Post-convection spreading phase in the Northwestern Mediterranean Sea

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Abstract

This is a study about the spreading of newly formed deep waters following open ocean deep convection in the Northwestern Mediterranean Sea. The main results are from the SOFARGOS large scale float experiment initiated in 1994–1995. During the SOFARGOS project, CTD stations and Lagrangian observations of ocean currents were carried out in the Gulf of Lion from December 1994 to July 1995. Hydrological observations confirmed that deep water formation occurred very early during winter 1994–1995 (late December, early January) in conjunction with atmospheric cooling, deep convection penetrating down to 2000 m in the so-called Medoc area. Numerous eddies (both anticyclonic and cyclonic) drifted away from the convection area and advected newly formed deep waters far away from the source region. In particular, compact anticyclones appeared to be the most coherent (long-lived) eddies and capable of transporting newly formed Western Mediterranean Deep Waters several hundreds of kilometers away from the convection area. Characterized by an inner core of about 5 km in radius, these eddies are submesoscale features in the outer domain and appear as key elements of the open ocean convection processes. During their long journeys, these eddies interacted with larger scale features such as the Northern Boundary Current, the North Balearic Front, topographic Rossby waves, and Sardinian eddies. These interactions influenced the long-term behavior of the eddies (mean drift, composition) and represented an important part of (1) the spreading phase following deep convection and (2) the large scale thermohaline circulation.

Keywords: Deep convection; Submesoscale eddies; Western Mediterranean Sea

1. Introduction

Open-ocean deep convection usually happens in winter in very few regions of the world ocean resulting in oceanic deep water formation. Evidence for this process has been found in the Mediterranean Sea, the Weddell Sea, the Labrador Sea and the Greenland Sea. Its role in forming deep ocean water masses is well documented, but details about what is occurring during the different phases of convection are still unclear. In particular, the spreading of the newly formed deep waters following deep convection is poorly understood. The first academic description of open-ocean deep convection, provided by Medoc Group (1970) in the
Northwestern Mediterranean Sea, is still commonly used today in all studies concerning deep convection processes, in all locations of deep water formation. Deep convection in the Gulf of Lion was described by the Medoc Group in three successive phases:

- **The preconditioning** of the area by a cyclonic gyre circulation in the whole Northern basin of the Western Mediterranean Sea producing a doming of isopycnals toward the surface at the center of the gyre and exposing a large body of weakly stratified waters to local cooling and evaporation due to Mistral and Tramontane winds (dry and cold winds blowing over the Gulf of Lion from the Alps and Pyrénées mountains, respectively).

- **The vertical mixing** due to buoyancy loss generated by intense surface cooling and evaporation reaching 1000 W/m² for short periods, creating a weak stability of the water column and ultimately allowing overturning.

- **The spreading** phase of the newly formed deep waters away from the formation site.

In addition, further studies have also shown the important role of baroclinic instability in relation to deep convection. This role was first proposed for the Northwestern Mediterranean Sea by Killworth (1976, 1979) and Gascard (1977, 1978). In brief, the violent atmospheric forcing events triggering deep convection in the center of the preconditioned area resulted in a stationary chimney of mixed and dense waters. At a later stage, once the atmospheric forcing had considerably lessened, the chimney became highly unstable baroclinically and eddy shedding occurred rapidly, followed by a general breakup of the chimney on a time scale of a few weeks. In 1975 and 1976, Gascard (1978) and Gascard and Clarke (1983) observed eddies of about 5 km radius associated with a baroclinic wave-like pattern at the periphery of the convection area in the Northwestern Mediterranean and the Labrador Sea, respectively. Gordon (1978) observed in 1977 a compact and coherent eddy related to deep convection in the Weddell Sea which survived for several months.

Killworth (1976, 1979) and Gascard (1977, 1978) also emphasized the fact that the chimney formation could be due to the presence of preexisting eddies, formerly generated by baroclinic instability. These mesoscale eddies characterized by a doming of isopycnals toward the surface would precondition for local deep convection at a smaller scale than large scale preconditioning that occurs within a gyre circulation.

We have to point out that baroclinic instability was presented at that time as a mechanism that could occur throughout the deep convection process, from the preconditioning to the spreading phase. In particular, the baroclinic instability mechanism would also contribute to vertical mixing by inducing vertical velocities.

During the late eighties, observations made by Schott and Leaman (1991) in the Northwestern Mediterranean Sea and by Schott et al. (1993) in the Greenland Sea with moored ADCPs gave birth to new concepts to describe deep convection, allowing a different role for the mesoscale eddies. Intense vertical velocities O(10 cm/s) were observed in cells with horizontal and vertical scales of O(1 km). Stimulated by these observations, Jones and Marshall (1993) and Maxworthy and Narimousa (1994) developed a theory, supported by both numerical modeling and tank experiments (also see Marshall and Schott, 1999 for a review), to explain these so-called “plumes”. The experiments helped to characterize and better understand the vertical mixing phase of these small-scale mixing agents. A **Mixed Patch** would result from an integral effect of these non-penetrative “plumes” (Send and Marshall, 1995), characterized by a horizontal length scale directly related to the surface buoyancy fluxes induced by the strong meteorological forcing, and balanced by lateral buoyancy fluxes due to eddies (Visbeck and Marshall, 1996). This **Mixed Patch** would then break up by baroclinic instability, after the violent forcing events had ceased (Legg and Marshall, 1993; Jones and Marshall, 1997).

Nevertheless, these experiments were all based on initial conditions considering a homogeneous ocean forced by a heterogeneous atmosphere. Indeed, both tank and numerical experiments consisted of a disc-shaped atmospheric forcing with a typical horizontal scale identical to the size of the oceanic **Mixed Patch**. Straneo and Kawase (1998) and Legg et al. (1998) pointed out that this forcing introduced peculiar conditions not likely to occur in the real world, corresponding more to a heterogeneous ocean at small scale forced by a atmosphere which can be considered as homogeneous at these small scale. Accordingly, numerical simulations showed that mesoscale surface intensified cyclonic eddies (presenting a doming of isopycnals toward the surface) could act as mesoscale preconditioners.
triggering local deep convection. Lherminier et al. (1999) extended this concept to subsurface anticyclonic eddies also presenting a doming of isopycnals in the upper layer. Legg and McWilliams (2000) also proposed that the homogenization of the newly formed deep waters was more likely due to the turbulent geostrophic eddy field.

More recently and mainly based on observations, Submesoscale Coherent Vortices (SCV, introduced by McWilliams, 1985) have been discovered in the three deep convection sites: the Labrador Sea (Lilly et al., 1999), the Greenland Sea (Gascard et al., 2002), and the Western Mediterranean Sea (Testor and Gascard, 2003). They have been described as a new mode of deep ventilation in Gascard et al. (2002). They all appear to have similar characteristics (small radius, large aspect ratio and long lifetime on the order of a year) and can act as preconditioners for deep convection the following year. SCVs also appeared to be involved in the large scale circulation of the newly formed deep waters. McWilliams (1985) observed a SCV at mid-latitude in the North Atlantic Ocean characterized by a core made of the deep waters generated by deep convection in the Labrador Sea. Based on two Lagrangian experiments undertaken in the Western Mediterranean Sea SOFARGOS (1994/1995) and MATER (1997/1998), Testor and Gascard (2003) identified SCVs related to deep convection evolving very far \(O(500 \text{ km})\) from the convection area. Highlighted by the discovery of these compact, long-lived eddies far away from the deep convection sites, the spreading phase of newly formed deep waters resulting from deep convection is attracting a lot of interest in the community for understanding the connection of deep convection and thermohaline circulation.

In the present paper, we will describe and analyze the results obtained from the SOFARGOS and MATER experiments, providing a more complete and extended description of the spreading phase following deep convection in the Northwestern Mediterranean Sea. We will focus on the SOFARGOS 1994/1995 experiment and first describe the area under study from a hydrological and meteorological point of view, based on CTD vertical profiles, ECMWF fluxes and Lagrangian float data. Then we will discuss the large scale spreading of the newly formed deep waters by submesoscale eddies and their importance for the thermohaline circulation. We will also examine various aspects of submesoscale eddy dynamics interacting with large and mesoscale events frequently encountered in the Western Mediterranean Sea.

