1 **THE SECONDARY AGEOSTROPHIC CIRCULATION IN THE IBERIAN**
2 **POLEWARD CURRENT ALONG THE CANTABRIAN SEA (BAY OF BISCAY)**

3
4
5 **Julio Gil**

6 Instituto Español de Oceanografía, Centro Oceanográfico de Santander, Promontorio de San Martín s/n. E-39004.
7 POBox 240 Santander, Spain

8 **Damià Gomis**

9 Grup d'Oceanografia Interdisciplinar, IMEDEA (UIB-CSIC). Campus UIB, 07071 Palma de Mallorca, Spain
10
11

12 **Abstract.**

13 We present a diagnostic study of the secondary ageostrophic circulation associated with a
14 meander of the Iberian Poleward Current (IPC). The study is undertaken from a hydrographic
15 survey covering a small domain of the Cantabrian Sea, along the northern Spanish coast. The
16 upper 250 m of the system consist of a wave-like eastward flow (the IPC) with maximum
17 geostrophic velocities of 15 cm/s at surface. A deeper westward counter-current (6 cm/s at 350
18 m) and an anticyclonic ring detached from the main zonal flow and extending through the
19 whole water column are the other major constituents of the system. The vertical velocity field
20 shows areas of ascent and descent motion that reach maximum absolute values of 8 m/day at
21 about 200 m. Buoyancy advection and vorticity advection contribute almost equally to the
22 observed ageostrophic vertical flow. The horizontal ageostrophic circulation (< 2 cm/s) is
23 consistent with the divergence pattern observed at upper and lower levels. Intense
24 salinity/temperature intrusions observed in different profiles are examined in the light of the
25 diagnosed ageostrophic motion.
26
27

28 **Keywords**

29 Iberian poleward current, Bay of Biscay, secondary ageostrophic circulation, mesoscale
30

1. Introduction

The arrival in the Cantabrian Sea of the warmer and saltier Iberian Poleward Current (IPC) occurring by the beginning of every winter is well documented (Ambar et al., 1986; Frouin et al., 1990; Pingree and Le Cann, 1990; Garcia-Soto et al., 2002; Gil, 2003). The IPC is detected in the form of temperature increases of about 2°C, more marked in sub-surface layers, and salinity increases of about 0.2 at surface, resulting in the intensification of horizontal gradients (Gil, 2003). After a first time stage, in which the flow advects negative density anomalies, it reaches an almost steady state with negligible density advection (Gil, 2003).

Meridional density gradients, the slope, and wind forcing interactions are the mechanisms for generating the IPC. Peliz et al. (2003), using a high-resolution primitive equation model, showed that the baroclinicity of the poleward current system can generate vorticity structures (cyclone/anticyclone eddy pairs). In fact, infrared satellite pictures often show a well-defined wave-like pattern with ridges and troughs separated about 50-100 km alongshore (see for instance Fig. 1a in Garcia-Soto et al., 2002). Some of these mesoscale meanders can detach from the zonal current (Gil, 2003) and, under abrupt topographic steering (canyons and capes), IPC waters can be injected into deeper regions, resulting in warm anticyclonic eddies (Pingree and Le Cann, 1992; Sánchez and Gil, 2004; Fernández et al., 2004). Observations reveal that eddies are mostly in geostrophic balance, with a weak secondary ageostrophic circulation along the eddy periphery (Sánchez and Gil, 2004).

Residual currents obtained from long-term moorings deployed on the Cantabrian Sea slope show that the eastward flow extends from surface down to 400-500 dbar. A weak return equatorward flow has been reported below the IPC, extending down to 900 dbar (Pingree and Le Cann, 1990; Haynes and Barton, 1990). The current-meter results of Pingree and Le Cann (1990) did not show evidences of seasonal fluctuations in the deep equator-ward flow, which suggests that this underlying current is not linked to the IPC. The most likely link is with the large-scale general circulation of the Bay of Biscay, which is weakly anticyclonic (Saunders, 1982).

While the overall structure of the IPC is fairly well documented, much less is known about the mesoscale dynamics associated with the IPC, and in particular about the vertical component of the velocity field. Many hydrographic surveys carried out in the region consist of shallow and widely spaced CTD casts (e.g. Gil, 2003), then not being suitable to carry out diagnostic studies involving the two/three differential steps required to go beyond the estimation of the geostrophic velocity (e.g., to undertake the computation of the vertical velocity and the

horizontal ageostrophic motion). Another group of studies have based on remote sensing data (e.g. Garcia-Soto et al., 2002); they have produced valuable descriptions of the surface signature of the IPC, but not of the 3D current structure. The few studies that have focused on the mesoscale dynamics of the region have concluded that it plays a key role in the biochemistry of the central Cantabrian Sea. Namely, regions of flow convergence/divergence seem to force trophic pathways and fish larvae retention, whereas vertical velocities seem to enhance the production of large-size phytoplankton (González-Quirós et al., 2004).

The interest of the mesoscale circulation associated with the IPC is therefore clear: it is poorly known and it seems to control important biogeochemical processes. The objective of this work is to shed some light on the secondary circulation, paying special attention to the vertical component of the velocity field. To do this, we use hydrographic data from an intense survey carried out in the Cantabrian Sea in February 1995, when the 1994 IPC was in its latest stage. The data set resulting from that survey is not ideal, but is up to date the best one to undertake a diagnostic study of the IPC at the mesoscale.

The structure of the paper is as follows. In section 2 we present the data set and the methodology used to estimate the secondary circulation. In section 3 we present a water mass analysis followed by the computation of the geostrophic circulation. Section 4 is devoted to the secondary circulation, presenting both the vertical and horizontal components. Results are discussed in section 5, paying particular attention to the sensitivity of the computations and to the role of the secondary circulation. Conclusions are outlined in section 6.

