Observations and Laboratory Modeling of Meddy Generation at Cape St. Vincent

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(Manuscript received 19 April 2000, in final form 13 March 2001)

ABSTRACT

The AMUSE, CANIGO, and AMPOR observations provided a large amount of subsurface float data showing evidence of several distinctly different flow behaviors within the Mediterranean Water layers in the vicinity of Cape St. Vincent (southwest Portugal). Some of the floats followed the coastline and moved northward, and some were trapped either in anticyclonic lenses of Mediterranean Water (meddies) generated at Cape St. Vincent or in cyclonic structures. These observations clearly indicated that the anticyclones were accompanied by cyclones, at least during their early stages. As the dipolar structure moves away from the cape, the cyclone tends to shift slowly away from the anticyclone while decaying so that after some time only the meddy persists. These in situ observations are compared with large-scale laboratory experiments and the agreement between the two sets of observations is remarkably good. Near the cape and shortly after the generation, the nondimensional radius, relative vorticity, and rotation period of the anticyclonic lenses observed in situ and in the model are in good agreement. Thus, the experiments and the 3D velocity measurements, which are possible in the test tank, provide useful details of the generation mechanisms and of the lens dynamics. They prove that the Rossby and Froude numbers of the (Mediterranean) current upstream of the cape are determining factors in predicting the behavior at the cape. For sufficiently low Rossby number (0.1 ≤ Ro ≤ 0.6) the laboratory measurements show that there is formation of a dipolar structure at the cape either by instability of the current at the cape or by coupling of the anticyclone created at the cape by the current and of a cyclone created upstream and advected by the current. This dipolar structure then detaches from the cape and, as is observed in nature, the cyclone tends to weaken in strength while separating from the cyclone, whereas the anticyclone moves away from the cape and has a much longer lifetime. The laboratory measurements also show that there is a quick upward diffusion of the negative relative vorticity.

1. Introduction

Mediterranean Water (MW) penetrates into the North Atlantic as an intermediate layer between about 500 and 1300 m with salinity and temperature values higher than the surrounding Atlantic water. One of the interesting features of this water mass in the Atlantic is the vertical splitting into two main cores (Zenk 1970; Ambar and Howe 1979a, b) centered at depths of about 800 m (upper core) and 1200 m (lower core). The submesoscale eddies detached from the Mediterranean undercurrent—the meddies—transport salty and warm water of Mediterranean origin and act like point sources of this water mass in the northeast Atlantic basin (Armi and Stommel 1983). Since the first time that a meddy was recognized as such (McDowell and Rossby 1978), many of these lenslike structures were identified either by their anomalous thermohaline characteristics (e.g., Armi and Zenk 1984; Käse et al. 1989; Prater and Sanford 1994) or by the anticyclonic rotation present in subsurface float trajectories (e.g., Armi et al. 1989; Richardson et al. 1989; Bower et al. 1995). In general, the reported meddies have shown the two maxima (upper and lower cores) in the thermohaline structure but, in some cases, only
one maximum was present (e.g., Pingree and Le Cann 1993b). A census of the published observational evidence of meddies, mostly using CTD data, was presented by Richardson et al. (1991). Information about the trajectories and the possible sites of meddy formation was obtained using freely drifting subsurface RAFOS floats (Bower et al. 1995, 1997). More recently, Richardson et al. (2000) have summarized the results of several sets of recent subsurface float data showing the main pathways taken by the meddies and documenting the frequent meddy disruption due to collisions with seamounts and the resultant dispersion of large amounts of warm salty MW.

The description of the dynamical characteristics of the meddies has been mainly based on field observations, although some numerical models (e.g., Beckmann and Käse 1989) and laboratory experiments (Hedstrom and Armi 1988; Bormans 1992) were able to simulate partly the observed behavior and to provide some details about their dynamics. In a “typical” meddy the velocity field is composed of a core in near solid-body rotation with periods of about 3–7 days, the periphery of which corresponds to the velocity maximum. In the outer region, the velocity decays exponentially with radius (Richardson et al. 1989; Schultz Tokos and Rossby 1991). The estimates of meddy (vertical and horizontal) dimensions yield quite different values depending on their being based on dynamical considerations or on the characteristic thermohaline anomaly field. In fact, the clockwise rotation associated with the meddies seems to extend vertically much farther than the depth range where the thermohaline anomaly is detected (~500–1300 m) and may reach the sea surface (Pingree and Le Cann 1993a; Schultz Tokos et al. 1994; Paillet et al. 1999; Oliveira et al. 2000) or extend down to levels around 2000 m (Pingree 1995).

The interior region of a meddy is weakly stratified and bounded by a sharp decrease in temperature and salinity (Armi et al. 1989; Hebert et al. 1990). The meddy center corresponds to a region of minimum stability (Brun–Väisälä frequencies of the order of 1 cph) bounded by high stability (~2 cph) regions above and below (Pingree and Le Cann 1993b; Prater and Sanford 1994). The front corresponding to the maximum of variance of the thermohaline properties is usually found at distances from the meddy center greater than the maximum velocity radius (Schultz Tokos and Rossby 1991; Hebert et al. 1990) and apparently coinciding with the potential vorticity front where the transition from the inner strong anticyclonic relative vorticity to the outer weak cyclonic vorticity occurs (Armi et al. 1989; Schultz Tokos and Rossby 1991). The relative vorticity of the core of the meddies has been estimated as a fraction of the planetary vorticity (f), ranging from −0.9f to −0.3f (Prater and Sanford 1994; Bower et al. 1997). The peripheral perturbations surrounding the meddy core are in the region of maximum positive (cycloidal) vorticity (Pingree and Le Cann 1993a).

The meddy generation mechanisms that have been proposed so far in the literature include mixing followed by geostrophic adjustment of the mixed water column (McWilliams 1985, 1988), baroclinic instabilities generated in a meridional current along an eastern boundary layer (Käse et al. 1989) or along a vertical wall on an f plane (Baey et al. 1995; Baey 1997; Sadoux et al. 2000), intermittences in the Mediterranean Undercurrent (Nof 1991), and generation of anticyclonic vorticity by separation of the frictional boundary layer of a coastally trapped jet downstream of a sharp corner (D’Asaro 1988).