2. SOFARGOS data set presentation

The SOFARGOS project (MAS2-CT92-0046) was jointly carried out by the Laboratoire d’Océanographie Dynamique et de Climatologie (LO-DYC—Paris), the Institut de Ciencies del Mar (CSIC—Barcelona), and the Stazione Oceanografica—Consiglio Nazionale delle Ricerche (SOCNR—La Spezia) from July 1994 to July 1995. The main objective of the SOFARGOS project was to develop and to test a quasi-Lagrangian float prototype designed for observing large scale and mesoscale subsurface currents in the Mediterranean Sea. The SOFARGOS final report was issued in 1996 (MAST2—EU Report, 1996). Here we concentrate on SOFARGOS Lagrangian (float) data and CTD data collected in winter-spring time in the so-called Medoc area for analyzing the spreading of newly formed deep waters resulting from deep convection in the Northwestern Mediterranean Sea.

2.1. Lagrangian float data

Twenty-five isobaric floats like Swallow floats (Swallow, 1955) were deployed. Precisely ballasted, these floats drift according to horizontal currents at a predetermined and quasi-constant depth. They are localized by triangulation using an array of moored acoustic sources transmitting long range low frequency signals (SOFAR: sound fixing and ranging). Floats receive signals transmitted six times per day for 1 year by acoustic sources at known locations and times controlled by stable clocks (10\(^{-7}\)s accuracy). Therefore, they are so-called RAFOS floats (inverted SOFAR technology) in contrast to SOFAR floats emitting acoustic signals. The mean absolute accuracy for float positioning is about 1–2 km, limited mainly by the geometry of the array with respect to the float's locations and uncertainties in sound speed estimation and position of the acoustic sources, but relative precision is much better (\(\sim 300 \text{ m}\)). Floats also measure in situ temperature and pressure every 2h. In addition, tilted vanes attached around the cylindrical float induce rotations around the float’s vertical axis in response to relative vertical displacements of water passing by the float. The combination of the pressure and rotation measurements allow these floats to be considered as Vertical Current Meters.
(VCM, Voorhis, 1971; Gascard, 1973). At the end of their mission, the floats drop a dead weight (~2 kg) to reach the surface and start transmitting data to NOAA satellites equipped with the ARGOS system. Over the entire duration of the SOFARGOS project, four cruises were carried out for the deployment and the recovery of the acoustic sources and floats, and the completion of hydrological surveys in the area (MAST2—EU Report, 1996) involving three different ships (R/V Garcia del Cid from Spain, R/V Thetys II from France, and R/V Urania from Italy).

- Ten RAFOS-VCM floats were launched from R/V Thetys II at the beginning of December 1994 across the Northern Current along 6°E south of Toulon (Fig. 1), all at about the same depth (350 m), to document mainly the large scale circulation of LIW in the Western Mediterranean Sea. They drifted for about 7 months upstream and downstream of the Northwestern Mediterranean convection area.

- Fifteen floats were launched in late January 1995 from R/V Thetys II in the deep water formation zone, the so-called Medoc zone (around 42°N, 5°E) at five different locations forming a cross-like pattern (Fig. 1) and at three different depths: around 350, 600 and 1200 m (minimum and maximum depths: 310 and 1450 m, respectively). They drifted for about 5 months from the deep convection zone and downstream.

![Fig. 1](image-url)
Float depths are presented in Table 1. Most of the 25 floats and 3 out of 6 acoustic sources were retrieved in July 1995 from the R/V Urania. Three acoustic sources were lost because of mooring failures. Floats were located from the remaining sources but with a loss of accuracy.

2.2. CTD data set

CTD casts were performed near the floats’ launching sites during the two cruises carried out by R/V Thetys II in December 1994 in the Northern Current south of Toulon and in January 1995 in the Gulf of Lion. In this paper, we will make an extensive use of the CTD data collected between 21 and 24 January 1995, during the second float deployment in the center of the deep convection area, in addition to the float trajectories collected across the Northwestern Mediterranean basin from December 1994 until June 1995.

3. Medoc area after deep convection events

3.1. Atmospheric fluxes

Strong meteorological forcing (cooling and evaporation), necessary to trigger deep convection, occurred very early during winter 1994–1995. Time series of atmospheric fluxes from ECMWF (European Center for Medium-Range Weather Forecasts) show intense heat losses in December and at the very beginning of January (peak values of more than 800 W/m²). Three meteorological events characterized by large heat fluxes from the ocean to the atmosphere appeared between December 20, 1994, and January 15, 1995 (Fig. 1). By late January, the fluxes had considerably decreased (peak values of ~100 W/m²) and remained weak until late February when fluxes reached intermediate peak values of 100–500 W/m². From late February to mid-March the cumulative heat loss was significantly lower than between mid-December and mid-January. In addition, the floats did not measure any intense vertical velocities, showing only maxima order of 0.5 cm/s (at near-inertial period) when drifting in the convection area just after the second deployment. So it appears that late January corresponds to the end of the violent mixing phase due to deep convective events occurring in 1995.

Subsequently, the 15 floats launched inside the convection region began to drift after the strong meteorological events had passed over the area. The first set of 10 floats that were launched earlier in December remained mostly outside the deep convection region. Some of them reached the convection area at about the same time that the second set of 15 floats was launched. Consequently, all the floats documented mainly the spreading phase following deep convection.

3.2. Water mass identification

CTD casts were performed just after the second float deployment in late January in the center of the Gulf of Lion (see Section 2). The CTD casts confirmed that deep water formation had already occurred (1–2 months earlier than usual) and revealed that deep convection had penetrated down to 2000 m depth in some places, but apparently did not reach the bottom. Corresponding potential temperature vertical profiles and θ–S diagrams are plotted in Fig. 2:

- “Stratified” stations exhibit surface salinity minima corresponding to Modified Atlantic Water (MAW) at about 38.15, characterized by a large variability in the θ–S values. Underneath this surface layer, one can find maxima of potential temperature and salinity at about 13.40 °C and 38.57 typical of the Levantine Intermediate Water (LIW) and, at greater depth, (old) Western Mediterranean Deep Water (WMDW) with potential temperature of 12.78 °C and salinity of 38.45.
- “Mixed” stations show homogeneous vertical profiles of potential temperature and salinity at intermediate depths extending in extreme cases from 0 down to 2000 m depth. This is typical of the newly formed Western Mediterranean Deep Water (nWMDW) which is characterized by a potential temperature of about 12.90 °C and a salinity of 38.48 at all depths. Since deep convection did not reach the bottom during this
particular winter, these “mixed” stations also indicate the presence of old WMDW \((12.78 \, ^\circ C, 38.45)\) below nWMDW.

The nWMDW is fresher, colder and denser than LIW but significantly warmer (+0.1 °C) and saltier (+0.03) than old WMDW, although densities of both deep waters are very close to each other. MAW (with salinities around 38.15 in this region) is progressively modified along its past path from Gibraltar Strait and a great variability can be observed in its characteristics depending on its age and path. In winter, mixed layer processes likely due to atmospheric forcing combined with frontal dynamics (Yoshikawa et al., 2001) can finally turn this surface water into a mode water, called Winter Intermediate Water (WIW). This water mass is characterized by a minimum in potential temperature at about 12.65 °C (even colder than WMDW) and a salinity <38.30. Hence, we consider WIW to be an ultimate state of MAW, forming a thick and homogeneous surface mixed layer. It does not correspond to a “mixed” station from a deep convection point of view, but rather to a “stratified” profile according to our classification. Considering the floats’ drifting depths (Table 1) and these hydrological specifications, we can distinguish two sets of drifting floats according to their depth ranges: 310–660 and 1100–1450 m. In each set, old/stratified and new/mixed waters are fairly recognizable by the temperatures of the water masses measured (in situ) by deep floats. The typical temperature of nWMDW is ~12.9 °C at all depths.

- In the deep layers (1000 m to the bottom), nWMDW is characterized by warmer temperatures (~12.90 °C) than old WMDW (~12.80 °C).
- At intermediate depths (350–650 m) and out of the convection area, temperatures of 12.90–13.05 °C are typical of nWMDW. Such cold temperatures indicate the influence of nWMDW on the water properties at this level, which are in contrast to the properties of the LIW (typical temperatures of ~13.10–13.40 °C at these depths in the Northwestern Mediterranean Sea, out of the convection area).

3.3. Main features of the convection area

The sections shown in Figs. 3a and 3b along 4°30'E and 42°15'N respectively depict the hydrological structure of the northern domain of the deep convection area during the post-convection phase.