2. Dataset and methodology

2.1 The dataset and the spatial objective analysis

The main difficulty encountered when studying mesoscale systems is to complete a survey with an adequate spatial and temporal resolution, since there is a trade-off between the spatial density of the sampling and the synopticity of observations (Allen et al., 2002; Gomis et al., 2001). The characteristic scale of mesoscale structures such as fronts and eddies can be estimated for the region of interest from the Rossby radius ($R_d = f^{-1}(g'D)^{1/2}$). For the IPC, the density difference over its vertical scale (D) is approximately 0.4 kg/m^3 in $D=400 \text{ m}$; hence, the reduced gravity g' has a value of 0.004 m/s^2 , which altogether with the value of the Coriolis parameter f at 44°N yields a value of about 12.5 km for R_d (i.e., a wavelength of about $2\pi R_d = 75 \text{ km}$). Hence, the separation between stations should be at most equal to 37.5 km , in order to

1 have a Nyquist wavelength of the same order or smaller than the typical wavelength of
2 mesoscale structures. In fact, the station separation should preferably be significantly smaller
3 than the Nyquist wavelength unless an extensive data array is available (not our case).

4 The sampling of the February 2005 cruise consisted of 4 transects separated about 20
5 km along-shore (the direction of the dominant flow), i.e. roughly a half of the Nyquist
6 wavelength. In the cross-shore direction the 7 stations of each transect were separated about 9
7 km (see Fig. 1). The CTD casts were performed with a Sbe25 Sea-Bird profiler and reached
8 between 700 and 900 m at open sea (down to the ocean floor over the shelf and shelf edge).
9 The sampling period was 5 days of the second half of February, 1995. The quasi-geostrophic
10 (QG) advective time scale $T = L/U$ is of the order of $75 \text{ km}/0.1 \text{ m/s} = 7.5 \text{ days}$. Therefore, the
11 data collected can in principle be considered as quasi-synoptic.

12 In a first step, potential temperature and salinity profiles were used to obtain potential
13 density and specific volume anomaly profiles. Dynamic height was computed with respect to a
14 700 m reference level (see section 3.2 for more details on the selection of this depth). Station
15 values of all variables (temperature, salinity, density and dynamic height) were then
16 interpolated onto a 27×33 regular grid of $0.03^\circ\text{lon} \times 0.022^\circ\text{lat}$ ($2.4 \text{ km} \times 2.4 \text{ km}$) using an
17 Statistical Optimum Interpolation technique. We assumed a gaussian correlation model of the
18 type $C(r) = \exp\{-r^2/(2S^2)\}$, setting the characteristic scale S in such a way that the correlation
19 model fit observed correlations. The value of the best fit was obtained for $S=10 \text{ km}$; sensitivity
20 tests showed that results were not significantly influenced by the choice of this parameter
21 provided it was kept within reasonable limits (7-15 km). An additional normal-error filter
22 convolution was also applied in the way proposed by Pedder (1993), in order to filter out scales
23 which can not be resolved by the sampling. In our case, the cut-off wavelength (λ_c) was set to
24 40 km (twice the separation distance between transects and four times the separation distance
25 along transects). For a complete explanation on the interpolation process, and in particular on
26 the need to perform a scale selection see Gomis and Pedder (2005).

27 The described process was repeated (for each observed variable) at 70 horizontal levels
28 spaced 10 dbar in the vertical, then spanning the whole vertical range. All the horizontal and
29 vertical sections shown in the following have been extracted from the 3D grid resulting from
30 the overlapping of the 70 horizontal grids.

2.2 The computation of the secondary circulation

The computation of derived variables will base on the quasi-geostrophic formulation. The QG assumption is justified in the region, since taking the smallest structure retained by the analysis ($L \equiv \lambda_c/2 = 20$ km) and typical velocity values reported from currentmeter data (15-20 cm/s, see Pingree and Le Cann, 1990) the Rossby number is $Ro = U/fL \sim (0.15-0.20 \text{ m/s})/(10^{-4} \text{ s}^{-1} 2 \cdot 10^4 \text{ m}) \leq 0.1$. The computation of the geostrophic component of the velocity from gridpoint values of dynamic height is straightforward. Conversely, the computation of the vertical and the horizontal ageostrophic component of the velocity field (constituting the secondary circulation) are more cumbersome.

The vertical velocity (w) can be obtained from the QG omega equation (Holton, 1992). A form of the omega equation suitable for the analysis of mesoscale features in the ocean can be written as:

$$N^2 \nabla_h^2 w + f_o^2 \frac{\partial^2 w}{\partial z^2} = f_o \frac{\partial}{\partial z} (\mathbf{V}_g \cdot \nabla_h \xi_g) + \frac{g}{\rho_o} \nabla_h^2 (\mathbf{V}_g \cdot \nabla_h \rho) \quad (1a)$$

where N is the mean-state buoyancy frequency, ξ_g is the geostrophic relative vorticity, ∇_h is the horizontal gradient operator, and other symbols have conventional meaning. The omega equation is often used in its Q-vector form, derived by Hoskins *et al.* (1978) to avoid cancellation problems between the two terms on the right hand side of (1a):

$$N^2 \nabla_h^2 w + f_o^2 \frac{\partial^2 w}{\partial z^2} = 2 f_o \nabla_h \cdot \bar{\mathbf{Q}} \quad (1b)$$

where

$$\bar{\mathbf{Q}} = g/\rho_o [(\partial u_g/\partial x)(\partial \rho'/\partial x) + (\partial v_g/\partial x)(\partial \rho'/\partial y), (\partial u_g/\partial y)(\partial \rho'/\partial x) + (\partial v_g/\partial y)(\partial \rho'/\partial y)]$$

