic  
637 configuration, and the multiplicity and greater variability of upwelling filaments along the West-  
638 Iberian shelf could be explained by this more complex bottom topography. If the isolated Cape  
639 Ghir promontory is a scholar case for the generation of well defined trapped anticyclone, trig-  
640 gering a large long lived stationary filament, the multiple *PVA* poles generated over the complex  
641 Iberian topography must lead to non-linear interactions, including fragmentation and merger of

642 the *PVA* poles leading to the observed shorter-lived, less coherent filament with higher spatial  
643 and temporal variability.

644 ???YM ATTENTION, je pense que les conditions atmospheriques peuvent aussi pas mal jouer  
645 : les vents sur la faade iberique sont assez inconstants et expliquent sans doute aussi pas mal la  
646 variabilite observee. ???YM

647 Note also that the present study does not exclude the processes invoked by previous studies to  
648 explain the generation of upwelling filaments. Indeed, topographically trapped eddies can com-  
649 bine or interact with other filamentation processes. The Southern North-West African upwelling  
650 system, offshore of Mauritania is a good example of the multiplicity of dynamics of the up-  
651 welling filaments : Recent satellite and in-situ observations (SOLAS-ICON cruise, unpublished  
652 yet) showed a whole range of filaments with different sizes, shapes and behaviour. The intrinsic  
653 instability of the upwelling front generates pinched off meanders, developing and propagating all  
654 along the front between the Arguin's bank and Cape Verde whereas a quasi permanent filament  
655 associated with an anticyclonic eddy is found over the Cape Blanc promontory. Even though  
656 this filament is quasi-permanently rooted over the Cap Blanc promontory, its offshore extension  
657 shows a high spatial and temporal variability. This variability is believed to result from interac-  
658 tions with an external mesoscale turbulence field resulting from the presence of the baroclinically  
659 unstable Cape Verde front in the vicinity of the Mauritanian upwelling system. This confirms the  
660 idea of [Strub *et al.*, 1991] that filaments could result of the combined effect of various processes  
661 like jet instability and the interaction with an offshore eddy field. The role of bottom topography  
662 is finally added to justify the stationarity and repeatability of filaments. ???YM ATTENTION: je  
663 trouve que ce dernier paragraphe brouille le message. Il faut que le lecteur reparte avec comme  
664 idee claire que jusqu' present seul le mecanisme propos ici permet d'expliquer la presence de long  
665 filament pigs se developpant systmatiquement au mme endroit. D'autres mecanismes peuvent  
666 soit gnrer des filaments (mais a priori ayant des caracteristiques en terme de stationnarite/pigeage  
667 different) soit interagir avec le present effet pour en diminuer ou en augmenter les effets. Je suggere  
668 de recrire ce paragraphe sur la base de ce fil conducteur. Tel quel ce paragraphe remet tout zro  
669 mon sens : on comprend que finalement d'autre mecanisme peuvent trs bien expliquer le mme  
670 comportement sans qu'on donne d'explication cela. ???YM

671 !!!??YM

672 The two-layer adiabatic model we have used is however very simplified and in nature, many  
673 other parameters and many additional processes, that have not been studied here, can influence  
674 the development of topographic eddies and formation of upwelling filaments, among which :

- 675 • the existence of an extended shelf and a continental slope ;
- 676 • the existence of a more complex and realistic large scale circulation (presence of a deep  
677 poleward undercurrent) associated with or preceding the upwelling development;
- 678 • the influence of a more complicated stratification and of the specific dynamics of the mixed  
679 layer or the influence of mixing in general;
- 680 • the influence of the planetary  $\beta$  effect;
- 681 • the influence of bottom friction.

682 The beta effect seems of particular importance as it strongly influences the dynamics of eddies  
683 and induces westward propagation. We can thus imagine, for Eastern boundary upwellings,  
684 that any localized source of vortices such as bottom topography, but also capes, could favor the

685 offshore development of filaments : indeed the cyclonic vortex developing downstream of a cape  
686 would be advected westward entraining upwelled cold waters and forming a localized filament  
687 instead of a circular patch as observed in the present study.

688 Another interesting subject is the influence of bottom friction. We have found here that its  
689 influence was weak, but this was expected since we considered a deep ocean. Bottom friction  
690 would obviously plays a stronger role than observed here for upwelling developing above shallow  
691 areas (corresponding to extended shelves). As its effect is to reduce the currents in the bottom  
692 layers, we expect this would limit the strength of the upwelling barotropic velocity and down-  
693 stream transport. However, meanwhile it would also renew the reservoir of positive PVA above  
694 the promontory. As a result, after a period of relaxation of the winds, despite the fact that the  
695 initial positive PVA moved away from the topography, the same mechanism could be repeated  
696 for a new upwelling event.

