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Monitoring of a quasi-stationary eddy in the Bay of Biscay by means of satellite, *in situ* and model results

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ABSTRACT

The presence of a quasi-stationary anticyclonic eddy within the southeastern Bay of Biscay (centred around 44°30′N–4°W) has been reported on various occasions in the bibliography. The analysis made in this study for the period 2003–2010, by using *in situ* and remote sensing measurements and model results shows that this mesoscale coherent structure is present almost every year from the end of winter-beginning of spring, to the beginning of fall. During this period it remains in an area limited to the east by the Landes Plateau, to the west by Le Danois Bank and Torrelavega canyon and to the northwest by the Jovellanos seamount. All the observations and analysis made in this contribution, suggest that this structure is generated between Capbreton and Torrelavega canyons. Detailed monitoring from *in situ* and remote sensing data of an anticyclonic quasi-stationary eddy, in 2008, shows the origin of this structure from a warm water current located around 43°42′N–3°30′W in mid-January. This coherent structure is monitored until August around the same area, where it has a marked influence on the Sea Level Anomaly, Sea Surface Temperature and surface Chlorophyll-a concentration. An eddy tracking method, applied to the outputs of a numerical model, shows that the model is able to reproduce this type of eddy, with similar 2D characteristics and lifetimes to that suggested by the observations and previous works. This is the case, for instance, of the simulated MAY04 eddy, which was generated in May 2004 around Torrelavega canyon and remained quasi-stationary in the area for 4 months. The diameter of this eddy ranged from 40 to 60 km, its azimuthal velocity was less than 20 cm s^{-1}, its vertical extension reached 3000–3500 m depth during April and May and it was observed to interact with other coherent structures.

1. Introduction

Several studies have shown observational evidence of westward propagation of mesoscale eddies in the open ocean. In the subtropical eastern North Atlantic a subtropical ring called STORM that moved westward near Azores in 1995, was analysed by means of current metre moorings, ARGOS buoys and subsurface floats (Pingree et al., 1996). This mesoscale structure was subsequently studied by Pingree and Sinha (1998) combining, in this occasion, *in situ* data with altimetry. Based also on *in situ* and remote sensing (altimeter/SeaWiFS) measurements, a westward propagating eddy at 26°N (generated south of the Canary Islands) and called SWESTY was analysed and followed for > 500 days in 1993 and 1994, by an ARGOS buoy (Pingree, 1996; Pingree and Garcia-Soto, 2004). More recent studies using satellite altimetry also show observational evidence of the westward propagation of eddies (*e.g.* Morrow et al., 2004; Chelton et al., 2007; Sangrà et al., 2009).

The westward propagation of mesoscale eddies in a β-plane occurs, regardless of their polarisation, due to the variation of the Coriolis parameter with latitude (Cushman-Roisin and Beckers, 2011). Also consistent with theory, global observations show small opposing meridional poleward and equatorward deflections of cyclones and anticyclones, respectively (Chelton et al., 2007, 2011). However, the translational behaviour of mesoscale eddies has been observed to be more complicated when other processes are added to the potential vorticity balance in a β-plane, due to (i) the presence of a sloping bottom (topographic β-effect), which as the Coriolis gradient with latitude, causes a translation of eddies with the shallower region to its right in the northern hemisphere

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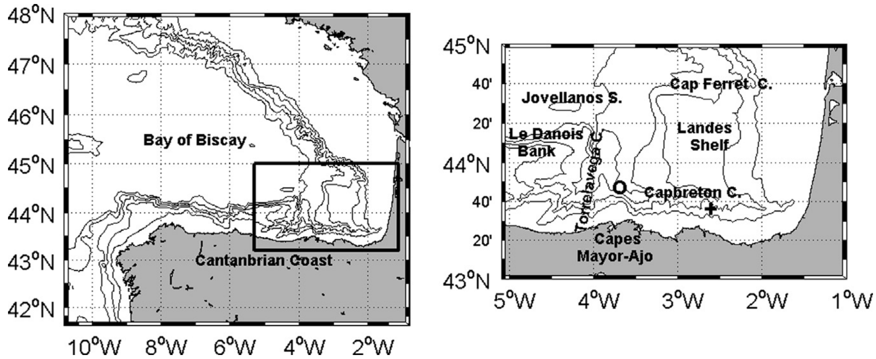

Fig. 1. (Left) Location of the area of study, within the Bay of Biscay. The solid line represents the limits of the grid in the model simulation. (Right) Study area and the main topographic structures; the location of the AGL buoy (2850 m) is indicated with a circle, whilst the location of the Matxitxako buoy (550 m) is indicated with a cross. Isobaths (m): 200, 1000, 2000, 3000 and 4000.