In our case, both equations yielded very similar results. More important are the boundary conditions that have to be assumed to solve (1a) or (1b). Usually w is set to 0 at the upper and lower boundaries of the 3D domain, which is a reasonable assumption provided the domain ranges from the ocean surface down to a rest state level. Lateral boundary conditions are usually set in a more arbitrary way, common choices being setting $w=0$ or Neumann conditions ($\partial w/\partial n=0$). Provided the horizontal scale of the structures is smaller than the size of the domain, the ellipticity of (1) makes that the interior solution for w is relatively insensitive to the imposed conditions (see for instance Gomis and Pedder, 2005). In our case we chose to set $\partial w/\partial n=0$ at the four lateral boundaries.

The secondary circulation is completed by giving the ageostrophic horizontal velocity. This can be determined in a qualitative way by taking advantage that Q vectors point in the sense of ageostrophic horizontal velocities (Hoskins *et al.*, 1978). However, the magnitude of the velocities cannot be estimated in this way. A proper estimation can be obtained from the QG formulation provided the local time derivatives of the geostrophic velocity are known. The latter can be obtained from the tendency field $\chi = \partial\phi/\partial t$, where ϕ denotes dynamic height, and for which the QG theory has also a diagnostic equation: the so-called tendency equation.

The QG tendency equation is in fact very similar to (1) from the formal point of view. It relates dynamic height and its local time derivative without explicitly determining the distribution of the vertical velocity w (Holton, 1992). It has previously been applied to ocean dynamics by Bush *et al.* (1996) and more recently by Gomis *et al.* (2005) and Flexas *et al.* (2006), who wrote the equation in a common oceanographic notation:

$$N^2 \nabla_h^2 \chi + f_o^2 \frac{\partial^2 \chi}{\partial z^2} = -N^2 f_o (\mathbf{V}_g \cdot \nabla_h \xi_g) + f_o^2 \frac{g}{\rho_o} \frac{\partial}{\partial z} (\mathbf{V}_g \cdot \nabla_h \rho) \quad (2)$$

Equation (2) can be solved in a very similar way to (1), except in the boundary conditions. The most important formal difference with respect to the omega equation is that χ cannot be set to zero at surface. From the QG density conservation equation it can be shown that a surface boundary condition consistent with setting $w=0$ is $\partial\chi/\partial z|_s = (g/\rho) \mathbf{v}_g \cdot \nabla_h \rho$. At the reference level χ must obviously be set to 0. The lateral boundary conditions can be set in the same (arbitrary) way as for the omega equation. Hence we set $\partial\chi/\partial n=0$ at the four lateral boundaries.

Once the dynamic height tendency field has been obtained, the ageostrophic velocities (u_{ag}, v_{ag}) can be subsequently computed as:

$$u_{ag} = -f^{-1} \{ f^{-1} \partial\chi/\partial x + u_g \partial v_g/\partial x + v_g \partial v_g/\partial y \} \quad (3a)$$

$$v_{ag} = -f^{-1} \{ f^{-1} \partial\chi/\partial y - u_g \partial u_g/\partial x - v_g \partial u_g/\partial y \} \quad (3b)$$

where (u_g, v_g) is the geostrophic velocity and the other symbols have conventional meaning.

3. Water mass analysis and the geostrophic circulation

3.1 Water mass analysis

With the aim of relating the flow to the different water masses present in the region, the surface dynamical topography (with respect to 700 dbar) is shown in Fig. 1. In the sampled domain the eastwards current is severely distorted by several meanders that depart the flow from the

1 isobaths. The most pronounced one is the anticyclonic ring observed to the north of the domain
2 (stations 20 and 21).

3 Figure 2 shows salinity, temperature and density at different depths. The IPC waters are
4 saltier and warmer than the surrounding waters, but the sharpest horizontal gradients of salinity
5 and temperature are observed at different depths: whereas salinity differences are largest at
6 approximately 100 dbar, the most intense temperature gradients are observed at 250 dbar (see
7 Figs. 2a, e). At upper levels (down to 100 m) the contribution of salinity gradients separating
8 the IPC from open-sea waters is mostly cancelled out by the relatively smooth thermal
9 gradients separating both water masses (Figs. 2a, b). The consequence is that upper-layer IPC
10 waters are not distinguished from open-sea waters in the density field (Fig. 2c). A different
11 feature is the salinity and density minima located over the shelf (Fig. 2a, c); this structure is due
12 to the mixing of warm IPC waters with fresh continental waters, resulting in low densities
13 confined near the coast.

14 Below 150 m the intense temperature gradients are not compensated by the very weak
15 salinity differences: at 250 dbar, for instance, the salinity of st. 7 (offshore) and st. 18 (within
16 the IPC) are quite similar (Fig. 2d), whereas the temperature is 1.6°C higher at st. 18 than at st.
17 7 (Fig. 2e). This results in intense density gradients separating IPC waters from open-sea waters
18 (Fig. 2f). The baroclinic pattern of the IPC waters causes an intense weakening of the surface
19 IPC flow, to the point that a zero velocity could be reached in between 250 and 300 m depth
20 (see section 3.2). Below 450 dbar, the density slopes reverse with respect to upper layers (see
21 stations 1-4 at figs. 2f and 2i).