Several experimental studies were devoted to surface gravity flows along a plane vertical wall, either in rotating channels (Vinger and Mcclimans 1980; Mcclimans and Green 1982) or in rotating tanks (Griffiths and Linden 1982; Obaton 1994), and showed that a baroclinic instability could lead to the formation of anticyclonic vortices. In a comprehensive parametric study, Baey et al. (1995) showed that an intermediate current flowing along a plane vertical wall has a very deterministic behavior, which can be predicted by knowing the initial parameters (stratification and rotation) and the boundary conditions (volume flow rate of injected intermediate water). For the lowest values of both the Ekman number and the Rossby number characterizing the intermediate current, a mixed baroclinic and barotropic instability was observed, which led to the formation of long-lived anticyclonic lenses of intermediate water, hence showing that bottom topography is not crucial for the generation of meddylike structures. Jungclaus (1999) simulated numerically an outflow of dense water in a linear sloping bottom and showed that topographic variations are not needed to destabilize the flow. The dipole formed is shed from the current with the anticyclonic part growing and then separating from the weaker cyclone. When considering the interaction of a surface or a subsurface current and a cape, D’Asaro (1988) suggested that the lateral friction at the coast was responsible for an anticyclonic coastal layer that, together with the overshooting of the current at the cape, was responsible for eddy generation. However, he did not mention any condition for the upstream current nor for the cape angle. Bormans and Garrett (1989) also evoked that for a corner with a small enough radius of curvature (Roed 1980), the current overshot at the corner and reattached to the wall, generating an anticyclone in between. In Stern and Whitehead (1990) experiments, dipoles were detached from a cape for a sufficiently large corner angle, but the current did not reattach to the downstream wall. Klinger (1994) performed f plane experiments for a surface current and also concluded that the cape angle is an important parameter: the vortex formation at the cape occurs only for cape angles less than 135° and vortex detachment is observed for cape angles less than 90°. It seems that, when there is lens production, it always occurs at the same location, and the detachment involves an active lower layer associated...
with a weaker cyclone. The cyclone and the anticyclone formed a heton, which tended to drift due to mutual advection of the vortices. On a $\beta$ plane and also for a surface current, Cenedese and Whitehead (2000) observed experimentally that no detachment at the cape (simulated by a radial thin vertical wall) occurs unless the bottom slope, and hence the topographic $\beta$ effect, is sufficiently large. On the other hand, Pichevin and Nof (1996) concluded that the $\beta$ effect and/or advection are necessary for vortex detachment to occur; on an $f$ plane, an ever-growing eddy is generated at the cape but never detaches. Studying the interaction of an intermediate current and a cape, Sadoux et al. (2000) showed that stable ($\text{Ro} > 0.4$) upstream conditions never led to lens generation at the cape. However, when the upstream current was itself unstable ($\text{Ro} \approx 0.1$), then the cape was a privileged place for lens generation, although not the only one. When the upstream current was stable but with a moderate to low Rossby number ($0.6 \leq \text{Ro} \leq 0.2$), there was dipole formation.

The purpose of this paper is to describe some interesting new aspects of meddy generation and shedding mechanisms that are observed in field observations and simulated in laboratory experiments. Section 2 describes the data collection both in nature and in the rotating tank, and section 3 refers to the main aspects that are common to these two datasets of different origin. The analysis, comparison, and discussion of the results are presented in section 4, and finally section 5 summarizes them.

2. Data and methods

a. Field measurements

The observations discussed here were obtained during the field program AMUSE (A Mediterranean Undercurrent Seeding Experiment) and in the frame of the CANIGO (Canary Islands Azores Gibraltar Observations) and the AMPOR (Physical, Chemical, Geological, and Biological Aspects Associated to the Presence of the MW off the Portuguese Coast) projects. The main objectives of AMUSE, which was carried out in 1993–94, were to study the spreading of the MW off the Iberian Peninsula and the formation of meddies (Bower et al. 1995, 1997). A total of 49 subsurface RAFOS floats were deployed over a period of about 8 months off the southern coast of Portugal at about 1200 m, the level where the lower core of MW is centered, and were acoustically tracked for an 11-month period. The hydrological setting was provided by a high-resolution (1.5 nautical miles spacing) XBT section K (Fig. 1) across
the Mediterranean undercurrent, which was repeated each time RAFOS deployments took place. The selection of the site for the float deployment was made in accordance with the information given by the XBT line and trying always to seed the main body of the undercurrent.

The CANIGO and the AMPOR field experiments, whose data will be used here, had some similarities with AMUSE: deployment of RAFOS floats for a period of one year (Sep 1997–Sep 1998) and repetition of a high-resolution (2.5 nautical miles spacing) XBT line S (Fig. 1) each time the deployments were done. The floats were ballasted for the levels of 800 and 1200 dbar, where the upper and lower cores of MW are centered, respectively (Ambar and Howe 1979a,b). In several instances the thermal structure indicated that, besides the main body of the undercurrent against the slope, there was also an offshore blob that could eventually be an eddy or a meander of the undercurrent. In these cases, an additional pair of floats was deployed within the offshore structure.

**b. Experimental setup**

For the laboratory experiments the LEGI-Coriolis 14-m diameter rotating platform was used. The 13-m diameter, 1.2-m deep cylindrical tank was filled with a linearly stratified fluid. The rotation period $T$ and the density gradient, and thus the Coriolis parameter $f = 4\pi/T$ and the Brunt–Väisälä frequency $N = (-\rho^{-1}g\partial\rho/\partial z)^{1/2}$ respectively, could be adjusted to cover a wide range of variation of the relevant parameters. Before introducing the intermediate current, the fluid in the tank was allowed to reach a solid body rotation regime, at least at first order. In fact, there was a very weak $O(1 \text{ mm s}^{-1})$ anticyclonic circulation in the upper layer that was practically impossible to avoid since it was due to the relative velocity of the air with respect to the rotating tank. However, for some experiments, a plastic cover was put on almost the whole surface of the tank, allowing the reduction of the drift current to $O(0.1 \text{ mm s}^{-1})$. Figure 2a is a front view of the experimental setup showing the injector of intermediate water, the vertical wall along which flowed the intermediate current, and the cape. Figure 2b shows top and front sketches of the installation. A collector was placed next to the injector at the same level, thus ensuring a constant water volume in the tank and a water injection at exactly its equilibrium level in the stratified fluid. The volume flow rate $Q$ of the intermediate current could be adjusted and was kept constant with time for the whole duration of each experiment. The velocity and interface level measurements done close to the injector showed the effectiveness of that device for obtaining an intermediate current in geostrophic balance after leaving the injector (Sadoux et al. 2000). In order to study the interaction between an intermediate current and a cape, a 4-m long curved vertical wall was installed (the “upstream” wall) par-
allel to the tank wall and 2 m away from it. Another plane vertical wall (the “downstream” wall) was set at an angle α with the upstream wall. The injector was placed so that the undercurrent moved along that upstream wall. The cape angle α was taken as 70°, which corresponds approximately to the shape of the 1000-m isobath off Cape St. Vincent. Sadoux et al. (2000) showed that for Cape angles smaller than 120° the cape angle has no direct influence upon the observed phenomena but changes the characteristic temporal evolution of the produced anticyclonic vortices.