3.3.1. The Northern Current

The Northern Current originates in the channel of Corsica and follows the topography westwards. This current advects stratified waters, mainly coming from the Tyrrhenian Sea, along the continental slope. In Fig. 3a, the “stratified” station 15 documents this boundary current with a MAW surface layer and a pronounced maximum of potential temperature (13.4 °C) and salinity (38.58) typical of LIW lying above WMDW. As revealed by floats drifting in the Northern Current at 350 m depth in the LIW core, this current closely follows the 500–1000 m isobath. Typical velocities of 5 cm/s are revealed by floats at 350 m depth. Station 14, which also presents a pronounced maximum of temperature and salinity due to the presence of LIW, is situated at about 25 km from the 500 m isobath and shows that the Northern Current can
have a lateral extension of about 30 km at 4°30'E. Here the water column is stratified and presents an internal radius of deformation \( \left( R_d = N H/f \right) \), where \( f \) is the planetary vorticity, \( N = 2 - 5f \) is the Brunt–Väisälä frequency, and \( H = 2-2.5 \text{ km} \) is the distance from the surface to the bottom) reaching about 10–15 km.

3.3.2. The Mixed Patch

In the southern part of the section, lying above WMDW (in dark blue), the large amount of nWMDW at 12.90°C in the south (in blue) extending from the surface down to approximately 2000 m depth confirms that deep convection did occur and reached great depths. This corresponds to the Mixed Patch (“mixed” stations 21, 11 and 1 in Fig. 3). The Mixed Patch is rather homogeneous and shows quite a different stratification from the Northern Current, corresponding to an internal radius of deformation equal to 1–3 km. Float data combined with CTD casts allowed us to delineate the Mixed Patch approximately. Its northern limit has a frontal shape which extends from 42°15'N, 4°10'E to 42°30'N, 5°30'E. The eastern and western limits are approximately situated along 4°10'E and 5°30'E, respectively. The southern limit of the Mixed Patch is not well defined, neither by floats nor by CTDs, but we estimated it at about 41°45'N. The corresponding volume of newly formed deep waters is estimated to be about \( 9 \times 10^{12} \text{ m}^3 \) in 1995.

3.3.3. The Transition Zone

Between 42°15'N and 42°30'N along 4°30'E (Fig. 3), CTD casts highlighted a transition area between the mixed and the stratified regions as in Schott et al. (1996). Based on additional CTD casts and float data, this Transition Zone surrounded the Mixed Patch at the western, northern and eastern boundaries and was about 30 km wide. Stations 12, 13, and 14 presented some perturbations compared to typical profiles (“mixed” and “stratified”), corresponding to a more gradual evolution of the water mass distribution between the Northern Current and the Mixed Patch (Fig. 3a).

In this Transition Zone, close to the Northern Current, the water column is unperturbed, with the exception of the surface layer mainly composed of WIW instead of MAW (Fig. 3a, rather “stratified” station 14), whereas close to the Mixed Patch the LIW is quite eroded and located closer to the surface and above homogeneous layers of nWMDW (Fig. 3a, stations 12 and 13). nWMDW is overlapped by eroded LIW, indicating lateral exchanges between the Mixed Patch and the surroundings. In this Transition Zone, the deformation radius ranges from 3 to 7 km. The West–East potential temperature section (Fig. 3b) also confirms active lateral exchanges by interleaving between the more “mixed” (stations 1, 2, 4, and 5) and the more “stratified” (stations 18, 6, and 8) waters.
3.4. **Baroclinic instability in the Transition Zone**

3.4.1. *In situ observations*

As pointed out in Section 3.1, the SOFARGOS float data sample mainly the post-convection phase. Just following deep convection events and before any export of nWMDW was observed far from the convection area, the float data reveal a very rich eddy activity in the convection area involving both anticyclonic and cyclonic circulations, in particular in the *Transition Zone* between the *Mixed Patch* and the surroundings. Eddies can be identified by looping trajectories and the water mass composing their cores, by temperature measurements. Fig. 4 shows a detail of float trajectories drifting in the *Transition Zone* between 350 and 1400 m during 3 weeks in February 1995. Despite significant additional variability, one can see the alternation of mixed and stratified pools along bathymetric contours in Fig. 4b, where trajectories are color-coded according to potential temperature. Within the upper 1000 m, mixed water parcels are colder than the stratified surroundings and warmer below (see Section 3.2).

Small cyclonic and anticyclonic eddies are embedded in these pools with corresponding positive...
and negative vertical velocities, respectively (Fig. 4a). The vertical velocities have been low-pass filtered \((T > 1 \text{ day})\) to remove the signals corresponding to internal waves and the subinertial components are about \(\pm 0.05 \text{ cm/s}\). At these depths, the stratified waters \((\text{LIW less dense than nWMDW})\) tend to upwell and the mixed waters tend to sink.

The eddy naming convention is to associate one or two letters and a number for each eddy. “A” stands for anticyclones (for instance A3) and “C” for cyclones. A following “S” stands for the “stratified” waters in their cores (for instance CS4) in contrast to nWMDW eddies. Two of these eddies having already been mentioned in Testor and Gascard (2003) under the denomination of SCV-N1 and SCV-N2, their old names are appended to the names chosen according to this new convention.

Note that the distance (about 25 km) between the centers of two paired eddies of opposite signs is rather regular. On the west–east section (Fig. 3b) carried out in late January in the Transition Zone, one may also associate the alternation of “mixed” and “stratified” stations to the vertical structure of a developing wave. The length scales associated with interleaving of mixed and stratified waters are in good agreement with float observations, although the horizontal spacing between the CTD casts (about 10 km) is certainly insufficient to resolve it properly.

These bipolar features and the corresponding vertical velocity structure suggest a baroclinic instability wave pattern already mentioned in the literature (see Killworth, 1976; Gascard, 1978; Madec et al., 1991; Hermann and Owens, 1993; Jones and Marshall, 1993; Send and Käse, 1998) for the lateral exchanges between the Mixed Patch and the surroundings. Based on these observations, we will apply a simple model of baroclinic instability to the Transition Zone to compare with in situ measurements.

### 3.4.2. Linear model

To characterize instabilities possibly resulting from the frontal structure observed in the Transition Zone, we will introduce a Phillips model of baroclinic instability (Phillips, 1954). The topographic effect can be linearized like a \(\beta\)-effect \(\|\beta_T\| \approx (f_0/H_0)\|\nabla h\|\sim O(10^{-9} - 10^{-10})\). The planetary \(\beta\)-effect \(\sim O(10^{-11})\) is neglected. In a quasi-geostrophic approximation and a two layer model, we consider a basic state where \(U_1\) and \(U_2\) are independent of \(y\) (south–north direction). The equations of linearized perturbations can be written as

\[
\begin{align*}
\left[ \frac{\partial}{\partial t} + U_n \frac{\partial}{\partial x} \right] \left[ \Delta \phi_n + (-1)^n F_n(\phi_1 - \phi_2) \right] \\
+ \left[ \beta_T - (-1)^n F_n(U_1 - U_2) \right] \frac{\partial \phi_n}{\partial x} &= 0, \\
n &\in \{1; 2\},
\end{align*}
\]

where \(H_n\) is the thickness of layer \(n\), \(\rho_2 - \rho_1\) is the density jump at the interface, and \(F_n = f_0^2/\rho_0([\rho_2 - \rho_1]/\rho_0)H_n\). Using the internal radius of deformation of a two layer model \(R_0 = (1/f_0)\sqrt{((\rho_2 - \rho_1)/\rho_0)(H_1H_2/(H_1 + H_2))}\), we can express \(F_n\) as a function of the deformation radius \(R_n\) for each layer \(n\) and layer thicknesses: \(F_n = 1/R_n = (H_1/(H_1 + H_2))/1/R_0, i \neq n, (i, n) \in \{1; 2\}\). The stream function of the perturbation in the layer \(n\) is \(\phi_n = A_n \cos \beta x e^{ik(x-c)}\) where \(k\) is the wave number along the front direction, \(l\) is the wave-number across the front, \(x\) is the west–east direction, \(t\) is time, and \(c\) is the phase velocity of the wave.