22 Apart from the general pattern described above, some temperature and salinity profiles
23 show sharp vertical gradients at upper levels (between 100 and 200 dbar) that suggest the
24 occurrence of water intrusions. In some cases, they look like intrusions of fresher and colder
25 water (see Fig. 3a for instance) and in others the intruding water is saltier and warmer, coming
26 from the poleward current (Fig. 3b). Some of the observed profile anomalies can hardly be
27 explained from the IPC geostrophic current, and suggest the presence of
28 divergence/convergence processes related to ageostrophic transport. These intrusions will
29 therefore be discussed later on in this work, once the ageostrophic circulation has been
30 determined.

3.2 *The geostrophic velocity field.*

A key issue was to determine a proper reference or no-motion level in absence of ADCP data. Residual currents obtained from long-term moorings deployed on the Cantabrian Sea slope show that the eastward IPC flow extends from surface down to 400-500 dbar, and that the returning equatorward flow extends down to 900 dbar (Pingree and Le Cann, 1990; Haynes and Barton, 1990). Other authors (e.g. Frouin et al., 1990) have reported in situ data suggesting that the lower boundary of the IPC can be much shallower, between 150 and 200 dbar. Therefore, two possibilities were contemplated to determine the reference level: i) to use the interface between the IPC and the reported deep countercurrent as no-motion level; ii) to use the deepest level available (700 m over the whole domain) as no-motion level.

Choosing the first option is cumbersome: a visual inspection of the isopycnals along each transect reveals that they are more or less horizontal in between 400 and 450 dbar (not shown); below that depth, the density slopes reverse with respect to upper layers, but not over the whole domain. This implies that the no-motion level should be above these levels, in order to obtain a westwards current at lower levels. We carried out several tests setting the reference level at different depths in between 300 and 450 m, but none of them yielded a realistic flow structure both above and below the reference level. The reason is probably that the velocity does not vanish at a constant depth all over the domain.

Setting the reference level to 700 dbar over the whole domain resulted in a more realistic flow structure. Although it is true that the velocity probably does not completely vanish at that level, the fact that velocity values are much smaller than at upper levels makes that the impact of not exactly fulfilling that assumption is smaller than when assuming no motion at an upper level. Under this assumption, the IPC is shown to be restricted to the upper 300 m and the countercurrent reaches maximum values between 300 and 400 m. Both features hold when choosing 900 m as reference level, though the values and vertical extension of the undercurrent change.

In order to give an overall view of the geostrophic velocity associated with the system described above, we present two horizontal sections that are representative of the flows dominating the upper and lower layers: the eastward IPC in the upper layer (10 dbar, Fig. 4a) and the westward underlying countercurrent (300 dbar, Fig. 4b). The current vectors have been drawn on the dynamic topography of the respective levels, in order to highlight the anticyclonic/cyclonic areas.

Figure 4 is complemented by Fig. 5, which shows the component of the geostrophic velocity across vertical sections oriented along the four station transects. The IPC is apparent in the upper 200 m, though the different location occupied in the panels reveals the marked meandering of the current within the domain. The speed of the IPC core across the sections of Fig. 5 (i.e., the zonal component) ranges between 10 and 15 cm/s, the highest values being always obtained at surface. However, over the entire domain, maximum speeds correspond to the meandering of the current around the anticyclonic eddy, i.e., where the current is not zonal, but practically meridional (see the geostrophic current at 100 dbar, Fig. 6). Around the eddy, maximum geostrophic speeds range between 15 and 20 cm/s and keep almost constant from surface to 100 dbar.

Figure 5 also shows that the transition between the IPC and the westwards countercurrent would be located between 200 and 300 dbar (i.e., closer to the depth reported by Frouin et al., 1990, than to the depth reported by Pingree and Le Cann, 1990). However, it must be recalled that geostrophic computations are based on assuming a no-motion level at 700 dbar, an assumption that clearly introduces some uncertainty as recognized when discussing the location of the reference level. The core of the westwards countercurrent is located at about 300-400 dbar, and the location over the domain roughly follows the terrain: at the westernmost transect, where isobaths deflect to the southwest, the countercurrent also turns southwestwards (Fig. 4b). The zonal speed of the undercurrent core is maximum at the central transects (Figs. 5b, c), in part because there the velocity is practically zonal, but also because of a reinforcement of the current due to the presence of the eddy. Maximum values are of the order of 6 cm/s.

4. The secondary circulation

4.1 The horizontal ageostrophic circulation

Figure 7 shows the horizontal ageostrophic velocity at 100 and 300 dbar. Typical values are of the order of 2 cm/s at 100 dbar and much weaker at lower levels. A first worth noting feature is that at 100 dbar the ageostrophic circulation is mostly clockwise around the anticyclonic eddy located to the north of the domain and also around the marked cyclonic meander C1. This is a consequence of the speed underestimation/overestimation inherent to the geostrophic balance in presence of non-negligible centripetal accelerations associated with marked anticyclonic/cyclonic curvature. This contribution has been shown to be mostly non-divergent (Gomis et al, 2001), so that it does not have a necessary impact on the forcing of vertical velocities.

A second worth noting feature is the presence of apparent divergent/convergent velocity patterns in some regions of the domain. At 100 dbar (Fig. 7a) there is a clear divergent pattern located at about 43.85°lat , -3.7°lon , and a convergent pattern located at about 43.90°lat , -3.4°lon (the latter is less clear from the vector field, but shows up when computing the divergence). A similar structure is obtained at all levels above 200 dbar. Conversely, at lower levels (see for instance at 300 dbar, Fig. 7b), the divergence/convergence regions have opposite sign to those observed at upper levels: although velocities are much smaller, some convergence is apparent at about 43.85°lat , -3.7°lon , while there is some divergence at about 43.90°lat , -3.4°lon . These structures are consistent with the location of the maximum values of vertical velocity described in the next section.