The density profiles along several vertical lines in a plane perpendicular to the wall were measured with ultrasonic probes by recording the variations of the propagation time of an acoustic signal between two small sensors placed horizontally 2 cm apart and moving vertically at a constant speed. This allowed us to define the upper and lower layer interfaces of the current along the upstream wall or of a detached lens away from the cape. These measurements have shown that the flow was in geostrophic balance for the “stable” upstream condition experiments and made possible to measure the width and thickness of the current immediately upstream of the cape. Use was made of a noninvasive Correlation Imaging Velocimetry (CIV) system developed by Fincham and Spedding (1997) and recently extended to quasi-3D measurements (Fincham 1997; Didelle and Fincham 1998). Polystyrene beads of diameter ~0.7 mm were seeded during spinup throughout the tank over the whole depth. The bead density ranged from the density of a layer located ~5 cm below the free surface to the density of a layer located ~10 cm above the bottom so that there were neutrally buoyant beads in the whole body of the fluid between these two levels. Additional beads with the intermediate water current density were also directly introduced into the current at the injector. A laser beam was swept vertically by an oscillating mirror and reflected by a submerged 45° mirror fixed in the tank to illuminate the advection of particles within a horizontal sheet ~0.5 cm thick. The laser sheet was moved vertically to scan a fixed volume of water (2.75 × 2.02 × 0.43 m³) and a digital camera was placed directly over the flow to capture images. At each level, on a grid with Δx = 3.6 cm and Δy = 4.2 cm resolution, the horizontal components of the velocity field could be determined with an accuracy O(1%–2%) by cross-correlations of the intensity of the light reflected by the particles between subregions of sequential image pairs centered on each grid point. There were measurements at 25 levels in each volume scan, allowing us to compute derived quantities, such as the vertical component of vorticity and the velocity divergence, at these levels, and to get a quasi-3D image of these variables. The 3D reconstructions of the velocity field enabled us to estimate the current width and thickness at the wall, which proved to be in good agreement with the estimates obtained through the density profile measurements mentioned earlier. Hence the CIV measurement and the vertical density profilers provide two alternative and independent ways to determine the current properties. The characteristic values of the velocity maximum and of the vorticity for the coastal current and for the meddy were estimated by averaging the corresponding measurements at several levels encompassing the current thickness.

Previous studies (Baey et al. 1995; Sadoux et al. 2000) of the stability of an intermediate water current flowing along a plane vertical wall and initially in geostrophic equilibrium showed that the relevant physical parameters of the flow are the Rossby (or alternatively the Burger), the Froude, and the Ekman numbers, defined as Ro = U/L (Bu = UNh/fL²), Fr = U/Nh, and Ek = νfL², where U is the current maximum velocity; L and h its characteristic width and thickness along the wall, ν the kinematic viscosity; N the Brunt–Väisälä frequency, and f the Coriolis parameter. For the experiments discussed in this paper, the Rossby and the Froude numbers varied in the range [0.1–1.2] and [0.1–0.5], respectively, and the Ekman number was kept at an average value of 5 × 10⁻⁴.

3. Description of the observations

a. Observations in the ocean

Figures 3–5 present some results from two independent datasets (AMUSE and CANIGO/AMPOR) of RA-FOS floats and XBT drops. The different types of behavior of the AMUSE floats after deployment are illustrated in Fig. 3. Only the first 30 days of the trajectories were plotted and the dots separate each 3-day segment. The bathymetric contours correspond to the 1000, 2000, and 3000 m.

The first two pictures of the sequence in Fig. 3 show two examples of the presence of a cyclone coupled with a meddy at the formation site. In the first one (Fig. 3a), the float closer to the continental slope started looping anticyclonically in a newly formed meddy (A1) after crossing the 1000-m isobath as this line turns meridionally off Cape St. Vincent (Bower et al. 1997). It performed three complete loops and afterwards it was expelled from the meddy. Meanwhile, the other contemporary float revealed the presence of a cyclone (C1) with its center 50 km to the southwest from the center of the meddy. This float completed one single cyclonic loop in a period of approximately 10 days, while the looping period of the float within the meddy was between 3 and 4 days. The azimuthal velocity records of the floats within the meddy and within the cyclone showed strong amplitude oscillations (from 15 to 45 cm s⁻¹ and from 5 to 20 cm s⁻¹ respectively), the float in each vortex accelerating when closer to the companion structure. Both floats registered temperatures higher than 11.5°C during the time they were in the respective coherent structure. The second example of coupled cyclone (C2) and anticyclone (A2) involved three floats.
(Fig. 3b), two in the cyclone and one in the anticyclone, and the main difference relative to the previous example is the site where the float within the meddy started looping. This time the meddy was already formed when it reached the western edge of the Cape St. Vincent spur. Furthermore, in the previous example the meddy float took a path closer to the 1000-m isobath than in this latter case, diverging from the upper slope westward from 9°W. The looping period of the float within this second meddy was 5–6 days, the azimuthal speed varied between 10 and 25 cm s⁻¹, and the recorded temperature was higher than 11.5°C. The float in the cyclone centered at 10°W performed loops at a speed oscillating between 10 and 20 cm s⁻¹ while registering temperatures between 11.5° and 12.0°C.

Figures 3c and 3d show two floats that moved rapidly to the northwest, and this can be attributed to the influence of existing dipolar structures. This is certainly the case of the float in Fig. 3c, which passed in the composed velocity field between the cyclone C1 and the anticyclone A1 of Fig. 3a.

Another typical observed behavior is illustrated in Figs. 3e and 3f. These show two floats that kept following the continental slope after turning to the right (facing downstream) at Cape St. Vincent without being entrained in or influenced by any vortical structure. The following two examples (Figs. 3g and 3h) show two other floats whose depths were 850 and 1100 m, respectively, 10 and 30 days) while registering high temperatures (respectively above 12.5° and 12.0°C).

Finally, Figs. 3i and 3j exemplify floats that turned south before reaching the cape, this being due to the influence of a cyclone present in the Portimão Canyon region, as will be seen next in this section.

The synoptic view of the float trajectories helps us to understand the relation between the structures present and to get insight on the evolution of the flow field. Figure 4a is a composite of information from the AMUSE and the CANIGO float experiments. The figures in the first and second rows (Figs. 4a to 4f) result from dividing in segments of 15 days each of the AMUSE float trajectories between 15 January 1994 (15/94) and 15 April 1994 (105/94). The third row gathers sequential data from some of the CANIGO floats, divided in periods of 30 days from 28 January 1998 (28/98) to 28 April 1998 (118/98). So each figure in the sequence is a snapshot of float paths in the MW layers. Four repetitions of the AMUSE XBT line K (see Fig. 1 for location) and two of the CANIGO XBT line S (Fig. 1) are presented in Fig. 5 in order to characterize the thermal field at the site and at the time the floats were deployed. In these figures, only the isotherms in the range 10.0° to 14.0°C are contoured (the shade corresponds to temperatures higher than 11.5°C), and the solid dots indicate where the RAFOS floats were deployed.

In the first and second time periods of Fig. 4 (15/94 to 45/94) we can follow the formation of the dipole (C1–A1) that was present in Fig. 3a, as well as identify the trajectory of the float passing between the cyclone and the anticyclone of this dipole (Fig. 3c) and of the float staying in the St. Vincent Canyon (Fig. 3g).