Eq. (1) is a system whose determinant must vanish to obtain nontrivial solutions for \(\phi_1\) and \(\phi_2\). This leads to the dispersion relation

\[
c = U_1 + \frac{UK^2(K^2 + 2F_2) - \beta_T(2K^2 + F_1 + F_2)}{2K^2(F_1 + F_2)} \ldots
\]

\[
\pm \frac{[\beta^2_t(F_1 + F_2)^2 + 2 \beta_T UK^4(F_1 - F_2) - K^4U^2(4F_1F_2 - K^4)]^{1/2}}{2K^2(F_1 + F_2)}
\]

where \(K^2 = k^2 + \hat{f}^2\) and \(U = U_1 - U_2\).

The growth rate is given by \(\sigma = K \text{Im}(c)\), where \(\text{Im}(c)\) denotes the imaginary part of \(c\), has a maximum value (corresponding to the wave of maximum instability) with adimensional wavenumber \(K\) equal to 1.31 taking \(R_0 = 5 \text{ km}, f_0 = 10^{-4} \text{ s}^{-1}, f_0/H_0 \times \nabla h/\partial y = 4.55 \times 10^{-10} \text{ m}^{-1} \text{ s}^{-1}, H_1 = 500 \text{ m}, H_2 = 1700 \text{ m}, U_1 = 30 \text{ cm/s},\) and \(U_2 = 3 \text{ cm/s}\).

These are typical values corresponding to 500 m depth deduced from observations and geostrophic calculations in the Transition Zone. The slope of isopycnals within the Transition Zone is approximately \(x = -(\partial \rho/\partial y)/(\partial \rho/\partial z)\sim 1\) to 3% around 500 m depth.

That leads to a wavelength of \(\lambda \approx 50 \text{ km}\), an e-folding time scale of 5.6 days (period \(T = 2\pi/\sigma_{\text{max}} \approx 8\) days), and a phase speed of \(\text{Re}(c) \approx 10 \text{ cm/s}\) (Re(c) denotes the real part of \(c\)) for the most unstable baroclinic wave solution. The
topographic effect does not significantly change the wavelength of the most unstable wave but has a weak stabilizing effect by prohibiting baroclinic instability at large wavelengths—see Fig. 5. These results are in good agreement with observations of anticyclonic eddies composed of nWMDW and associated cyclonic vortices composed of more stratified waters (A3-CS3-A4-CS4, Section 3.4.1). So, the Transition Zone separating mixed waters from outer more stratified waters appears to be a site where baroclinic instability can grow.

This has already been mentioned by Gascard (1978) for this area. Early in March 1975, a cyclonic eddy (C75) 10 km in diameter and 2–3 days period was observed with isobaric floats drifting at 600, 1000 and 1200 m depth. It was centered at about 42°05′N, 4°30′E. At the same time, an anticyclonic circulation (A75) appeared centered 20 km to the west of the center of the cyclonic eddy. These earlier observations are very similar to the present observations of dipolar structures (CS4-A4-CS3-A3). According to Gascard (1978), the time and space bipolar structure A75-C75 (30–40 km lengthscale, 5 days growth rate) was in agreement with a two layer model of baroclinic instability without \( \beta_T \) effect (Tang, 1975). The Medoc 1975 eddy was located at the southern edge of the Mixed Patch instead of the northern edge for SOFARGOS 1995 observations, indicating this process can occur all around the Mixed Patch.

In summary, there is new evidence that baroclinic instability contributes locally to (a) the sinking and the spreading of nWMDW moving away from the Mixed Patch and (b) the upwelling of stratified waters moving from the Transition Zone towards the Mixed Patch and restratifying the area.

4. nWMDW eddies

Figs. 6 and 7 depict the large scale spreading of nWMDW as revealed by SOFARGOS floats in the intermediate and deep layers. In Fig. 6, one can clearly see (as revealed by CTD casts in Fig. 3) the Northern Current composed of warm waters at intermediate levels (see Section 3.2), following the continental slope; the Mixed Patch (in blue); and the Transition Zone in between composed of mixed and stratified waters.

Typical velocities of the Northern Current around 350 m are about 5 cm/s. Advection by the boundary current is important for setting the properties of the water in this region where water masses are constantly renewed by stratified waters coming from the East. Hence, CTD casts correspond mainly to stratified profiles (station 15, Fig. 3). Fig. 6 also reveals eddies and meanders more likely resulting from instabilities of this boundary current (Crépon et al., 1982). Fig. 7 shows float data from the deep layers, with a Mixed Patch centered above the Rhône’s deep sea fan (colored in red–yellow compared with the stratified waters in dark blue, see Section 3.2).

The month following deep convection, nWMDW was mainly advected by eddies towards the southwest at a mean speed of about 5 cm/s. They seemed to follow the general cyclonic circulation around the basin, within 50 km offshore the continental slope and parallel to the Northern Current. Most of the nWMDW turned eastwards in the vicinity of point J when the general pattern of circulation became more complex.

4.1. nWMDW anticyclones and cyclones

As already mentioned in Section 3.2, nWMDW is fairly recognizable on the temperature records of a float, depending on its depth: it corresponds to a cold anomaly at intermediate depths and a warm anomaly at great depths. In this section, we will
characterize these eddies with dynamical parameters (radius, relative vorticity, lifetime, and vertical extent) and analyze float trajectories according to the following three steps:

- An estimate of the eddy center locations is made by applying low-pass filtering to float trajectories, allowing an approximation of the radial distribution of orbital velocity. One can then estimate basic parameters such as relative vorticity and eddy radius, defined as the distance at which the maximum mean orbital velocity is observed. The radial structure is approximated by quasi-solid body rotation near the center, in order to estimate relative vorticity near the center of the eddy. Floats drifting near a float trapped within an eddy, at about the same depth but not too much influenced by the eddy rotations, also provide upper bounds for the horizontal eddy scale. This is particularly important when an eddy has not been properly sampled to detect the maximum orbital velocity in the radial structure.

- Estimates of the lifetime of the eddies are given by the duration of time the floats stayed trapped within them. Of course, this does not necessarily mean that the eddies disappear when floats leave them or that the eddies form when floats begin to drift inside of them. Hence, the eddies' lifetime estimated from the floats is underestimated but should scale with actual eddy lifetime.

- Floats drifting near a float trapped within an eddy but at a different depth provide information about the vertical eddy scale. Floats drifting together at different depths in the same eddy provide a lower bound for the eddy vertical extent. A float passing above or below a float trapped within an eddy, and not influenced by the rotations, also gives limits to the eddy's vertical extent. NOAA/AVHRR SST images provide information about the surface (given image resolution of 1 km and cloud cover). Hence they can be used to determine if an eddy, sampled by deep floats, has a surface signature or not which also helps to estimate the eddy’s vertical scale.
Based on these estimates, we identified eight different eddies transporting nWMDW out of the Mixed Patch. One can clearly see these eddies in Figs. 6 and 7 by looking at float trajectories and float temperatures typical of nWMDW. Eddies of both signs were identified: A1, A2, A3, A4, A5 for anticyclones and C1, C2 and C3 for cyclonic eddies (see Tables 2 and 3). These latter cyclonic eddies, having a core typical of nWMDW, are distinct from the two cyclones (CS3 and CS4) presented in Section 3.4 that appeared to be involved in the restratification process.

All the anticyclones composed of nWMDW were very similar to the one presented in Fig. 8 (right panel). Fig. 8 (left panel) is typical of cyclones composed of nWMDW. Hence, we may classify eddies involved in the spreading of nWMDW in these two groups: the nWMDW anticyclones and the nWMDW cyclones. They all appeared as subsurface eddies since we had no evidence of SST anomalies related to these eddies. Nevertheless, we suspect they have a large vertical scale of at least 1 km centered at around 1000 m depth, as indicated in Tables 2 and 3. They were characterized by a rotational period of a few days and a high vorticity anomaly (relative vorticity $|\zeta|/f$) implying high Rossby numbers (see Tables 2 and 3 for a detailed description of each eddy). All these eddies had a radius of 5–10 km, which is comparable to the internal radius of deformation in the convection area. Outside the convection area this scale is significantly smaller than the deformation radius. Therefore, these nWMDW eddies correspond to submesoscale features for the Northwestern Mediterranean basin when found outside of the convection zone.