4.2 The vertical velocity field

As stated in section 2.2, the results obtained for the vertical velocity are very robust in front of the used formulation (the omega equation and the Q-vector formulation). Maximum absolute values are obtained at about 200 dbar (Fig. 7c), reaching $+8$ m/day and -8 m/day. They are approximately collocated with the convergence/divergence centres observed in the horizontal ageostrophic velocity (Figs. 7a,b and 8). Another nucleus of descending motion (values of up to -7 m/day) is located at about 43.75°lat , -3.9°lon .

Combining the vertical velocity with the horizontal ageostrophic velocity we obtain the secondary circulation. The meridional section along transect 2, which crosses areas of ascent and descent motion, shows the clear signature of a secondary circulation cell (Fig. 8b). It is worth noting that the ageostrophic velocity is mostly meridional along this section, so that the vectors represented in Fig. 8b practically account for the whole ageostrophic circulation. This Figure also illustrates how the main area of ascent motion referred on Fig. 7c results in the upper-level horizontal divergence observed in Fig. 7a (at 43.85°lat , -3.7°lon).

Figures 8a and 8c show the same fields as Fig. 8b but for the westernmost transect and a section crossing the convergence area close to the eastern boundary, respectively. In the westernmost transect (Fig. 8a) there is a clear downwelling region at stations 3 and 4 (43.8°lat), whereas Fig. 8c illustrates how the upper-level horizontal convergence observed in Fig. 7a results in the descent motion nucleus shown in Fig. 7c (at 43.90°lat , -3.4°lon). For these two transects the ageostrophic velocity is also predominantly meridional, though it is more undefined and rather weak in the case of Fig. 8a. All vertical sections show that ageostrophic velocities extend almost down to 600 dbar, though they are more important in the upper 300

1 dbar. Actually, the horizontal component is mostly confined to the upper 250 dbar, while the
 2 vertical component reaches a maximum at 200 dbar and extends with significant values down
 3 to more than 500 dbar.

4 To investigate the mechanisms responsible for the observed vertical velocities we can
 5 examine the two terms of the vertical forcing conforming the rhs of the omega equation (1a),
 6 namely the vertically differential advection of geostrophic vorticity and the laplacian of
 7 buoyancy advection (Figs. 9a, b). In the two main regions of ascent/descent motion, the two
 8 terms have the same sign. The lack of cancellation between the two terms explains why the
 9 results of the omega eq. (1a) and of the Q-vector formulation (1b) give identical results (the rhs
 10 of the omega equation was re-written in terms of the Q-vector precisely to avoid accuracy
 11 problems derived from the approximate cancellation between the two terms; if the two terms of
 12 the rhs of the omega equation do not suffer from cancellation problems, the two formulations
 13 are practically equivalent). In fact, the values of the two terms are practically identical in the
 14 central part of the domain (yielding the upward motion region), whereas the buoyancy
 15 advection term has a some more intense pattern to the east (at one of the downward motion
 16 region) and the vorticity advection has a some more intense pattern to the west (at the other
 17 major downward motion region).

18 The dynamical forcing of the vertical motion can also be illustrated plotting the vorticity
 19 and density fields against the geostrophic flow (Figs. 9c, d). In the region of ascent motion
 20 there is a clear advection of coastal light waters (i.e., $\mathbf{V}_g \cdot \nabla_h \rho > 0$) towards the western
 21 boundary of the anticyclonic eddy; conversely, in the region of descent motion (in particular in
 22 the eastern one) there is a clear advection of open-sea dense water (i.e., $\mathbf{V}_g \cdot \nabla_h \rho < 0$) towards
 23 the zonal current (Fig. 9d). This results in the negative/positive values for the laplacian of
 24 buoyancy advection reported in Fig. 9b (the Laplacian of a wave-like structure has opposite
 25 sign to the structure itself).

26 Figure 9c shows that the central area of ascent motion is also linked to the advection of
 27 positive vorticity ($\mathbf{V}_g \cdot \nabla_h \xi_g < 0$) from the cyclonic nucleus located to the SW of the eddy
 28 towards the western boundary of the anticyclonic ring. Conversely, in the two regions of
 29 descent motion negative vorticity is advected (i.e., $\mathbf{V}_g \cdot \nabla_h \xi_g > 0$); in the eastern one negative
 30 vorticity from the eastern side of the anticyclonic ring is pushed into the zonal current, whereas
 31 in the western one the negative vorticity comes from the continental waters confined near the

coast (Fig. 9c). The vertical derivative of vorticity advection keeps the sign of the vorticity advection itself, then resulting in the negative/positive values reported in Fig. 9a.

5. Discussion

5.1 On the reliability of the computations

One of the main concerns of mesoscale diagnostic studies is the accuracy of the derived fields. The secondary circulation is obtained after the integration of differential equations that contain non-linear combinations of terms consisting of up to third-order derivatives of observed variables. Hence, the accuracy of the gridded observed fields is essential. Statistical optimal interpolation provides an estimation of the errors associated with the observations and the interpolation process, under the assumption that the parameters of the method (basically the correlation scale and the noise-to-signal ratio) are correct. This is never exactly the case, due to the hypothesis of isotropic correlation, for instance, but at least the method provides an approximation to actual errors. Following Gomis and Pedder (2005) we have computed not only the total error field, but also its partition in two contributions: a first one due to the presence of errors in the observations and a second one due to the discrete sampling. A sample of the obtained error fields is shown in Fig. 10, expressed as a fraction between the error standard deviation and the standard deviation of the field. Namely, we show the distribution for dynamic height (the key field regarding dynamical computations) at 200 dbar (the level at which the vertical velocity is more intense).