In the repetition of the XBT line K of 08/94 (Fig. 5a) a large blob of warm water (T > 12.0°C) is clearly identified at the offshore edge of the line. During the next four repetitions of the same line (not shown) an offshore structure was still present, although showing less and less temperature contrast until the repetition performed in 44/94 (Fig. 5b) where only the two offshoremost stations have temperatures higher than 11.5°C at mid depth. Looking back to the first three 15-day snapshots (Fig. 4a to 4c), it is interesting to see that several floats revealed the presence of the cyclonic structure C2 in the Portimão Canyon region, which can be identified with the aforementioned offshore blob. In the period 45/94 to 60/94 (Fig. 4c), the cyclonic structure moved westward from the site where it had been intersected by the XBT line, translating along the southern edge of the Mediterranean undercurrent. The presence of this cyclone is clearly identified by the movement of several floats, such as those that were forced by it to move south (Fig. 4c) after being caught in its western side. During this period other floats (those of Figs. 3e and 3f) were moving northward against the continental slope, and this occurred after the shedding of the dipolar structure (C1–A1) off Cape St. Vincent (Fig. 4b).

The temperature field measured along the XBT line K of 57/94 (Fig. 5c) shows a peculiar axisymmetrical structure almost completely detached from the slope, with temperatures reaching 13.6° and 12.4°C at the levels of the upper and lower MW cores. Three floats were then deployed within this structure (see dots in the figure). Seven days later (on 64/94), the warm blob had disappeared from this line (Fig. 5d), and in the meantime the offshoremost float deployed in it reached Cape St. Vincent and started looping anticyclonically (Fig. 3b). This poses the question whether that blob observed in 57/94 was in fact the meddy A2 revealed by that float a little farther downstream.

The sequence of the RAFOS trajectories (Figs. 4a to 4d) shows that the cyclonic structure C2, which moved

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<Fig. 3. Different types of behavior of the AMUSE floats after deployment: (a)–(b) floats showing coupled cyclone and anticyclone, (c)–(d) floats passing between a cyclone and an anticyclone, (e)–(f) floats following northward along the continental slope contours, (g)–(h) floats staying within St. Vincent canyon, and (i)–(j) floats turning south before reaching Cape St. Vincent. The dots separate 3-day periods and the bathymetric contours correspond to the 1000, 2000, and 3000 m depth.>
Fig. 4. Trajectories of RAFOS floats: (a) to (f) from AMUSE experiment, (g) to (i) from CANIGO experiment. The AMUSE float sequence is divided in 15-day periods from 15 Jan 1994 (15/94) to 15 Apr 1994 (105/94) and the CANIGO float sequence in 30-day periods from 28 Jan 1998 (28/98) to 28 Apr 1998 (118/98). The dots correspond to the last float position in each time period. Also shown are the depth contours 1000, 2000, and 3000 m and the locations of the (K or S) XBT lines.
away from the Portimão Canyon, interacted with the above referred meddy A2. In the period 60/94 to 75/94 (Fig. 4d) the cyclone C2 was very close to the meddy A2 (their centers being 35 km apart). Although the float that was within this cyclone has left it (around day 75/94), the presence of this structure was revealed by another float in the period 75/94 to 90/94 (red trajectory in Fig. 4e) but now situated to the southwest of the meddy. The centers of rotation of each of these vortices were 50 km apart during that period. In the last period shown for the AMUSE floats (Fig. 4f), the cyclone float stopped looping, thus not revealing further evolution of the cyclone. It is interesting to notice that in this period meddy A2 moved faster than in the previous period and took a more westward oriented path. Furthermore, as soon as it moved away from the slope, the float that remained for more than 30 days in the St. Vincent canyon was able to move north (Fig. 4f). The CANIGO dataset presented a similar situation to the one just described. In the thermal field along the XBT line S of 28/98 (Fig. 5e) a large blob of relatively warm water ($T > 12.5^\circ C$ at $\sim 750$ m) was found offshore centered at about $36^\circ N$. In the next XBT line (52/98) the structure is still identifiable (Fig. 5f) although with slightly lower temperatures. Back to Fig. 4 (last row), one can see the trajectories of the floats deployed in these two repetitions of XBT line S (see also dots in Figs. 5e and 5f). Once again the offshore blob can be associated to a cyclonic structure (C3). This time it was possible to evaluate that its vertical extent was at least 800 m since the upper float, which was deployed on day 52/98 and ended up at around 400 m, was caught in the same structure as the 1200-m float deployed in the same day. Geostrophic computations (not shown), based on CTD
data obtained along a CANIGO section coinciding with the XBT line S, pointed to twice as large a vertical extent of the vortex structure, its influence spanning from the surface to about 1600 m. In the time period from 58/98 to 88/98 (Fig. 4h) the cyclone C3 moved to the west and paired with an anticyclone A3 present in the Cape St. Vincent region. The lack of some intermediate data prevented the tracking of the whole path of the cyclone, but the following period of 30 days (Fig. 4i) shows the cyclone still present south of the meddy.

The similarity between some of the flow patterns observed in the AMUSE and in the CANIGO programmes is striking. Examples are the evidence of the Mediterranean undercurrent close to the slope, the presence of cyclonic structures in the Portimão Canyon region, the westward translation of these vortices at the edge of the undercurrent, the coupling of cyclones with meddies near Cape St. Vincent, and the west-northwest movement of the dipolar structures as they get away from the generation region.

b. Observations in the rotating tank

Figure 6a illustrates what happens in the case of an upstream stable current (Ro ~ 1) for a 90° cape angle. One can see the stable current flowing along the upstream wall, circling around the edge of an anticyclone, and then propagating along the downstream wall passed the vortex. Sadoux et al. (2000) checked that both the upstream and the downstream currents were in geostrophic equilibrium, and the density profiles and velocity measurements performed during this set of experiments confirmed that conclusion. The vortex is created downstream of the cape due to overshooting of the current at the end of the upstream wall. The subsequent reattachment to the downstream wall is due to the Coriolis force. The anticyclone diameter increased with time, but this vortex never detached. In some instances it was slowly advected by the current along the downstream wall and replaced by another one at the cape. Sadoux et al. (2000) verified that, whenever there was an anticyclone formed and remaining at the cape, its diameter grew linearly with time. The respective growth rate was independent of the rotation velocity for all upstream conditions and increased with the cape angle. This growth rate was also independent from the volume flow rate of the intermediate current. Figure 6b shows an example of a CIV-computed velocity field at the mean level of the current over a 2.75 × 2.02 m² area in the case of a 70° cape, while Fig. 6c represents the vertical component of the vorticity corresponding to that velocity field. One can see the presence of the anticyclone located immediately downstream of the cape. It is also worth noticing that there is an O(5 cm) band of cyclonic vorticity almost all along the offshore edge of the current and of the cape anticyclone. Moreover, in all experiments with a stable upstream current, there were return currents above and below the intermediate flow, which were generated by the intrusion of this current in the (almost) quiescent fluid. The return current present below the intermediate current was always weak and vanished with time whereas the return current located above remained all the time. This upper countercurrent was
FIG. 6. (Continued)
responsible for the generation of a cyclone immediately upstream of the cape. With time, this cyclone tended both to extend downward and to be shifted slightly to the edge of the current. Soon this cyclone and the anticyclone at the cape interacted. For these high Rossby numbers it always appeared that the cyclone was moved anticyclonically around the standing and growing cape anticyclone. The CIV measurements showed that during that process the cyclone was extending both upward and downward in a barotropization process. Thus, the cyclone appeared to be looping around the anticyclone and then advected along the downstream current. This process repeated itself with a period $O(3-4 \times T_c)$ so that, after a while, several cyclones could be recorded along the outside edge of the anticyclone and the downstream current.