697 However, even though these processes can have a strong influences on the result and their effect  
698 is worth investigating, we believe that they would not substantially modify our main result : the  
699 bottom topography plays a key role on the formation of the long upwelling filaments, whatever  
700 the direction of the current and details of the topography, PVA has to form when the flow passes  
701 over topography, developing a trapped circulation whose signature extends over the entire water  
702 column and influences the upper layer dynamics advecting water parcels offshore.

703 Finally, more in-situ observations of early stage development of upwelling filaments are  
704 needed to possibly confirm our results and the main influence of the barotropic circulation associ-  
705 ated with topographic eddies. This implies an extended mapping of the circulation, !!! including  
706 the deep mesoscale features which are often neglected during the mesoscale surveys because of  
707 the difficulty to sample at great depths without loosing time and thus synopticity. !!!

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714 **References**

- 715 [Alvarez-Salgado *et al.*, 2001] Alvarez-Salgado, X.A., M.D. Doval, A.V. Borges, I. Joint, M. Frankignoulle, E.M.S.  
716 Woodward, F.G. Figueiras, Off-shelf fluxes of labile materials by an upwelling filament in the NW Iberian Up-  
717 welling System. *Progress in Oceanography*, 51, 321-337, 2001.
- 718 [Alvarez-Salgado *et al.*, 2007] Alvarez-Salgado, X.A., Arstegui, J., Barton, E.D., Hansell, D.A., Contribution of up-  
719 welling filaments to offshore carbon export in the subtropical Northeast Atlantic Ocean. *Limnology and Oceanog-*  
720 *raphy*, 52, 1287-1292, 2007.
- 721 [Bang and Andrews, 1974] Bang, N.D., and Andrews, W.R.H., Direct current measurements of a shelf-edge frontal jet  
722 in the southern Benguela system, *Journal of Marine Research*, 32, 407-421, 1974.
- 723 [Barth, 1989 a] Barth, J.A. Stability of a coastal upwelling front 1. Model developing and a stability theorem, *Journal*  
724 *of Geophysical Research*, Vol. 94, 10844-10856, 1989.
- 725 [Barth, 1989 b] Barth, J.A. Stability of a coastal upwelling front 2. Model results and comparison with observations,  
726 *Journal of Geophysical Research*, Vol. 94, 10857-10883., 1989.
- 727 [Barth, 1994] Barth, J.A. Short-wavelength instabilities on coastal jets and fronts, *Journal of Geophysical Research*,  
728 Vol. 98, 16095-16115, 1994.
- 729 [Barton, 2001] Barton, E.D., M.E Inall, T.J, Sherwin and R. Torres., Vertical structure, turbulent mixing and fluxes  
730 during Lagrangian observations of an upwelling filament system off Northwest Iberia, *Progress in Oceanography*,  
731 Vol. 51, 249-267, 2001.
- 732 [Barton, 2004] Barton, E.D., Aristegui, J., Tett, T. and Navarro-Pérez, E., Variability in the Canary Islands Area of  
733 Filament-Eddy Exchanges, *Progress in Oceanography*, Vol. 62, 71-94, 2004.
- 734 [Batteen, 1997] Batteen, M.L., Wind-forced modeling studies of currents, meanders, and eddies in the California Cur-  
735 rent system *Journal of Geophysical Research-Oceans*, Vol 102, C1, 985-1010, 1997.