Some of the nWMDW anticyclones have been followed for more than 5 months and observed very far from the Medoc area (300 km), whereas we had no evidence that nWMDW cyclones could last more than 2 months or could advect nWMDW much farther away than 100 km from the convection zone (Tables 2 and 3 and Figs. 6 and 7). Based on another experiment (MATER) undertaken 2 years later and revealing nWMDW anticyclones in the Algerian basin, nWMDW anticyclones may propagate even further (Testor and Gascard, 2003), implying a lifetime greater than 1.5 year for the

![Figure 7. SOFARGOS experiment, trajectories of five RAFOS-VCM floats situated between 1100 and 1450 m depth from December 3, 1994, until July 1, 1995, color-coded according to potential temperature. Float numbers indicate the ends of the trajectories. Black arrows indicate eddies that were carefully examined to be typical of the newly formed deep waters. Note that the newly formed deep waters are colored in red/yellow/green and the old deep waters in deep blue. Approximate location of point J is indicated by a green star.](image-url)
The nWMDW eddies are characterized by their large distances from the mixed patch, a lifetime of about 1 year, and predominantly negative buoyancy anomaly. These eddies are thought to be formed by the decoupling of deep convection processes, resulting in the formation of a density anomaly that leads to the creation of SCVs. Eddies identified as nWMDW can be formed by plumes or breaking internal waves. However, the core of nWMDW eddies is composed of mixed waters, suggesting that some kind of mixing had occurred during the period of deep convection that occurred 1 month before the observations. The eddies are constrained to form at the same time float observations are made but could have been formed earlier. In particular, the core of nWMDW eddies is composed of mixed waters, some kind of mixing had to occur during the period of deep convection that occurred 1 month before we started our observations (Fig. 1). According to McWilliams (1985), SCVs may be created by diapycnal mixing followed by a gradient wind adjustment. In case of deep convection, a localized density anomaly, possibly leading to the formation of an nWMDW eddy, could be generated by breaking internal waves as suggested by StGuily (1972) and Gascard (1973) or by diapycnal mixing of water masses resulting from baroclinic instability. This latter process would potentially be active from the preconditioning to the spreading phase of deep convection as also claimed by Gascard (1978).

On the other hand, mixing agents like plumes (Voorhis and Webb, 1970; Schott and Leaman, 1991) are less likely to be good candidates. Plumes occur during the violent mixing phase of deep convection in a quasi-homogeneous fluid, but they are thought to be non-penetrative (Marshall and Schott, 1999). Hence, other mechanisms have to be involved to create a local density anomaly resulting in the formation of nWMDW eddies.

### 4.2. Origin of the submesoscale nWMDW eddies

#### 4.2.1. Diapycnal mixing

As we already said, nWMDW eddies were not constrained to be formed at the same time float observations were made, but could have been formed earlier. In particular, the core of nWMDW eddies is composed of mixed waters, some kind of mixing had to occur during the period of deep convection that occurred 1 month before we started our observations (Fig. 1). According to McWilliams (1985), SCVs may be created by diapycnal mixing followed by a gradient wind adjustment. In case of deep convection, a localized density anomaly, possibly leading to the formation of an nWMDW eddy, could be generated by breaking internal waves as suggested by StGuily (1972) and Gascard (1973) or by diapycnal mixing of water masses resulting from baroclinic instability. This latter process would potentially be active from the preconditioning to the spreading phase of deep convection as also claimed by Gascard (1978).

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### Table 2

<table>
<thead>
<tr>
<th>Eddy identifier</th>
<th>A1</th>
<th>A2</th>
<th>A3 (SCV-N1)</th>
<th>A4</th>
<th>A5 (SCV-N2)</th>
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<td>#00, #11</td>
<td>#93, #99</td>
<td>#03, #89</td>
<td>#07</td>
</tr>
<tr>
<td>$P_{\text{float}}$ (dbar)</td>
<td>1450, 1100</td>
<td>1400, 600</td>
<td>1400, 350</td>
<td>1400, 350</td>
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</tr>
<tr>
<td>$T$ (days)</td>
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<td>2.8 ± 0.5</td>
<td>3.5</td>
<td>2.9 ± 0.7</td>
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<tr>
<td>$u$ (cm/s)</td>
<td>6.2 ± 2.5</td>
<td>4.0 ± 3.6</td>
<td>8.0 ± 6.1</td>
<td>4.7 ± 2.3</td>
<td>7.2 ± 5.2</td>
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<tr>
<td>$r$ (km)</td>
<td>6.4 ± 2.9</td>
<td>2.0 ± 2.3</td>
<td>5.0 ± 1.0</td>
<td>3.7 ± 1.5</td>
<td>5.2 ± 0.5</td>
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<tr>
<td>$T_{\text{obs}}$ (days)</td>
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<td>&gt;150</td>
<td>&gt;4</td>
<td>&gt;150</td>
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<tr>
<td>$V_{\text{dift}}$ (cm/s)</td>
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<td>5.2 ± 3.1</td>
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<tr>
<td>$H$ (m)</td>
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<td>&gt;1100</td>
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<tr>
<td>$\zeta$ (10$^{-4}$ s$^{-1}$)</td>
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<td>−5.4</td>
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</tr>
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<tr>
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<td>−0.56</td>
<td>−0.56</td>
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<td>−0.53</td>
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<td>&gt;0.4</td>
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</tbody>
</table>

$n_{\text{float}}$ is the ID of the floats revealing these eddies, $P_{\text{float}}$ is float depth, $T$ is an estimate of the rotation period, $u$ is an estimate of orbital velocity, $r$ is the distance of floats to the eddy center, $T_{\text{obs}}$ is the period during which these eddies were observed, $V_{\text{dift}}$ is an estimation of translation speed, $H$ is an estimate of the vertical extent, $R$ is an estimate of radius, $U$ is an estimate of the maximum orbital velocity, $R_o = \zeta/f$ is the Rossby number, and $B_o = (NH/2fR)^2$ is the Burger number.

### Table 3

<table>
<thead>
<tr>
<th>Eddy identifier</th>
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<th>C3</th>
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<td>2.5 ± 1.1</td>
<td>5.0 ± 1.5</td>
</tr>
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<td>$r$ (km)</td>
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<td>$T_{\text{obs}}$ (days)</td>
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<td>&gt;68</td>
<td>&gt;20</td>
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<tr>
<td>$V_{\text{dift}}$ (cm/s)</td>
<td>5.3 ± 2.3</td>
<td>4.0 ± 2.2</td>
<td>2.1 ± 1.1</td>
</tr>
<tr>
<td>$H$ (m)</td>
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<td></td>
</tr>
<tr>
<td>$\zeta$ (10$^{-4}$ s$^{-1}$)</td>
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<td>5.8</td>
<td>3.4</td>
</tr>
<tr>
<td>$R$ (km)</td>
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<td>(&gt;8)&lt;12</td>
<td>(&gt;10)&lt;12</td>
</tr>
<tr>
<td>$U$ (cm/s)</td>
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<tr>
<td>$R_o$</td>
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</table>

See Table 2 for explanation.
from diapycnal mixing that will finally adjust to form an nWMDW anticyclone.

4.2.2. Baroclinic instability
Besides its role in diapycnal mixing, could baroclinic instability also be a mechanism involved in the formation of nWMDW eddies? This question arises from the fact that A3/SCV-N1 was first observed embedded in a wave pattern developing in the Transition Zone, described in Section 3.4 (Fig. 4), which was consistent with the Phillips model of baroclinic instability applied to this area.

Indeed, the Phillips model leads to a slanted convection pattern roughly describing what happens in the Transition Zone. It is only a simple and linearized model describing the beginning phase of the instability for an idealized ocean (smooth initial conditions typical of the gyre circulation) and cannot fully describe finite amplitude eddies.

The real “initial conditions” observed in the real ocean should involve very active mesoscale structures superimposed on this gyre circulation, and the unstable baroclinic waves developing within the Transition Zone might very well interact with the mesoscale ambient field, redistributing vorticity in a wave pattern similar to the one observed in February 1995 (see Section 3.4). In such a case, Section 3.4 simply documents how these nWMDW eddies interact with an unstable baroclinic wave in the Transition Zone. Indeed, baroclinic instability could have contributed to the formation of nWMDW eddies, but the full eddy formation process is certainly more complex.

4.2.3. Preconditioning
The float trajectories showed high eddy activity (scales of 5–10 km radius) in the convection area during the post-convection phase, but this might not result just from deep convection events. Being baroclinically unstable (Crépon et al., 1982), the Northern Current is a prime candidate to frequently generate mesoscale structures, detaching from the boundary current and moving offshore toward the deep basin interior before, during, and after deep convection events. However, baroclinic instability may also induce eddies anywhere else in the gyre, and this is undoubtedly a major source for the observed eddy activity.