For smaller Rossby numbers ($0.2 \lesssim \text{Ro} \lesssim 0.6$), the same process occurred but with a weaker anticyclone generated downstream of the cape. As Ro decreased, the growth rate of both the cyclone and anticyclone became comparable. Hence, dipole generation at the cape could happen with the dipole following a line that was roughly an extension of the upstream wall. For the largest values of Ro in the range $0.2 \lesssim \text{Ro} \lesssim 0.6$, both the cyclone and anticyclone had large diameters $O(1 \text{ m})$, but the anticyclone was always larger than its cyclonic counterpart. Figure 7a is a picture of the flow, illuminated by the laser sheet in a plane located at the level of the intermediate current, which shows the dipole shortly after it moved away from the cape. The diameter of each vortex decreased with decreasing Ro and the structure became more symmetric. As Ro decreased, the cyclonic motion penetrated progressively deeper, even reaching levels below the intermediate current. Figures 7b and 7c show the fields of the velocity and of the vorticity vertical component for this example.

Figure 8 is a sequence of images representing the vertical vorticity averaged over nine planes and, for the corresponding time, a 3D representation of vertical vorticity isolines showing the formation and detachment of three lenses when the upstream current is unstable (Ro = 0.2). It appears that there is always a cyclone associated with the anticyclone, for there is a continuous transition from the growing anticyclone immediately downstream the cape (Ro > 0.6) to the present case. However, most of this cyclone is located in the layers above the intermediate current so that its signature at the level of the intermediate current is weak. On the other hand, the anticyclone extends from the level of the intermediate current until the uppermost level of the illuminated volume so that this anticyclonic signature has a much larger vertical extension than the density anomaly, which is confined to the level of the intermediate current. The coupling between the cyclone and the anticyclone is most likely responsible for the self-propagation of this dipolar structure away from the cape. As this movement proceeded, the cyclonic part weakened and tended to separate from the anticyclonic part so that at some distance from the cape only the anticyclone is observed. In most cases (a few without any significant residual current and some with a residual current, either in the same direction or in the opposite direction relative to the intermediate current), these heltonlike structures followed a path that was roughly in the extension of the upstream wall, as previously noted for the dipoles. This is particularly noticeable since, even close to the cape, the lenses that were produced along the upstream or downstream walls always moved away from the wall along a line practically perpendicular to the wall. As is clear in Fig. 8, as soon as a lens detaches from the cape, another one immediately forms and the process repeats itself periodically. It could be noticed later in the experiment that a cyclone advected from upstream by the current got close to the cape and enhanced the detachment of small anticyclonic structures.

When the upstream flow is so unstable (Ro < 0.05) that only a small part of it reaches the cape, this cor-
Fig. 8. Time sequence of averaged vertical vorticity (on the left) and the corresponding 3D isosurfaces (on the right) from an experiment with unstable upstream conditions ($Ro = 0.15; Ek = 2 \times 10^{-4}; Fr = 0.12$): (a) $t = 3.5T_i$ ($T_i = 25$ s), (b) $t = 5.2T_i$, and (c) $t = 7.3T_i$ ($T_i$: inertial period).
responding to a mean current velocity at the cape lower than 0.2 cm s\(^{-1}\), that is, \(O(0.1Nh)\), then the current remains always attached to the wall without any overshooting or lens generation.

4. Comparisons and discussion

a. Phenomenological comparison

The Rossby number was proven to be an important parameter for determining the flow regime (Baey et al. 1995; Sadoux et al. 2000). For \(Ro \sim 1\) the flow in the rotating tank is stable and there is only the production of an anticyclone at the cape due to the overshooting of the current, which then reattaches to the wall downstream of this vortex. To our knowledge, there is no observational evidence of such a standing anticyclonic vortex at the cape; hence we can probably assume that such stable conditions are seldom, if ever, occurring in the Mediterranean Undercurrent upstream of Cape St. Vincent. At the other end of the Ro range of variation (\(Ro < 0.1\)), the upstream flow is so unstable that very little flow actually reaches the cape, and what remains of the current moves around the cape without producing any perturbation. This could correspond to the in situ observations when floats followed along the continental slope, either directly moving in the northward direction (Figs. 3e and 3f) or stagnating in the St. Vincent canyon for some time before being released and then moving northward (Figs. 3g and 3h).

When the Rossby number decreases from unity to a value \(O(0.1)\), the flow field regime in the rotating tank changes from totally stable to unstable. The gradual decrease in the Rossby number is accompanied by the setup and growth of a cyclonic circulation at the edge of the anticyclone so that, for moderately unstable conditions (\(Ro \sim 0.4\)), both vortices are of the same intensity and dipole production can occur. In very unstable conditions (\(Ro \sim 0.1\)), anticyclonic monopolar lenses are generated, but there is still the presence of companion cyclones.

As described above, one of the new features exhibited by the in situ datasets is the presence of cyclones, which are likely to be involved in the detachment of meddies. The AMUSE float data showed some (apparently) isolated meddies, but this could be due to the low density of floats in the water, which did not allow the observation of the companion cyclones. On the contrary, when the number of floats was large enough to resolve the major aspects of the flow field, meddies were always observed with companion cyclones.

b. Flow regime diagram

The estimation of the Rossby (\(Ro = U/\text{fL}\)) and Froude (\(Fr = U/Nh\)) numbers for the case of in situ observations is not as straightforward as for the case of the model observations, where the CIV measurements enable the characterization of the three-dimensional structure of the velocity. Before comparing the nondimensional numbers obtained in both cases, an explanation is needed of how the velocity, the width, and the thickness of the current were evaluated for the ocean dataset. The region considered for these calculations was limited between the site of the float deployment (\(\sim 8^\circ30’ W\)) and Cape St. Vincent (\(\sim 9^\circ W\)). For each of the repetitions of the XBT line, a graph of the cross-stream distribution of the undercurrent velocity was constructed by using the velocities of the floats that followed to the west in that region. The fitted curves (Fig. 9) were constrained to the vanishing of the velocity at the side wall (in this case, the continental slope at the depth of about 1100 m), to a quasi-linear increase between the wall and the velocity maximum and to an exponential decrease offshore from this maximum. Since the AMUSE floats were deployed on a nearly regular time basis, the fitted lateral distributions of velocity can give an estimate of the time evolution of the maximum velocity (\(U\)) and of the width (\(L\)) of the undercurrent, taken as the offshore distance from the slope where the velocity decreases to a value of 0.5 cm s\(^{-1}\). For the estimation of the Froude number, the current thickness (\(h\)) was taken as that of the MW layer, and this was estimated to be approximately 700 m based on the hydrology sections during AMUSE and CANIGO. The Brunt–Väisälä frequency (\(N\)) was computed for every CTD cast in the region under consideration and averaged to give a single \(N\) profile (Fig. 10), from which the value 0.0027 s\(^{-1}\) was taken as representative of the undercurrent layer. For the laboratory experiments, the Rossby and Froude numbers were computed for the intermediate current upstream of the site of generation of anticyclonic lenses. Table 1 summarizes the main characteristic values and nondimensional parameters for the current in the case of in situ and model data.
The Ro–Fr diagram is presented in Fig. 11 for both in situ data (open symbols) and model data (solid symbols) for the same range of Ro. Regarding the in situ data, the circles correspond to floats that ended up in newly formed meddies at Cape St. Vincent. The squares correspond to cases where the floats deployed stayed in St. Vincent Canyon or continued to follow the continental slope downstream of the cape. All other cases are plotted with diamond-shaped symbols. Whenever a meddy was generated, the Froude and Rossby numbers were within the range $[0.13-0.23]$ and $[0.09-0.13]$, respectively. It is also worthwhile noting that for lower values of the Froude and Rossby numbers (respectively below 0.13 and 0.09) there was no meddy production. The values below those limits correspond to cases of slope-trapped flow or floats staying for long periods in the canyon.