- 736 [Batteen *et al.*, 2007] Batteen, M.L., Martinho, A.S., Miller, H.A. and McClean, J.L., A Process-Oriented Study of the  
737 Coastal Canary and Iberian Current System, *Ocean Modelling*, vol. 18, 1-36, 2007.
- 738 [Bleck and Boudra, 1986] Bleck., R. and Boudra, D., Wind driven spin-up in eddy-resolving ocean models formulated  
739 in isopycnic and isobaric coordinates, *Journal of Geophysical Research*, vol. 91, 7611-7621, 1986.
- 740 [Bleck and Smith, 1990] Bleck., R. and Smith, L., A wind driven isopycnic coordinate model of the North and equatorial  
741 Atlantic Ocean: Model development and supporting experiments *Journal of Geophysical Research*, vol. 95, 3273-  
742 3285, 1990.
- 743 [Bleck *et al.*, 1992] Bleck., R., Rooth, C., Hu, D., and Smith, L., 1992. Salinity driven thermocline transients in a wind  
744 and thermocline forced isopycnic coordinate model of the North Atlantic. *Journal of Physical Oceanography*, vol.  
745 22, 1486-1505.
- 746 [Brink, 1983] Brink, K.H., The near-surface dynamics of coastal upwelling, *Progress in Oceanography*, vol. 12, 223-  
747 257, 1983.
- 748 [Bretherton, 1966] Bretherton, F.P., Critical layer instability in baroclinic flows, *The Quarterly Journal of the Royal*  
749 *Meteorological Society*, vol. 92, 325-334, 1966.
- 750 [Capet and Carton, 2004] Capet, X.J. and Carton, X.J. Nonlinear Regimes of Baroclinic Boundary Currents, *Journal*  
751 *of Physical Oceanography*, vol. 34, 14001409, 2004.
- 752 [Charney and Stern, 1962] Charney, J.G. and Stern, M.E., On the Stability of Internal Baroclinic Jets in a Rotating  
753 Atmosphere, *Journal of the Atmospheric Sciences*, vol. 19, 159172, 1962.
- 754 [Dewey *et al.*, 1991] Dewey, R.K., Moum, J.N., Paulson, C.A., Caldwell, D.R. and Pierce, S.D., Structure and Dynamics  
755 of a Coastal Filament, *Journal of Geophysical Research*, vol. 96, 14885-14907, 1991.
- 756 [Flament *et al.*, 1985] Flament, P., Armi, L. and Washburn, L., The Evolving Structure of an Upwelling Filament,  
757 *Journal of Geophysical Research*, vol. 90, 11765-11778, 1985.
- 758 [Garvine, 1971] Garvine, R.W., A simple model of coastal upwelling dynamics, *Journal of Physical Oceanography*, 1,  
759 169-179, 1971.
- 760 [Garvine, 1973] Garvine, R.W., The effects of bathymetry on the coastal upwelling of homogeneous water, *Journal of*  
761 *Physical Oceanography*, 3, 47-56, 1973.
- 762 [Gill and Clarke, 1974] Gill, A.E. and A.J., Clarke, Wind-induced upwelling, coastal currents and sea level changes,  
763 *Deep-Sea Research*, vol. 21, 325-345, 1974.
- 764 [Haidvogel *et al.*, 1991] Haidvogel, D.B., Beckmann, A. and HedStrm, K.S., Dynamical Simulations of Filament For-  
765 mation and Evolution in the Coastal Transition Zone, *Journal of Geophysical Research*, vol. 96, 15017-15040,  
766 1991.
- 767 [Haynes *et al.*, 1993] Haynes, R., Barton, E.D. and Pilling, I. Development, Persistence, and Variability of Upwelling  
768 Filaments, *Journal of Geophysical Research*, vol. 98, 22681-22692, 1993.
- 769 [Herbette *et al.*, 2003] Herbette, S., Morel, Y.G. and Arhan, M., Erosion of a surface vortex by a seamount, *Journal of*  
770 *Physical Oceanography*, vol. 33, (8):1664-1679, 2003.