Gaillard et al. (2000) observed an important patchiness in the convection area throughout the winter in 1992, both at and below the surface. They associated this with numerous eddies that had evolved in that region before, during, and after deep convection events. They gave an horizontal scale of $O(30-40)$ km for these eddies, which is larger than the scale of the eddies we observed in the
post-convection phase. Nevertheless, using a Kalman filter for integrating a composite data set gathering tomography and a limited amount of floats and current meters data, Gaillard et al. (2000) could certainly not resolve length scales of 10 km, so this discrepancy could be a matter of resolution. Still, this is a good indication of very high eddy activity during all the deep convection phases.

Undoubtedly, one must take into account this active mesoscale eddy field considering that such eddies can act as preconditioners for deep convection. According to Killworth (1979), Legg et al. (1998), Straneo and Kawase (1998) and Legg and McWilliams (2000), surface intensified cyclonic eddies favor deep convection locally since their surface layers are denser and less stratified. Also, Lherminier et al. (1999) have shown that subsurface anticyclonic eddies characterized by quasi-homogeneous waters around the eddy center (their structure in density differs from cyclonic eddies only in the deep layers) may also act as preconditioners for triggering deep convection locally. Preconditioning is certainly a clue to help understand the formation of both nWMDW cyclones and anticyclones, in addition to diapycnal mixing and baroclinic instability processes.

Considering that nWMDW anticyclones might last over 1 year (SCVs), these features might survive from one winter to the next and might appear again in the convection area a year later. In so doing, they might create a hysteresis “memory” effect in the ocean for locally preconditioning deep convection. However, it seems this memory effect is less important for preconditioning deep convection in the Western Mediterranean Sea than it is for the Greenland Sea (Gascard et al., 2002) since, unlike the submesoscale eddies observed in the Greenland Sea that were also related to deep convection, the nWMDW eddies appeared to have a strong tendency to migrate far away from the formation area.

Fig. 9. SOFARGOS experiment, trajectories of all RAFOS-VCM floats from December 3, 1994, until March 21, 1995, color-coded according to potential temperature. Float numbers indicate the ends of the trajectories. Float drifting depths are indicated in Table 1. Black arrows indicate eddies that were carefully examined to be typical of the newly formed deep waters. Note that the trajectories in blue (~12.9°C) correspond to the newly formed deep waters everywhere and that trajectories in green–yellow out of the convection area also indicate the presence of this water mass that has been little warmed by the surrounding LIW. Trajectories in red and purple/deep blue correspond to stratified waters (intermediate and deep, respectively)—see Section 3.2. Approximate location of point J is indicated by the green star.
5. Large scale nWMDW spreading

In Section 4, we have demonstrated that most of nWMDW eddies escaped from the Mixed Patch toward the southwest and then turned eastwards, downstream from point J (location in Figs. 6 and 7). In the following, we will present different kinds of interactions between the submesoscale nWMDW eddies and larger scale features and indicate how these interactions might influence the large scale and long-term behavior of these eddies, and ultimately the large-scale spreading of the newly formed deep waters in the Northwestern Mediterranean Sea.

In the first part (Section 5.1) we will describe more in detail the spreading of nWMDW in the vicinity of the convection region while nWMDW eddies were following the along-slope circulation upstream of point J. In the second part (Section 5.2) we will treat the long-term behavior of these nWMDW eddies far away from the convection area, downstream from point J, once they have left the vicinity of the boundary circulation.

5.1. Spreading phase in the vicinity of the convection region

Within 100 km around the Mixed Patch, all the submesoscale eddies revealed by SOFARGOS floats moved southwestward. During this spreading phase, floats drifted parallel to the continental slope with mean advection speed of approximately 5 cm/s, comparable to the Northern Current mean speed (Figs. 9 and 10).

5.1.1. Interaction with the boundary current

Point J is the approximate location of a moored current meter, deployed at 41°20’N, 3°40’E near the continental slope by Send et al. (1996) in 1994 to observe the spreading of deep waters formed in the Medoc area. Send et al. (1996) reported a cold anomaly of 0.02°C observed in March 1994 (the year before SOFARGOS) from a current meter located at about 1450 m depth on mooring J. Send et al. (1996) interpreted the cold anomaly as nWMDW coming from the Medoc area and...
embedded in the *Northern Current*. The continental slope would have provided favorable conditions for the Coriolis force to enable dense waters to directly escape through the boundary current.

The 1D-mixed layer model used by Send et al. (1996) predicted that deep convection would have reached the bottom around February 7, 1994. Their model predicted that nWMDW was characterized by a cold temperature anomaly at all depths that year although there was no direct measurement in winter 1994 to confirm it. These estimates of nWMDW matched the cold temperature anomaly, suggesting an origin related to deep convection in 1994, in contrast with SOFARGOS observations indicating a warm temperature anomaly characterizing nWMDW at this depth.

The distance between the center of the convection region (42°N, 5°E) and mooring J being approximately 100 km, Send et al. (1996) estimated that it would have taken 20–30 days for nWMDW to reach mooring J if it was drained directly out of the Medoc area into the boundary current at the observed speed of 5 cm/s. This would also explain why the cold temperature anomaly was observed in March 1994. This anomaly had a time scale of 1 month and Send et al. (1996) extended it over a 50 km wide section based on Font (1987). According to Send et al. (1996), 50% of nWMDW would have been incorporated in the northern boundary current by the so-called “bleeding effect”.

5.1.2. Advection by submesoscale eddies in the interior of the basin

We point out that in 1995 (1 year after the Send et al., 1996 observations) SOFARGOS float data indicate a mechanism very different dynamically from the “bleeding effect” described by Send et al. (1996). Indeed, SOFARGOS floats highlighted several eddies transporting nWMDW out of the convection area and away from the northern boundary current. Several nWMDW eddies were observed at less than 50 km from the location of mooring J (Figs. 9, 10, 12 and 14). In order to compare 1994 and 1995 observations, time series of the distance of SOFARGOS floats to the point J, color-coded according to potential temperature at the positions of the floats, are shown in Fig. 11. The number of floats is not sufficient to sample the area properly, but assuming the vertical extension of the eddies to be about 1 km, centered around 1000 m

![Fig. 11. Time series of the distance of floats to point J color-coded according to in situ potential temperature in the (top) intermediate layer and (bottom) deep layer. A1, C2, A5, A3, and C3 are nWMDW eddies passing near point J.](image-url)
depth, temperatures typical of nWMDW observed by floats at intermediate levels should also be present at greater depth. In the same way, nWMDW temperatures observed by floats in the deep layer should also be present above it. This helps to identify nWMDW passing by the point J. Fig. 11a represents floats at intermediate levels (300–650 m depth) and Fig. 11b floats in the deep layers (1100–1450 m depth). At each level, one can clearly identify the nWMDW from the temperature measurements (see Section 3.2): in the deep layers, nWMDW appears as a warm temperature anomaly and above, in particular at intermediate depths where warm and salty LIW is usually encountered, nWMDW appears as a cold anomaly. This is in contrast with the Send et al. (1996) definitions of nWMDW in terms of temperature anomaly: in winter 1993/1994, the newly formed deep waters appeared to be characterized by a cold anomaly at all depths, possibly due to a colder winter.

Assuming, as did Send et al. (1996), nWMDW spreading will extend up to a 50 km wide section, we identified five eddies (A1, A5, C2, A3, and C3) passing through this 50 km wide section in a period of a month (March). These five nWMDW eddies are an underestimate of the number of eddies that could cross this section in 1 month, since we could have missed a lot of eddies because of float under-sampling. Also, anomalies typical of nWMDW are visible around 23 April, 8 May, 23 May 1995 and at the end of the experiment in July 1995 (Fig. 11), although the float trajectories did not allow us to confirm that these latter indications of nWMDW were really due to eddies.

Consequently, the period when nWMDW might spread away from the Mixed Patch could very well be at least 3 months, based on floats passing near point J and measuring temperatures typical of nWMDW, rather than the 1 month postulated by Send et al. (1996). The minimum period between the occurrence of two eddies crossing the 50 km wide section (east of point J) ranges from about 0 days (A5 and C2 crossed the area the same day) up to 10 days (between A1 and A5 or C2) and even 3 weeks (between A5 or C2 and A3). With a diameter of roughly 15 km and mean advection speeds of 5 cm/s, such nWMDW eddies could create temperature anomalies for about 3 days on a mooring record. Based on this typical eddy diameter, 3 nWMDW eddies could possibly get across the section once in 3 days, or a maximum of 30 eddies over a month. Extending these results over 3 months, up to 90 nWMDW eddies resulting from deep convection could get across this section each year, although this is an upper limit in this space and time frame since it corresponds to a domain saturated with eddies.