As it is usually the case for hydrographic data measured with a CTD, the contribution of observational errors is small over most of the domain (except at the boundaries): less than 0.05 times the standard deviation of the field (Fig. 10a). The contribution due to the sampling is higher and reflects the station distribution: it is less than 0.10 times the standard deviation of the field close to the transects, but reaches up to 0.14 in between transects (Fig. 10b). The sum of the two contributions (in terms of variance, not of standard deviation) gives the total error field (Fig. 10c). All figures shown previously in the paper have been restricted to the regions where total errors are less than 0.15 times the standard deviation of the field.

It is well known that errors are larger for the derived variables than for the observed variables. Gomis and Pedder (2005) showed that even having dynamic height errors smaller than 0.10 times the standard deviation of the field could result in vertical velocity errors of up to 0.30 . These numbers are common to most hydrographic surveys and can only be reduced

1 through a dense sampling strategy (e.g. using a SeaSoar) or including accurate ADCP data in
2 the interpolation process.

3 More worrying is the eventual contribution of the lack of synopticity of observations.
4 Gomis et al. (2005) showed that in presence of rapidly evolving fields, this contribution can be
5 significantly higher than the ones due to observational errors or to the station distribution. The
6 problem is that the impact of the lack of synopticity can only be estimated if the time evolution
7 of the system is known. Some qualitative insight can be gained from the values obtained for the
8 dynamic height tendency. A sample of that field (at 200 dbar) is shown in Fig. 11: tendency
9 values are smaller than 0.08 dyn cm/day, which indicates that for two adjacent transects
10 (separated by a maximum sampling time of two days), the dynamic height of the first one could
11 have changed up to 0.15 dyn cm by the time the ship was sampling the second one. The
12 dynamic height difference between transects is of the order of 1 dyn cm at that depth, so that
13 the errors associated with the lack of synopticity would be of about 15% of the sampled
14 gradients. Although this is not a negligible amount, it is not higher than for most oceanographic
15 surveys carried out using a CTD probe.

16 ***5.2 On the role of the secondary circulation***

17 Detecting any impact of the secondary circulation on the water mass pattern is usually difficult
18 due to the dominance of the geostrophic circulation. However, the presence of salinity and
19 temperature gradients such as those separating the IPC from the outer waters make that some of
20 these impacts might be identified. In particular, the presence of apparent intrusions in several
21 station profiles may be due to advection processes linked to the secondary circulation
22 (altogether with the geostrophic advection by the main current).

23 As an example we will focus on the intermediate transect 2 (st. 8-14). In the upper 200
24 dbar, the geostrophic current flows almost northwards along this transect (Fig. 6), though with
25 some cyclonic curvature surrounding station 11 and some anticyclonic curvature surrounding
26 station 13. Strong similarities should therefore be expected between the upper layers of stations
27 11, 12 and 13. When looking at the profiles, however, station 12 looks rather different from the
28 others between 100 and 200 m (Fig. 12). A possible explanation comes from the ageostrophic
29 circulation along this transect shown in Fig. 8b: at a latitude of about 43.85° (where station 12
30 is located) there is a clear upwelling below 200 dbar. Hence, fresher and colder waters coming
31 from lower levels would reach the 100-200 dbar layer at station 12, whereas these waters
32 cannot reach such upper levels at stations 11 and 13. The layers of upwelled waters have been
33 marked with a thick line in Fig. 12.

5.4 On the role of the deep undercurrent.

From Figs. 1 and 4b it can be seen how both the cyclonic and anti-cyclonic areas (labelled C1-C2 and A) are bordered by the meandering IPC and also by the undercurrent at deeper layers. The region between C1 and A has been shown to account for maximum density and vorticity advection. In this region the ageostrophic circulation at 100 dbar (Fig. 7a) is mostly clockwise, whereas at 300 dbar (Fig. 7b) it is anticlockwise around C1. Our hypothesis is that the undercurrent may be the cause of instabilities in the entering IPC flow and the origin of the western branch.

Fig. 3a shows the temperature and salinity of the st. 03 along transect 1. Between 200 and 250 dbar st. 03 has low salinities and relatively cold waters (colder than IPC waters and warmer than those far from them). This pattern can hardly be explained from the IPC geostrophic current. The only waters with similar temperature-salinity at these levels are located around the anticyclonic ring (st.12, see Fig. 12), and therefore these waters could have reached st. 03 due to the horizontal advection (from st.12 to st.03) caused by the undercurrent (see Fig. 4b).

6. Conclusions.

In this study we have determined the geostrophic and ageostrophic circulation associated with the IPC current in a small region of the Cantabrian Sea. The system consists of a wave-like upper-layer eastward flow (the IPC), a deeper westward counter-current and an anticyclonic ring detached from the main zonal flow. Both the error analysis provided by the Optimal Statistical Interpolation method and the estimates on the synopticity of the sampling indicate that the data set is adequate to undertake a dynamical diagnosis at the mesoscale.

Regarding the geostrophic circulation, we have obtained an IPC that extends down to 250 m. Compared to previous results, it is shallower than the 400 m reported by Pingree and Le Cann (1990) or Haynes and Barton (1990) and deeper than the 150-200 m reported by Frouin et al (1990). After several tests changing the reference level we concluded that the obtained flow structure was robust in front of reasonable changes in the reference level.

Regarding the secondary circulation, we have determined the pattern of both the vertical and horizontal components. Maximum absolute values of 8 m/day have been obtained for the vertical motion. They are equally associated with buoyancy and vorticity advections taking place mostly in the region between a small cyclonic meander and the anticyclonic ring

1 detached from the main flow. The horizontal ageostrophic velocity reaches maximum values of
2 2 cm/s and shows a convergence/divergence pattern that is fully consistent with the vertical
3 motion diagnosed at different levels.