In all the tank experiments with several anticyclonic lenses produced at the cape, the Froude and Rossby numbers were, respectively, in the range $[0.05-0.14]$ and $[0.05-0.15]$ corresponding to values for which the current along the upstream vertical wall was unstable. The periodic generation of anticyclonic lenses, either before or at the cape, could then be associated with a mixed baroclinic–barotropic instability (Baey et al. 1995).

There is a general agreement between the results from the in situ measurements and from the model data, namely that there is an intermediate range of Fr and Ro for which meddies are generated and that for the lowest range of Fr and Ro there is no eddy production at the cape. However, the absolute values of the nondimensional parameters defining these ranges differ from in situ to model data, especially in what concerns the Froude number. The chosen mean values for the stratification and for the thickness of the Mediterranean Undercurrent can account for the differences found in the Froude number. The Rossby number values estimated for the Mediterranean undercurrent using the float data never exceeded 0.20, and this is compatible with a maximum value of 0.25 (using a mean value of 15 km for the current width, which can be considered a lower limit) based on the direct current measurements of the undercurrent (at 800 and 1200 m) off south Portugal in a 14-month period, obtained in the frame of the CANIGO project.

c. Characteristic quantities

Eight trajectories of RAFOS floats (six from AMUSE and two from CANIGO), corresponding to eight meddy formations at Cape St. Vincent, were analyzed in detail and were used to estimate the radius, the rotation period, and the azimuthal and translational velocities. Average values of these quantities have been already reported in Bower et al. (1997) or in Richardson et al. (2000). In order to characterize the lenses produced in the tank and compare them with the meddies observed in the ocean, the CIV measurements were vertically averaged over the nine horizontal planes that corresponded to the current thickness at the wall. The resulting fields of the horizontal velocity and of the vertical component of vorticity were taken as representative of the current and of the generated lenses.

Focusing first on the anticyclonic lenses, a very striking result is the high coherence of the experimental results: the radius, the rotation period, and the value of the vertical component of vorticity, both as the lens detaches from the cape and as it evolves away from it, are almost the same for all the lenses generated in several distinct experiments. Table 1 summarizes the main characteristic parameters both for meddies followed by RAFOS floats (during AMUSE and CANIGO) and anticyclonic lenses produced in the tank during different experiments. Some of these quantities are normalized in order to make the comparison easier. The meddy radius is normalized by the Rossby radius of deformation $R_d = Nh/f$ associated with the undercurrent; the rotation period $T_r$ and the formation period $T_f$ are normalized by the inertial period ($T_i = 2\pi f$, where $T_i = 20$ h in this oceanic region and $T_i = 25$ s in the tank); the translation velocity $V_t$ is normalized by the maximum velocity $U_{max}$ and the relative vorticity $\omega$ is normalized by the planetary vorticity $f$. As can be appreciated, there is good agreement between the nondimensional quantities computed for the ocean and for the model, as they are within the same range of variation.

The horizontal velocity fields measured in the laboratory confirmed that the core of the anticyclonic structures was in solid-body rotation. The same is practically true from the reconstructed in situ velocity fields since the azimuthal velocity increased linearly from the center of the structure until a certain radial distance and decreased outward. Thus, assuming solid body rotation for the cores of the in situ vortices (e.g., Schultz Tokos and Rossby 1991), the vertical component of vorticity was...
Table 1. Summary of all characteristic quantities estimated with the in situ and the model data for the cases of meddy and lens generation at the cape (see definition of the parameters in the text).

<table>
<thead>
<tr>
<th>Case</th>
<th>Current</th>
<th>Meddy</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$f$ (s$^{-1}$)</td>
<td>$N$ (s$^{-1}$)</td>
</tr>
<tr>
<td></td>
<td>Ocean</td>
<td></td>
</tr>
<tr>
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<td>43 000</td>
</tr>
<tr>
<td>Am116a</td>
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<td>35 000</td>
</tr>
<tr>
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<td>35 000</td>
</tr>
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</tr>
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</tr>
<tr>
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</tr>
<tr>
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</tr>
<tr>
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</tr>
<tr>
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</tr>
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</tr>
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</tr>
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</tr>
<tr>
<td>Fi01</td>
<td>0.25</td>
<td>0.31</td>
</tr>
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</table>

* Not enough floats in water.
FIG. 11. Flow regime diagram in the Ro–Fr plane for the experiments and for the in situ observations. The Rossby and Froude number are computed from the velocity, stratification, width, and thickness of the current upstream the cape. Open symbols: in situ data (○ floats that ended up in newly formed meddies; □ floats following the continental slope; ◻ other cases); solid symbols: laboratory experiments (● lens formation; ■ weak current following the wall; ◆ dipole formation).

computed as twice the angular frequency. Since the rotation period of the particles within the core in solid body rotation is the same whatever the distance to the center, the relative vorticity in the core should be constant. Thus, variations in the period of rotation of the vortex imply variations in the relative vorticity of the structures.

In what follows, the temporal evolution of meddy radius, rotation period, translation velocity, and relative vorticity are compared with the respective values computed for the lenses in the tank. Also, the relative vorticity of the companion cyclones are analyzed. Time is normalized by $1/\omega$; for the laboratory experiments, the time origin is taken as the detachment instant, whereas for the in situ data, it is taken as the time of the meddy first observation. In the case of the cyclones, the zero time in the vorticity graphs correspond to the estimated time when the interaction with the meddy at the cape was stronger, that is, when the two structures were closest to each other.