The amount of nWMDW formed in the winter 1994/1995 was roughly estimated to be $9 \times 10^{12} \text{ m}^3$ or 0.3 Sv averaged over the year. Considering eddy radius and vertical extension, each nWMDW eddy individually represented about 1% of the nWMDW production, which implies that up to 90% of the nWMDW production could have crossed the 50 km wide section, advected by nWMDW eddies. Only a limited number of SOFARGOS floats were launched in a sector of the Mixed Patch (about $\frac{1}{5}$ of the total area). Nevertheless, these floats revealed eight nWMDW eddies. Based on these observations, we can estimate that nWMDW eddies accounted for as much as 40% of nWMDW spreading away from the convection area, which is similar to estimations that Send et al. (1996) attributed to the “bleeding effect”.

Are advection by eddies and the “bleeding effect” two distinct and equally important mechanisms for the large scale spreading of nWMDW occurring every year? We already noted a discrepancy between these 2 years of observations: in 1995, on a section 50 km wide passing by mooring J, nWMDW would correspond to warm temperature anomalies at 1450 m depth instead of cold temperature anomalies as in 1994. Nevertheless, considering the short timescale variability (few days) on the time series produced by Send et al. (1996), these two mechanisms (that were evaluated to be of approximately equal importance) might not be so distinct but rather coexist.

5.2. Spreading phase far away from the convection region

In the first part we described the spreading phase 1 or 2 months after deep convection had occurred in the vicinity of the convection region centered in the Gulf of Lion, down to point J. We are now going to describe four distinct events occurring several months to years later and influencing the spreading of newly formed deep waters embedded in SCVs, as far away as the Algerian basin, 500 km downstream of point J. These four events are of a larger scale than the SCVs, and are related to (1) a topographic Rossby wave, (2) a deep frontal jet (the North Balearic front), (3) an eddy–eddy interaction, and (4) the Algerian Gyre.
5.2.1. Topographic Rossby waves

The behavior of the nWMDW eddies leaving the convection area was first constrained by mesoscale events characterized by waves of scales about 100 km long and 1 month period developing north of the Balearic Islands. Note the oscillation patterns indicated by black arrows along the float trajectories in Figs. 10, Fig. 12 and 14. The submesoscale eddies we could have followed until then are embedded in these waves. The motions of two nWMDW anticyclones (A3/SCV-N1 and A5/SCV-N2) are highlighted by grey arrows in Figs. 10, 12 and 14. Among these oscillating patterns, one is remarkable since it involved several floats drifting together (#04, #88 and #14, Fig. 12).

A method based on the works of Molinari and Kirwan (1975), Sanderson (1995) and Richez (1998) permits analysis of float trajectories in terms of differential kinematic parameters. A cluster of floats drifting close to each other can be seen as a water parcel. Assuming that the horizontal speed gradients are homogeneous, we can estimate those by inverting (using the least squares method) a system in which we express the velocity of each float in Taylor’s expansion about the centroid of the cluster. Once we have the speed gradients, we can compute the relative vorticity associated with the water parcel and verify a potential vorticity balance (logarithmic form) \( \frac{\partial \zeta}{\partial f} + \frac{df}{f} - \frac{dH}{H} = 0 \) where \( f = f_0 + \beta y \) is the Coriolis parameter, \( \zeta \) is the relative vorticity, and \( H \) is the thickness of the water parcel.

The computations for each term of this balance are presented in Fig. 13, \( H \) being the depth of the seafloor seen by the cluster of floats. There is a partial compensation between the \( \beta \)-effect, the topographic effect and the relative vorticity, suggesting this feature is due mainly to a barotropic topographic Rossby wave. However, the sum of each term is non-zero, and this might be due to the following factors: (1) the topographic Rossby wave observed is not necessarily purely barotropic and (2) three floats drifting at very different depths are not enough to provide precise estimates of the relative vorticity, especially because of the bias induced by any vertical shear in the calculations of vorticity.

The dispersion relation for short topographic Rossby waves for the barotropic case is

\[
\omega = \frac{-k(\beta - (f_0/H_0)(\partial H/\partial y)) - l(f_0/H_0)(\partial H/\partial x)}{k^2 + l^2},
\]

Fig. 12. Same as Fig. 9, trajectories of all RAFOS-VCM floats from April 23, 1995, until June 4, 1995, color-coded according to potential temperature.
where \((k, l)\) is the wave vector. The length scales roughly derived from Fig. 13 are \(\Delta X/2 \sim 75 \text{ km}, \Delta Y/2 \sim 50 \text{ km}\), corresponding to an elevation of the bottom topography \(\Delta H/2 \sim 100 \text{ m}\) (Fig. 13). With \(H_0 = 2600 \text{ m}, f_0 = 9.5 \times 10^{-5}, \beta = 1.7 \times 10^{-11}, \)
\(k = 2\pi \Delta X/(\Delta X^2 + \Delta Y^2), l = 2\pi \Delta Y/(\Delta X^2 + \Delta Y^2), \)
\(\partial H/\partial X = \Delta H/\Delta X\) and \(\partial H/\partial Y = \Delta H/\Delta Y\), this dispersion relation leads to \(\omega = 2.8 \times 10^{-6} \text{ s}^{-1}\) giving a time scale of approximately 53 days, which can be compared with the observed time scale of \(T/2 \sim 1\) month.

Following deep convection, there were numerous oscillations in the basin with approximately the same time and length scales as shown by floats. One oscillation, clearly observed by three floats, suggests that they might be related to topographic Rossby waves. Note that this particular oscillation differs from an ideal monochromatic wave as illustrated by the complex shape of real float trajectories. This might result from other processes including the large scale circulation, but the sampling of the float data did not allow us to characterize it more precisely.

### 5.2.2. North Balearic Front influencing SCV propagation

Noticeable is the jet-like pattern shown by floats #05, #07, #11, #90, and #03 between 350 and 1100 m depth in Fig. 14. This jet corresponds to the North Balearic Front identifiable from surface satellite imagery (Fig. 15). This front has been observed before (Lacombe and Tchernia, 1972) and is well documented by infrared imagery (Deschamps et al., 1984). In addition, the XBT survey of Fuda et al. (2000) also shows that the front has a subsurface temperature signature at the intermediate layers. However, there was no evidence of deep currents associated with this front so the deep velocity documented by the floats drifting in the area provides us with a first characterization of the flow at depth. The jet has velocities of the order of 8–10 cm/s between 350 and 1100 m depth. A nWMDW anticyclone (A5/SCV-N2) observed by float #07 was transported toward the southern basin by the jet-like current at high translation velocity (about 10 cm/s), preceded by float #05 about 60–70 km ahead and followed by floats #11, #90, and #03 at about 45–55, 70–80, and 75–80 km, respectively, drifting at about the same velocity. The length of the jet consequently reached more than 130 km.

At a later stage, this nWMDW anticyclone (A5/SCV-N2) went across the front and entered the Algerian basin (Fig. 15). This might be due to the high anticyclonic vorticity (about \(-f/2\)) of A5/SCV-N2, since eddies could cross jets that present lower horizontal shear than their vorticities according to Vandermeisch et al. (2003). Consequently, the horizontal shear of the jet should be lower than \(-f/2\). This would lead to an estimate of more than 20 km for the width of the jet. Also, the float temperature data suggest that the eddy–jet interaction might result in some water mass transformation inside the eddies, since the temperatures at float #07 (450 m) increased rapidly while crossing this front (Fig 14), suggesting some influence from warm ambient water, such as LIW.

### 5.2.3. Eddy–SCV interaction

After having crossed the North Balearic Front, A5/SCV-N2 entered the Algerian basin. In this basin, another kind of interaction with larger scale features can occur. Another Lagrangian experiment
Fig. 14. Same as Fig. 9, trajectories of all RAFOS-VCM floats from May 5, 1995, until July 1, 1995, color-coded according to potential temperature.