4 The robustness of the diagnosed ageostrophic fields has been indirectly proved by
5 their consistency with the anomalous features observed in temperature and salinity profiles at
6 stations where maximum ascent motion has been diagnosed.

7

8

References

- Allen, J. T., D. A. Smeed, A. J. G. Nurser, J. W. Zhang, M. Rixen, 2002. Diagnosis of vertical velocities with the QG omega equation: an examination of errors due to sampling strategy. *Deep-Sea Res. I*, **48**, 315-346.
- Ambar, I. J., A. F. G. Fiúza, T. Boyd, R. Frouin, 1986. Observations of a warm oceanic current flowing northwards along the coasts of Portugal and Spain during Nov-Dec 1983. *EOS Transaction. American Geophysical Union*, **67** (144), 1054 pp.
- Bush, A. B. G., J. C. McWilliams, W. R. Peltier, 1996. The formation of oceanic eddies in symmetric and asymmetric jets. Part II: Late time evolution and coherent vortex formation. *J. Phys. Ocean.*, **26** (9), 1825–1848.
- Fernández, E., F. Alvarez-Marques, R. Anadón, S. Barquero, A. Bode, A. García, C. Garcia-Soto, J. Gil, N. González, A. Iriarte, B. Mouriño, F. Rodríguez, R. Sánchez, E. Teira, S. Torres, L. Valdés, M. Varela, R. Varela, M. Zapata, 2004. An Overview of the Physical, Chemical and Ecological Properties of a Slope Water Oceanic Anticyclonic edDY (SWODDY) in the Southern Bay of Biscay. *J.Mar. Biol. Ass. U.K.*, **84**, 501-517.
- Flexas, M. M., D. Gomis, S. Ruiz, A. Pascual, P. León, 2006. In situ and satellite observations of the eastward migration of the Western Alboran Sea Gyre. *Prog. Ocean.*, **70**, 486–509.
- Frouin, R., A. F. G. Fiúza, I. Ambar, T. J. Boyd, 1990. Observations of a Poleward Surface Current off the Coasts of Portugal and Spain during Winter. *J. Geophys. Res.*, **5** (C1), 679-691.
- García-Soto, C., R. D. Pingree, L. Valdés, 2002. Navidad development in the southern Bay of Biscay: Climate change and swoddy structure from remote sensing and in situ measurements. *J. Geophys. Res.*, **107** (C8), doi: 10.1029/2001JC001012.
- Gil, J., 2003. Changes in the pattern of water masses resulting from a poleward slope current in the Cantabrian Sea (Bay of Biscay). *Estuarine, Coastal and Shelf Science*, **57**, 1139-1149.
- Gomis, D., S. Ruiz, M.A. Pedder, 2001. Diagnostic analysis of the 3D ageostrophic circulation from a multivariate spatial interpolation of CTD and ADCP data. *Deep-Sea Res. I*, **48**, 269-295.

- 1 Gomis, D., M. A. Pedder, 2005. Errors in dynamical fields inferred from synoptic
2 oceanographic cruise data. Part I: the impact of observation errors and the sampling
3 distribution. *J. Mar. Sys.*, **56** (3–4), 317–333.
- 4 Gomis, D., A. Pascual, M. A. Pedder, 2005. Errors in dynamical fields inferred from synoptic
5 oceanographic cruise data. Part II: the impact of the lack of synopticity. *J. Mar. Sys.*,
6 **56** (3–4), 334–351.
- 7 González-Quirós, R., A. Pascual, D. Gomis, R. Anadón, 2004. Influence of mesoscale physical
8 forcing on trophic pathways and fish larvae retention in the central Cantabrian Sea.
9 *Fish. Oceanog.* **13**:5, 1-14.
- 10 Haynes, R., E. D. Barton, 1990. A poleward flow along the Atlantic coast of the Iberian
11 peninsula. *J. Geophys. Res.*, **95**: 11425-11141.
- 12 Holton, J. R., 1992. *An Introduction to Dynamic Meteorology*. Academic Press, 511 pp.
- 13 Hoskins, B. J., I. Draghici, H. C. Davies, 1978. A new look at the Omega equation. *Quart. J.*
14 *Roy. Meteor. Soc.*, **104**, 31-38.
- 15 Pedder, M. A., 1993. Interpolation and filtering of spatial observations using successive
16 corrections and gaussian filters. *Mon. Wea. Rev.*, **117**, 1675-1708.
- 17 Peliz, A., J. Dubert, D. B. Haidvogel, B. Le Cann, 2003. Generation and unstable evolution of a
18 density-driven Eastern Poleward Current: The Iberian Poleward Current. *J. Geophys.*
19 *Res.*, **108**(C8), 3268, doi: 10.1029/2002JC001443.
- 20 Pingree, R. D., B. Le Cann, 1990. Structure, strength and seasonality of the slope currents in
21 the Bay of Biscay region. *J. Mar. Biol. Ass. U.K.*, **70**, 857-885.
- 22 Pingree, R. D., B. Le Cann, 1992. Three anticyclonic Slope Water Oceanic eDDIES
23 (SWODDIES) in the southern Bay of Biscay in 1990. *Deep-Sea Research*, **39**, 1147-
24 1175.
- 25 Sánchez, R., J. Gil, 2004. 3D structure, mesoscale interactions and potential vorticity
26 conservation in a swoddy in the Bay of Biscay. *J. Mar. Sys.*, **46**, 47-68.
- 27 Saunders, P. M., 1982. Circulation in the eastern north Atlantic. *J. Mar. Res.*, **40**: 641-657.

Figure Captions.