1) RADIUS

In order to compute the meddy radii, the float trajectories were low-pass filtered with cutoff periods of twice the respective meddy looping periods. This allowed us to obtain the trajectories of the meddy centers and to estimate the radii by subtracting these translation components from the float real trajectories. Figure 12a presents the time evolution of a composite of the estimated radii for five of the eight meddy observations. The solid line is a linear fit to the data. As could be expected, there is a sizeable data scattering, but it seems that the average size of the meddy is of $O(0.5R_d)$. The fit points to a slow increase of the radii, but this could mean that the floats tended to increase their distance to the center of the respective vortex due to radial motions associated, for instance, to elliptic instabilities (Hedstrom and Armi 1988) or that there is some diffusion of the vorticity anomaly caused by viscous effects. However, the meddies found in the Canary Basin have larger radii than the ones found near Cape St. Vincent, so an increase over time should be expected. Figure 12b presents the results for the tank experiments in which the lens radius was defined by the zero line of the vorticity vertical

![Figure 12](image-url)

Fig. 12. Evolution of the anticyclonic lens radii normalized by the Rossby radius of deformation associated with the current for (a) in situ data (○ Am103a; (◆) Am110; □ Am120b; ▼ Am129; ◆ Ca452) and (b) model data (○ Pa01; □ Pa02; ◆ Pa03; ▼ Pa04; (◆) Pa05; ▼ Ob01; △ Ob02; ◆ Ob03). The solid lines correspond to linear fits.
component. As can be seen in this figure, the radii of the anticyclonic lenses are $O(0.7R_d)$ when they detach from the current and increase steadily with time. The increase in the laboratory model is more pronounced than in the ocean. This can be related to frictional effects that are higher in the laboratory experiments than in nature. Because of friction there is entrainment of water so that the anticyclonic vortex diameter, measured either by the line of zero velocity or by the line of zero vorticity, is increasing. However, the lens diameter based on dynamic quantities increases, whereas the lens diameter based on the density anomaly varies little during the same time, thus showing that the former increase is due to lateral entrainment of water and not to diffusion. Such a lateral entrainment will also be responsible for slowing the rotation velocity and so an increase of the rotation period of the lens. In the ocean such radius increases occur more slowly.

2) ROTATION PERIOD

The rotation period of each meddy followed by floats was computed based on the estimated azimuthal velocity and the corresponding radius. The linear fit to the data is presented in Fig. 13a and points to a rotation period initially $O(6T_i)$ that remained almost unchanged over time. However, in the cases of two meddies that were tracked for more than six months, the rotation period increased significantly over that period. For the laboratory experiments, the rotation period at the time of each measurement was computed using the maximum velocity and the lens radius. The rotation period of the lenses in the tank shows the following pattern (Fig. 13b): it is practically constant within the range $5-10$ $T_i$ before the detachment time and, afterwards, it increases with time. As said above, the relatively faster increase compared with the in situ data was most likely due to frictional effects.

3) TRANSLATION VELOCITY

The translation velocity of each of the observed meddies was estimated by means of differencing the positions of the low-pass filtered trajectories (which have the looping part removed). Average values of translation velocity were reported by Richardson et al. (2000) using a composite of three different sets of RAFOS floats. The eight time series of translation velocity computed here (not shown) showed a similar behavior: a decrease from values of about 10 cm s$^{-1}$ that is, $O(0.06N_h)$, to values of 2 cm s$^{-1}$, that is, $O(0.02N_h)$, in the first inertial periods. The direction taken by the meddies was toward the northwest.

In the tank, the mean drift velocity of a lens as it moved away from the cape was in the range [1-2] mm s$^{-1}$, that is, $O(0.03N_h)$. In fact, that velocity varied with the residual current in the tank, but it is noticeable that whatever the strength or the direction of that current, all the lenses moved away from the cape initially along the line defined by the upstream wall. The advection by the Mediterranean undercurrent downstream of Cape St. Vincent can be responsible for the initial high translation velocity of the meddies (Richardson et al. 2000). However, in the tank, the intermediate current does not advect the lenses downstream of the cape after their formation. This would explain that the nondimensional drift of the lenses in the tank is similar to that of the meddies when they are not being advected by the undercurrent. Besides the effect of advection, the interaction between a meddy and its companion cyclone is likely responsible for the high translation speeds just after detachment.
4) Relative Vorticity

The anticyclonic relative vorticity of the meddies produced at Cape St. Vincent was estimated to be in the range between $-0.3 f$ and $-0.6 f$, as can be seen in Table 1 and reported in Bower et al. (1997). Meddy vorticities were frequently higher than the current vorticities. Figure 14a shows the evolution of the relative vorticity for the meddies that were observed with companion cyclones. In the tank, the lenses maximum anticyclonic vorticity, when they separate from the cape, is $O(-0.4 f)$, and that is always equal or larger than the upstream current vorticity (the ratio of the lens vorticity to the current vorticity varies from 1 to 2.5). In most instances, the lens vorticity had its maximum value at the detachment time. After it detaches from the cape (Fig. 14b), the vorticity decreases almost linearly due to lateral mixing and to viscous friction mainly at the top and lower interfaces of the lens. It is quite remarkable that all data collected from successive anticyclonic lenses in the two experiments fall on the same decay line.

The vorticity values computed here are close to most of the previous estimates by other authors. Pingree (1995) estimated a value of $-0.45 f$ for a meddy found in the Tagus Abyssal Plain and Schultz Tokos et al. (1994) a value of $-0.40 f$ for Meddy Aska. Prater and Sanford estimated a value a little higher ($-0.85 f$) for Meddy Cadiz.

The relative vorticities of the cyclones that were associated to the shedding of the three meddies discussed in section 3 fall between $0.10 f$ and $0.25 f$. Figure 15a, although with only a few data available, shows that the
cycloic vorticity presents a decreasing trend with time, the values being higher before the cyclones interact with the meddies at the cape \((f = 0)\). In the model (Fig. 15b), the cyclones have initially a slightly weaker \([0.15f \pm 0.35f]\) maximum vorticity when compared to the anticyclones. The evolution of the cycloic vorticity is similar to the evolution of the anticyloic vorticity in the lenses, although with much more scatter (Fig. 15b). This scatter can be partially due to the weak absolute values associated to the cycloic lenses. Companion cyclones observed in the model had relative vorticity higher than those observed in the ocean, but the negative trend was present in both datasets.

With respect to the three dipolar structures found, the ratios between the relative vorticity of the cyclone and of the associated meddy at the time of shedding were 0.34, 0.44, and 1.00. In the tank, when the cyclonic lens detaches, the ratio of cycloic vorticity to anticyclonic vorticity is \(O(0.7)\), and it remains practically constant during the whole lifetime of the cyclone, which is of the order of 15 to 30 \(T_f\). The anticyclone seems to have a much longer lifetime, as proved by several lenses that could be followed during 607, or even longer periods as they evolved out of the measurement area. Also in the ocean, meddies have been tracked for several hundreds of inertial periods, but there is not an example of a cyclone being persistently followed for more than 200 inertial periods (see Richardson et al. 2000).