Fig. 15. SOFARGOS experiment, trajectories of all RAFOS-VCM floats from June 3, 1995, to June 7, 1995, and composite SST image of June 5, 1995. Note eddy A5/SCV-N2 crossing the North Balearic Front.
carried out 2 years later (MATER) also revealed nWMDW anticyclones (SCVs) in the Algerian basin. From that study, Testor and Gascard (2003) presented an interaction with a Sardinian LIW eddy (30 km radius and 3-week period, also see Testor and Gascard, 2005) west of Sardinia. This interaction, documented by floats drifting at around 600 m, was accompanied with direct exchanges of water between the two eddies.

Examples of this were exchanges between the Sardinian eddy and SCV-S1 documented by a float that passed directly from one eddy to the other. Another example was given of this float drifting at the periphery of the Sardinian eddy and showing intrusions of nWMDW (cold temperature anomaly at 600 m) in the Sardinian eddy, characterized at this level by a strong LIW signature (warm temperature anomaly at this level).

The direct exchanges of water observed between the SCV and the Sardinian eddy made them, respectively, warmer and colder in a very short time period. These could modify the waters in the eddy cores that are otherwise protected from the ambient field by the dynamical barriers presented by the relative vorticity structures of the eddies.

5.2.4. Algerian Gyre influencing SCV propagation

In the Algerian basin, the nWMDW eddies found in the MATER experiment (Testor and Gascard, 2003) follow the general cyclonic and barotropic gyre circulation, the so-called Algerian Gyre (Gascard et al., 1999; Testor and Gascard, 2005). This gyre is characterized by a diameter of around 200 km, a rotation period of about 3–4 months, and velocities at its periphery of 5–10 cm/s.

During this MATER experiment, two nWMDW SCVs were observed drifting in the cyclonic gyre for at least a year, describing several loops in the Algerian basin (Testor and Gascard, 2003). The eddies were advected with a translation velocity comparable with the mean velocity of the Algerian Gyre, demonstrating that the eddy propagation was controlled mostly by the barotropic current. Considering the time it takes for the SCVs to reach the Algerian basin, these SCVs must have been more than 1.5 year old at the end of the MATER experiment, and are expected to finally dissipate there, although there was no evidence of such dissipation in our data.

In summary, the situations presented above demonstrate that the trajectories of the submesoscale nWMDW eddies are influenced by larger scale features (gyres, topographic Rossby waves, jets, or mesoscale eddies). These features take an active part in the large-scale spreading of nWMDW, since they affect the path of the nWMDW eddies and thereby control the spreading of the newly formed deep waters.

During their long journeys, the submesoscale nWMDW eddies slowly transfer their characteristics to the surrounding waters and vice versa. Their upper part is gradually warmed by the surrounding LIW, while their lower part is cooled by the older deep waters. Homogeneous nWMDW, embedded within SCV cores, is protected from the surroundings by dynamical barriers. This allows the advection of rather “pure” nWMDW over long periods of time and far away from the source region. But events like the interaction with a Sardinian eddy (or even the North Balearic Front) can drastically modify this slow process and allow for abrupt exchanges between nWMDW and the stratified surroundings.

6. Conclusion

Thanks to the SOFARGOS and MATER experiments (1994/1995 and 1997/1998), particularly to the Lagrangian data, we have discovered long-lived submesoscale anticyclonic eddies in both the northern and southern basins of the Western Mediterranean Sea, which are by-products of deep water formation. These are subsurface eddies characterized by a high Rossby number \( \left| R_o \right| \sim 0.5 \), a small radius \( R \sim 5–10 \) km, a vertical extension of 1–2 km centered at around 1000 m depth and a large aspect ratio \( \alpha = H/R \sim 0.05–0.1 \). They have weak or no surface signature. Cyclonic submesoscale eddies with similar radius and vertical extensions were also observed, but the anticyclonic eddies seemed to be much more coherent, having typical lifetimes of 1 or more years compared to 1 (or 2) month for the cyclonic eddies. By their small size and long lifetime, they can be considered Submesoscale Coherent Vortices (SCVs). They are characterized by a homogeneous core composed of newly formed deep water (nWMDW), implying that these SCVs are somehow related to deep convection events in the Gulf of Lion.

Each SCV core corresponds to a volume of approximately 1% of the total production of nWMDW formed each year in the Northwestern Mediterranean Sea. Considering that approximately 40–50 SCVs are generated each year in the Medoc
area, 40–50% of the large scale spreading of nWMDW might be provided by these nWMDW eddies. This can be compared with the “bleeding” process of nWMDW into the Northern Current introduced by Send et al. (1996) and that would account for nearly half of the large scale spreading of the nWMDW via the general cyclonic circulation along the boundary of the Western Mediterranean Sea. Our conclusion points to a different but equally important large scale process of advection in establishing the water mass properties in the interior of the basin.

Because of their long lifetimes, SCVs can advect nWMDW several hundreds of kilometers away from the source region, hence playing an important role in the thermohaline circulation of the Western Mediterranean Sea. We demonstrated that the large- and meso-scale dynamics throughout the entire Western Mediterranean Sea constrains the path of these nWMDW eddies, and consequently the spreading of the newly formed deep waters embedded within their eddy cores.

We have shown that the large scale circulation of the region, namely the Northern Current in the Liguro-Provençal basin, the deep jet associated with the North Balearic Front, and the Algerian Gyre in the South, have strong influences on the behavior of these eddies. In addition, interactions with mesoscale features such as topographic Rossby waves (100 km, 1 month) in the northern basin, and a mesoscale LIW eddy (described by Testor and Gascard, 2003) in the Algerian basin also contributed to the fate of nWMDW eddies in the basin as a whole.

During their long journeys, the nWMDW eddies transfer their characteristics progressively to the ambient waters. Corresponding lateral fluxes of heat and momentum may be significant and occur within the intermediate and the deep layers, so that nWMDW formation influences the large scale circulation in both layers through the eddy transport process described in this study.

The exchange process is slow: nWMDW trapped near a SCV center is quite protected from the outside by dynamical barriers due to the high vorticity anomalies that characterize SCVs. However, strong interactions with mesoscale features can locally contribute to the rapid dissipation of such eddies, accelerating the release of the newly formed deep water trapped near the eddy center. Topographic Rossby waves seem to have a moderate effect on the exchanges of properties between these eddies and the ambient fluid. Eddy–eddy interactions such as the encounter between a Sardinian eddy and an nWMDW eddy appear to be very active in rapidly transferring properties between the submesoscale eddies and the mean field.

The detailed processes leading to the formation of the nWMDW eddies (SCVs) remain to be determined. The first observation of these eddies occurred in the convection area where they were subjected to baroclinic instability at the periphery of the Mixed Patch. But this observation was made about 1 month following the deep convection events leading to the formation of the new deep waters. Various aspects such as preconditioning by pre-existing eddies or intense diapycnal mixing resulting from or interacting with baroclinic instability have to be considered when dealing with aspects of SCV formation. Also, the exceptional longevity of SCVs could very well introduce a long-term “memory” effect for deep convection processes if they remain in the convection area. In this case, they will participate less efficiently in the large scale spreading of nWMDW. According to SOFARGOS observations in the Northwestern Mediterranean Sea, the nWMDW eddies seem to have a propensity to leave the convection area and contribute efficiently to the large scale spreading of newly formed deep waters resulting from deep convection. This contrasts with other deep convection sites like in the Greenland Sea, where similar long-lived eddies have been observed remaining in the convection area for a much longer time (Gascard et al., 2002). In this latter case, SCVs resulting from deep convection and deep water formation seem to be more involved during the preconditioning phase than during the spreading phase.

In a more general context, since nearly half of the newly formed deep waters could very well be dispersed in the whole Western Mediterranean Sea by submesoscale eddies interacting with other water masses, this eddy transport process should thoroughly be taken into account for the study of the thermohaline circulation of the Mediterranean Sea. Because of their small size, the nWMDW eddies cannot be explicitly resolved in large-scale high-resolution OGCMs. Hence, it would be necessary to parameterize SCVs and all the related subgrid eddy transfer processes (involving wave–eddy, eddy–jet or eddy–eddy interactions) that determine the large-scale and long-term behavior of the nWMDW eddies. To achieve that, further field experiments and numerical studies are necessary focussing on the
SCV formation and its relation to deep water formation. Furthermore, a better understanding of the long-term evolution of SCVs and the dynamical interactions between SCVs and large and mesoscale circulations is needed.

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