Figure 1: Geographical context of the work, with the 28 CTD stations used to obtain all results. The eastwards IPC meandering current is well apparent in the 10 m dynamic height field, in dyn cm ($1 \text{ dyn cm} = 10^{-3} \text{ m}^2/\text{s}^2$) referred to 700 dbar. Isobaths (dashed lines) are superimposed.

Figure 2. salinity, potential temperature ($^{\circ}\text{C}$) and σ_{θ} (kg/m^3) at 100, 250 and 600 m.

Figure 3. Salinity and potential temperature profiles showing some relevant water intrusions: (a) at stations 03 and 24; (b) at stations 14 and 20.

Figure 4. Geostrophic current at 10 dbar and 300 dbar. Vector currents have been plotted on the dynamic topographic of the respective level in order to highlight the cyclonic/anticyclonic areas.

Figure 5. zonal geostrophic speed (cm/s) across the four meridional transects: a) the westernmost one (st. 1-7); b) the second transect (st. 8-14); c) the third transect, running across the eddy (st. 16-21); d) the easternmost transect (st. 22-28)..Positive values denote eastward velocities.

Figure 6. Geostrophic velocity at 100 dbar. Stations of transect 2 have been superimposed.

Figure 7. Horizontal ageostrophic velocity at 100 dbar (a) and 300 dbar (b); vertical velocity (in m/day) at 200 dbar (c).

Figure 8. Ageostrophic velocity on three vertical S-N sections; a) along transect 1 (st. 1-7); b) along transect 2 (st. 8-14); across the region of maximum divergence ($\approx 3.4^{\circ}\text{lon}$ meridian, close to transect 3). Note that the depth is in Dbar (aprox. tens of m). The sample vector of the left upper corner corresponds to 1 cm/s in the horizontal and to 1 m/day in the vertical. The background field is density.

Figure 9. distribution at 200 dbar of the two terms conforming the vertical forcing (in $10^{-17} \text{ m}^{-1}\text{s}^{-3}$): the vertical derivative of geostrophic vorticity advection (a) and the laplacian of density advection (b). The next two panels show the geostrophic velocity superimposed onto the geostrophic relative vorticity (in 10^{-5} s^{-1}) (c) and onto density (d).

Figure 10. Error distribution for the gridded values of dynamic height at 200 dbar, as given by the optimal interpolation method. They are expressed as a fraction between the error standard deviation and the standard deviation of the field, the spacing between isolines being 0.05. The

1 panels show the errors derived from errors in the observations (a) those derived from the
2 discrete station distribution (b) and total errors (c).

3 **Figure 11.** Tendency of dynamic height (in dyn.cm/day) at 200 m and the dynamic height field
4 (dyn cm) at the same depth.

5 **Figure 12.** Salinity and potential temperature profiles at three stations located on the second
6 transect (-3.76° lon): st. 11, 12 and 13.

7

8

Figure 1: Geographical context of the work, with the 28 CTD stations used to obtain all results. The eastwards IPC meandering current is well apparent in the 10 m dynamic height field, in dyn cm ($1 \text{ dyn cm} = 10^{-3} \text{ m}^2/\text{s}^2$) referred to 700 dbar. Isobaths (dashed lines) are superimposed.

Figure 2: salinity, potential temperature ($^{\circ}\text{C}$) and σ_{θ} (kg/m^3) at 100, 250 and 600 m.

Figure 3. Salinity and potential temperature profiles showing some relevant water intrusions: (a) at stations 03 and 24; (b) at stations 14 and 20.

Figure 4: geostrophic current at 10 dbar and 300 dbar. Vector currents have been plotted on the dynamic topographic of the respective level in order to highlight the cyclonic/anticyclonic areas.

Figure 5: zonal geostrophic speed (cm/s) across the four meridional transects: a) the westernmost one (st. 1-7); b) the second transect (st. 8-14); c) the third transect, running across the eddy (st. 16-21); d) the easternmost transect (st. 22-28). Positive values denote eastward velocities.

Figure 6: geostrophic velocity at 100 dbar. Stations of transect 2 have been superimposed.

Figure 7: horizontal ageostrophic velocity at 100 dbar (a) and 300 dbar (b); vertical velocity (in m/day) at 200 dbar (c).

Figure 8: ageostrophic velocity on three vertical S-N sections: a) along transect 1 (st. 1-7); b) along transect 2 (st. 8-14); c) across the region of maximum divergence ($\square 3.4^\circ$ lon meridian, close to transect 3). Note that the depth is in Dbar (aprox. tens of m). The sample vector of the left upper corner corresponds to 1 cm/s in the horizontal and to 1 m/day in the vertical. The background field is density.

Figure 9: distribution at 200 dbar of the two terms conforming the vertical forcing (in $10^{-17} \text{ m}^{-1} \text{ s}^{-3}$): the vertical derivative of geostrophic vorticity advection (a) and the laplacian of density advection (b). The next two panels show the geostrophic velocity superimposed onto the geostrophic relative vorticity (in 10^{-5} s^{-1}) (c) and onto density (d).

1

Error distribution for the gridded values of dynamic height at 200 dbar, as given by the optimal interpolation method. They are expressed as a fraction between the error standard deviation and the standard deviation of the field, the spacing between isolines being 0.05. The panels show the errors derived from errors in the observations (a) those derived from the discrete station distribution (b) and total errors (c).

2

3

4

5

6

7

8

9

10

11

12

Figure 11: Tendency of dynamic height (dyn cm/day) at 200 m and the dynamic height field (dyn cm) at the same depth.

Figure 12: salinity and potential temperature profiles at three stations located on the second transect (-3.76° lon): st. 11, 12 and 13.