5) Generation Period

The estimation of the generation period based on float data is prone to errors due to the fact that, to resolve the major flow features, a larger number of floats as well as a better time coverage would be needed. The floats deployed in both observational programs behaved similarly, traveling westward with the undercurrent until they reached Cape St. Vincent. From Table 1 (last column) one can see that the elapsed time between the generation of the meddies followed by floats Am166 and Am129 (corresponding to the period when there was a great number of floats in water; see Fig. 4) is 42\(T_f\). On the other hand, the minimum elapsed time between two generations is \(O(6T_f)\). In the tank detachment is defined by the time when the zero line of vorticity (surrounding the current and the detaching anticyclone) breaks and originates an isolated blob of negative relative vorticity. The generation period is thus the time separating two such successive breaks. It appeared to vary depending whether or not there was a residual current in the tank or other vortices in the vicinity of the cape. For \(Ro O(0.1)\), in cases when there was almost no residual current, the generation period was in the range \([9\text{–}31 T_f]\). Some reasons can be thought for the discrepancies in the period of formation estimates from in situ and model data. Time variations in the Mediterranean Undercurrent and the possible formation of cyclonic structures in the Portimão canyon region, like those described in section 3, can eventually affect the period of formation of meddies at Cape St. Vincent.

5. Summary and conclusions

The AMUSE, CANIGO, and AMPOR float datasets provide information about the characteristic radius, rotation period, translation velocity from the cape and relative vorticity of meddies generated at Cape St. Vincent. Furthermore, both float datasets show that cyclones are associated with the anticyclones. To our knowledge it is the first time that such consistent observations were made close to the meddies generation region. Such a coupling provides a possible mechanism for the detachment of an anticyclonic lens from the cape by the formation of a dipolar structure, which is self-propelling. The tank experiments show that such a lens production at the cape is part of a broader set of phenomena ranging from the formation of a growing anticyclone at the cape (not observed in nature) to the formation of a dipolar structure moving away from the cape and to the current hugging the slope all along its movement (both latter cases being observed in nature). From the tank experiments we know that the main parameter for determining the stability of the intermediate water current are the Rossby and Froude numbers (Baey et al. 1995; Baey 1997; Sadoux et al. 2000). These numbers computed from the estimated current width, thickness, and maximum velocity for the Mediterranean undercurrent are in relatively good agreement with the tank values. The discrepancies can be related to the difficulty in estimating the Mediterranean current width and maximum velocity. What is clear both in the experiments and in nature is that a limited range of Rossby and Froude numbers exists for which there is lens production. For the lowest values of these quantities both the tank experiments and the in situ data show that there is no lens detachment. It is most remarkable that, at least at and shortly after the detachment time, the nondimensional lens radius, relative vorticity, and rotation time agree well between the tank experiments and the observations. However, the damping is much larger in the tank than in nature due to the higher frictional effects, which lead to a greater lateral entrainment in the tank and thus to an increase in lens radius and in rotation time (or alternatively, a decrease in maximum rotation velocity of the lens).

From such good agreement one can infer that the tank experiments provide interesting and useful insights on the dynamics of the Mediterranean intermediate current, for the lens generation process, and for the initial movement of the lenses away from the cape. The lens detachment occurs only when the upstream current is unstable. This instability proved to be a mixed baroclinic and barotropic instability (Baey 1997), leading to anticyclonic lenses formation both along the upstream wall of the cape and at the cape itself. The cape then appears to be a privileged but not the only generation place for
Anticyclonic lenses. This fact can shed some light in explaining Meddy Cadiz (Prater and Sanford 1994), which was found at a longitude to the east of Cape St. Vincent. As the authors proposed, the meddy had to be formed on the slope near Portimão Canyon upstream of Cape St. Vincent. Swallow (1969) years before reported the observation of a cyclonic structure in the Portimão Canyon as well as an anticyclonic blob to the east of Cape St. Vincent. Those structures also could be explained by the instability of the Mediterranean undercurrent upstream of Cape St. Vincent. The experiments show that whatever the Rossby and Froude numbers are, an anticyclonic lens shed at the cape is always accompanied by a cyclonic lens and that the dipolar structure thus created is self-propelling and moves away from the cape in a direction roughly along the line of the upstream wall, that is, westward or northwestward in nature. In the tank experiments, for $0.1 \leq Ro \leq 0.2$, that cyclonic lens was most of the time created in the cape region forming a dipolar structure together with the anticyclone, whereas it seems that in nature, the cyclone is originated upstream from the cape, possibly in the complex Portimão Canyon region. This cyclone is advected by the current and then forms a dipolar structure that moves away from the cape in a direction more or less along the line defined by the upstream coast, as in the tank experiments. However, in these experiments, for $0.1 \leq Ro \leq 0.5$, there were also instances in which a cyclone formed upstream from the cape was advected, coupled with the anticyclone at the cape, and then moved with it as a dipolar structure. In the tank experiments, as well as in the in situ observations, the lifetime of the cyclone is much smaller than that of the anticyclonic lens, and the two tended to separate from each other as they move away from the cape. It is noticeable that, despite the short number of cyclone observations, the order of magnitude of the cyclonic vorticity (or its ratio to the anticyclonic vorticity) shows good agreement between the tank experiments and the observations. The tank experiments, thanks to the use of the 3D CIV measurements, provide information about the complete velocity, and hence vorticity field in a volume of water, and thus is a powerful complementary tool for interpreting the scattered data. In particular, the CIV measurements clearly show that there is a quick upward diffusion of the negative vorticity so that the water above the anticyclonic lens has also a negative anticyclonic vorticity. This feature is also present in some oceanic observations (Pingree 1995; Pailliot et al. 1999; Oliveira et al. 2000). This study exemplifies the usefulness of coupling laboratory and in situ observations for studying mesoscale processes; the observational evidences are used as fully as possible, and the laboratory experiments provide a dynamical framework to interpret and complement them.

Acknowledgments. The authors are indebted to the owners of the sailing vessels Kialoa II and Safa for their support in the field programs of AMUSE and CANIGO/AMPOR projects respectively; to T. Müller and M. Knoll, from the Kiel Institut für Meereskunde, for the mooring and retrieving of the sound sources; and to all our colleagues and students that helped launch the floats and XBTs in the field programs. The AMUSE project was funded by the Portuguese–American Foundation for Development-FLAD (Grant 54/93) and the U.S. National Science Foundation (Grants OCE-91-01033 and OCE-91-00724). The CANIGO project was a European Union MAST III (Contract MAS3-CT96-0060) funded project, which supported both the in situ measurements and the tank experiments. The AMPOR project was financed by the Fundação para a Ciência e Tecnologia (PRAXIS XXI) through contract 2/2.1/MAR/1741.95. N.S. acknowledges the support by the Fundação para a Ciência e Tecnologia (Grant BD/19535/99). Moreover, the LEGI/Coriolis gratefully acknowledges the CNRS support, and two of the authors (SS and DR) would also like to thank H. Didelle and R. Carcel who helped in the experiments as well as A. Fincham who provided the necessary support for the CIV measurements and data processing. This work was partly supported by the grant HPRI-CT-1999-50019 (HYDRIV project) of the European Commission.

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