- 771 [Herbette *et al.*, 2005] Herbette, S., Morel, Y.G. and Arhan, M., Erosion of a surface vortex by a seamount on the beta  
772 plane, *Journal of Physical Oceanography*, vol. 35, (11):2012-2030, 2005.
- 773 [Hoskins *et al.*, 1985] Hoskins, B., McIntyre, M. and W. Robertson, On the use and significance of isentropic potential  
774 vorticity maps, *Quarterly Journal of the Royal Meteorological Society*, vol. 111, 877-946, 1985.
- 775 [Ikeda, 1981] Ikeda, M., Meanders and Detached Eddies of a Strong Eastward-Flowing Jet Using a Two-Layer Quai-  
776 Geostrophic Model, *Journal of Physical Oceanography*, vol. 11, 526-540, 1981.
- 777 [Ikeda, 1989] Ikeda, M., Lygre, K. and Sandven, S., A Process Study of Mesoscale Meanders and Eddies in the Norwe-  
778 gian Coastal Current, *Journal of Physical Oceanography*, vol. 19, 20-35 1989.
- 779 [Killworth, 1980] Killworth, P.D., Barotropic and baroclinic instability in rotating stratified fluids, *Dynamics of Atmo-  
780 sphere and Ocean*, vol. 4, 143-184, 1980.
- 781 [Kostianoy and Zatsepin, 1996] Kostianoy, A.G. and Zatsepin, A.G., The West African coastal upwelling filaments and  
782 cross-frontal water exchange conditioned by them, *Journal of Marine Systems*, vol. 7, 349-359, 1996.
- 783 [Lentz and Chapman, 2004] Lentz, S.J. and D.C., Chapman, The importance of nonlinear cross-shelf momentum flux  
784 during wind-driven coastal upwelling, *Journal of Physical Oceanography*, 34, 2444-2457, 2004.
- 785 [Marchesiello *et al.*, 2003] Marchesiello, P., J. C. McWilliams and A. Shchepetkin, Equilibrium structure and dynam-  
786 ics of the California Current System. *J. Phys. Oceanogr.*, 33, 753-783, 2003.
- 787 [McIntyre and Norton, 1990] McIntyre, M., and W. Norton, Dissipative wave-mean interactions and the transport of  
788 vorticity or potential vorticity, *Journal of Fluid Mechanics*, vol. 212, 403-435, 1990.
- 789 [McWilliams and Gent, 1980] McWilliams, J.C., and Gent, P.R., 1980. Intermediate models of planetary circulations in  
790 the atmosphere and the ocean. *Journal of Atmospheric Sciences*, vol. 37, 1657-1678.
- 791 [Morel and McWilliams, 2001] Morel, Y. and McWilliams, J., Effects of Isopycnal and Diapycnal Mixing on the Sta-  
792 bility of Oceanic Currents, *Journal of Physical Oceanography*, vol. 31, 2280-2296, 2001.
- 793 [Morel *et al.*, 2006] Morel, Yves, G., Darr, D. S. and C. Talandier, Possible sources driving the Potential Vorticity struc-  
794 ture and long-wave instability of coastal upwelling and downwelling currents *Journal of Physical Oceanography*,  
795 vol. 36, 875-896, 2006.
- 796 [Morel and Thomas, 2009] Morel, Y. and Thomas, L.N., Ekman drift and vortical structures, *Ocean modelling*, vol. 27,  
797 185-197, 2009.
- 798 [Navarro-Pérez and Barton, 1998] Navarro-Pérez, E. and Barton, E.D., The Physical Structure of an Upwelling Fil-  
799 ament off the North-West African Coast during August 1993, *South African Journal of Marine Science*, vol. 19,  
800 61-73, 1998.
- 801 [O'Brien and Hurlburt, 1972] O'Brien, J.J. and Hurlburt, H.E., A numerical model of coastal upwelling, *Journal of*  
802 *Physical Oceanography*, vol. 2, 1972.
- 803 [Pringle, 2002] Pringle, J.M., Enhancement of wind-driven Upwelling and Downwelling by alongshore bathymetric  
804 variability, *Journal of Physical Oceanography*, 32, 3101-3112, 2002.
- 805 [Relvas *et al.*, 2007] Relvas, P., Barton, E.D., Dubert, J., Oliveira, P.B., Peliz, A., da Silva, J.C.B. and A.M.P. Santos,  
806 Physical oceanography of the western Iberia ecosystem: Latest views and challenges, *Progress in Oceanography*,  
807 74, 149-173, 2007.
- 808 [Rossi *et al.*, 2009] Rossi, V., Morel, Y. and Garçon, V., Effect of the wind on the shelf dynamics: Formation of a  
809 secondary upwelling along the continental margin, *Ocean Modelling*, 2009, doi : 10.1016/j.ocemod.2009.10.002.
- 810 [Roed and Shi, 1999] Roed, L. P., and X. B. Shi, A numerical study of the dynamics and energetics of cool filaments,  
811 jets and eddies off the Iberian Peninsula. *Journal of Geophysical Research*, 104(C12), 29817-29841, 1999.
- 812 [Sanchez *et al.*, 2008] R. F. Sanchez, P. Relvas, A. Martinho, and P. Miller, Physical description of an upwelling fil-  
813 ament west of Cape St. Vincent in late October 2004, *Journal of Geophysical Research*, Vol. 113, C07044,  
814 doi:10.1029/2007JC004430, 2008.
- 815 [Shi and Roed, 1999] X. B. Shi and Roed, L. P., Frontal Instabilities in a Two-Layer, Primitive Equation Ocean Model,  
816 *Journal of Physical Oceanography*, Vol. 29, 948-968, 1999.
- 817 [Stern and Chassignet, 2000] M.E. Stern and E.P. Chassignet, Mechanism of eddy separation from coastal currents ,  
818 *Journal of Marine Research*, Vol. 58, 269-295, 2000.
- 819 [Strub *et al.*, 1991] Strub, T.P., Kosro, P.M., and Huyer, A., The nature of the cold filaments in the California Current  
820 System, *Journal of Geophysical Research*, vol. 96, 14743-14768, 1991.
- 821 [Thomas, 2005] Thomas, L.N., Destruction of potential vorticity by winds, *J. Phys. Ocean.*, vol. 35, 2457-2466, 2005.
- 822 [Verron and Le Provost, 1985] Verron, J. and Le Provost, C., A numerical study of quasi-geostrophic flow over topog-  
823 raphy, *J. Fluid Mech.*, vol. 154, 231-252, 1985.
- 824 [Viera and Grimshaw, 1994] Viera, F. and Grimshaw, R., Topographic Forcing of Coastal Mesoscale Phenomena: Fil-  
825 amentation, Vortex Formation, and Eddy Detachment, *Journal of Physical Oceanography*, vol. 24, 1433-1448,  
826 1994.
- 827 [Washburn and Armi, 1988] Washburn, L. and Armi, L., Observations of Frontal Instabilities on an Upwelling Filament,  
828 *Journal of Physical Oceanography*, vol. 18, 1075-1092, 1988.
- 829 [Winther *et al.*, 2007] Winther, N.G., Morel, Y.G. and Evensen, G., Efficiency of high order numerical schemes for