(Nof, 1983; Cushman-Roisin and Beckers, 2011); (ii) the interaction with bathymetric irregularities (*e.g.* submarine seamounts, canyons); (iii) eddy–eddy interactions; or (iv) interactions of eddies with existing currents. Furthermore, on some occasions these processes can become an obstacle to the eddy translation. In these cases, eddies can remain trapped near the same position during part of their lifetimes. Examples of these so-called quasi-stationary eddies have been observed in different locations of the global ocean. In the northeastern Indian Ocean, a stationary counterclockwise eddy was observed, between the Cuvier basin and the Exmouth Plateau (Metso et al., 1986; Quadfasel et al., 1996). A similar situation occurred in the northern North Atlantic, where the North Atlantic Current formed quasi-stationary meanders (Shoosmith et al., 2005). In that study, looping RAFOS floats monitored a number of quasi-stationary eddies in different places. As an example, to the west of the Mid Atlantic ridge, an anticyclonic eddy remained over a slightly elevated sea-floor area for 9 months, from October 1997. In the California Current System, a ~500 m high nearly-cylindrically-symmetrical seamount and other topographic disturbances, may either generate and/or trap some offshore eddies by means of barotropic effects (Bernstein et al., 1977; Burkov and Pavlova, 1980; Simpson, 1982; Simpson et al., 1984). Above the topographic anomaly, Zapiola Rise (45°S and 45°W), an intense anticyclonic gyre of barotropic circulation has been observed and modelled in several studies (*e.g.* Saunders and King, 1995; Volkov and Fu, 2008). The Zapiola anticyclone is maintained by eddy–topography interactions and controlled by the bottom friction (De Miranda et al., 1999). Finally, these are some examples of quasi-stationary eddies in the eastern North Atlantic: a dipole partially trapped to East Thulean Rise and monitored during more than half a year by an Argos buoy in 1990 (Pingree and Le Cann, 1991); a MEDDY trapped for 11 months in 1997 northeast of the Charcot seamount (45°N, 11°30′W) (Paillet et al., 2002); A NACW anticyclonic eddy monitored from 2000 to 2001, which described a small total displacement, in the northeast Atlantic (~43°30′N, 15°–19°W) (Reverdin et al., 2009).

Within the southeastern Bay of Biscay (Fig. 1), quasi-stationary eddies have been reported for different years near the same position, around 44°N–4°W (Pingree and Le Cann, 1992a, 1992b; Garcia-Soto et al., 2002; Caballero et al., 2008). These eddies have been observed to remain stationary for a long time, in some cases up to 7 months (Pingree and Le Cann, 1992b). Due to the location of these eddies near 4°W in different years, they have been named “4°W eddies”. Pingree and Le Cann (1992a) concluded that the 4°W eddies were trapped by the topography, in a recess limited to the west by the combined slope feature of Le Danois bank and Torrelavega canyon and to the east by the Landes Plateau, and that they were usually associated with cyclones. The study of a 4°W eddy by means of SST images, Argos buoys and XBTs (Pingree and Le Cann, 1992b) concluded that (i) the characteristics of this eddy were similar to that of Slope Water Oceanic Eddies (SWODDIES) (Pingree and Le Cann, 1992b); (ii) it had a diameter of 100 km, a vertical extension of >2000 m (XBTs are limited to ~2000 m depth) and a rotation period of ~3.5 days; (iii) from May to September the eddy rotated clockwise without a significant net movement and from October to January the net movement was southeastward.

The interannual variability of the presence/absence of the 4°W eddy has been related to the inflow of the poleward winter slope flow off the northern Spain (*Navidad*). This is an extension, along the Cantabrian coast, of the Iberian Poleward Current (IPC). As with other Eastern Poleward Currents, which are driven by density forcing, the IPC is associated with eddies at the oceanic boundaries, generated as a consequence of the separation or evolution of perturbations in the flow regime. In the western Iberian coast, the IPC is responsible for the generation of migratory eddies (Pingree and Le Cann, 1993) and of persistent anticyclones that remain trapped in the lee of topography (Peliz et al., 2003). When the IPC/*Navidad* enters the Cantabrian coast, where the slope is parallel to the dynamic height contours, this current passes to a decaying phase (Pingree and Le Cann, 1990; Pingree et al., 1999; Le Cann and Serpette, 2009). Le Cann and serpette (2009) observed that the opposing effects of the dynamic forcing along the Cantabrian slope could create instabilities in the slope currents. The limit between both effects is around 4°W: positive (negative) wind stress curl east (west) of this position induce (restrain) the poleward direction of the slope current. These authors also argued that the alongshore slope current decay (and the subsequent instabilities) could be also induced by an offshore currents migration. Pingree et al. (1999) considered that the current decayed due to inertial overshoot relating to slope topography and loss of slope current transport and continuity. East of 5°W, the major change of the coast and slope orientations, together with the presence of some steep bathymetric irregularities (submarine canyons of Torrelavega, Capbreton and Cap Ferret and the Landes Plateau), can be expected to significantly affect the shelf/slope circulation by adding complexity to the regional ocean circulation. Indeed, the area between Capbreton and Cap Ferret canyons is known for its intense mesoscale activity (Le Cann and Serpette, 2009).

Subsurface water masses (>200 m) on the Cantabrian slope show a maximum temperature between mid-September and mid-March, when *Navidad* is developed (Pingree, 1994). A monitoring of this current during the last three decades show that a marked *Navidad* was developed 1/3 of the years (Garcia-Soto and Pingree, 2012) and specifically during the last decade it was observed in 2001, 2003, 2007 and 2010. The 4°W eddy is not generated when *Navidad* is confined to the western Cantabrian coast, but it can be observed when this winter flow is extended along the main part of