$\delta t_t$	$\delta t_c$	$\delta x$	$N_x$	$N_y$	$N_{days}$	$\nu$	$T_x$	$f$	$H_2 + H_1$
5 s	200 s	2000 m	601 pts	301 pts	50 days	$0.3 \text{ m}^2 \text{ s}^{-1}$	$-0.2 \text{ Nm}^{-2}$	$10^{-4} \text{ s}^{-1}$	2050 m

Table 1: Table of the model parameters kept fixed in all experiments.

Experiment	$d\rho \text{ (kgm}^{-3}\text{)}$	$Cd$	$H_1 \text{ (m)}$	Forcing duration (days)	$H_t \text{ (m)}$	$L_x \text{ (km)}$	$L_y \text{ (km)}$	$dL \text{ (km)}$
Ref	1	0	50	10	200	20	100	20
Inst	1	0	50	10	200	20	100	20
Notopo frc 10	1	0	50	10	<u>0</u>	<u>0</u>	<u>0</u>	<u>0</u>
Notopo frc 20	1	0	50	<u>20</u>	<u>0</u>	<u>0</u>	<u>0</u>	<u>0</u>
Notopo frc 30	1	0	50	<u>30</u>	<u>0</u>	<u>0</u>	<u>0</u>	<u>0</u>
Notopo frc 40	1	0	50	<u>40</u>	<u>0</u>	<u>0</u>	<u>0</u>	<u>0</u>
Cd 3e-3	1	<u><math>3 \cdot 10^{-3}</math></u>	50	10	200	20	100	20
Cd 5 e-3	1	<u><math>5 \cdot 10^{-3}</math></u>	50	10	200	20	100	20
frc time 20	1	0	50	20	200	20	100	20
frc time 30	1	0	50	30	200	20	100	20
frc time 40	1	0	50	40	200	20	100	20
$H_t$ 50	1	0	50	10	<u>50</u>	20	100	20
$H_t$ 100	1	0	50	10	<u>100</u>	20	100	20
$H_t$ 300	1	0	50	10	<u>300</u>	20	100	20
$H_t$ 500	1	0	50	10	<u>500</u>	20	100	20
$H_t$ 1000	1	0	50	10	<u>1000</u>	20	100	20
$H_t$ 1500	1	0	50	10	<u>1500</u>	20	100	20
$L_x$ 0	1	0	50	10	200	<u>0</u>	100	20
$L_x$ 50	1	0	50	10	200	<u>50</u>	100	20
$L_x$ 100	1	0	50	10	200	<u>100</u>	100	20
$L_y$ 50	1	0	50	10	200	20	<u>50</u>	20
$L_y$ 150	1	0	50	10	200	20	<u>150</u>	20
$L_y$ 200	1	0	50	10	200	20	<u>200</u>	20
dL 0	1	0	50	10	200	20	100	<u>0</u>
dL 10	1	0	50	10	200	20	100	<u>10</u>
dL 40	1	0	50	10	200	20	100	<u>40</u>
$H_1$ 25	1	0	<u>25</u>	10	200	20	100	20
$H_1$ 100	1	0	<u>100</u>	10	200	20	100	20
$H_1$ 200	1	0	<u>200</u>	10	200	20	100	20
$d\rho$ 0.5	<u>0.5</u>	0	50	10	200	20	100	20
$d\rho$ 2	<u>2</u>	0	50	10	200	20	100	20
$d\rho$ 3	3	0	50	10	200	20	100	20

Table 2: Table of the model parameters that were varied in the various experiments.