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Water mass circulation on the continental shelf of the Gulf of Cádiz

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Abstract

Acoustic Doppler Current Profiler data collected during three successive surveys in the Gulf of Cádiz in May–June 2001 have been used to analyse the surface circulation on the continental shelves of the Gulf of Cádiz and how this circulation matches the circulation in the ocean side of the Gulf. The wider and larger eastern continental shelf holds a cyclonic circulation bounded at the south by a shelf-break jet that is identified with the Huelva front. The coastal current that closes the gyre at the north is identified with the warm counterflow mentioned in the literature. Under westerly winds, this counter current recirculates toward the east while recent upwelled water near Cape Santa Maria is advected downstream by the shelf-break jet, leaving the cold signature at the surface that has been identified historically with the Huelva front. Under easterlies, part of the coastal counterflow invades the western continental shelf while the remaining recirculates eastward to close the cyclonic cell. The western continental shelf and slope is occupied by a larger-scale cyclonic eddy that extends into the deep ocean. This eddy has vertical length scale of hundreds of metres and is linked to the general wind forcing in the area. Both cyclonic structures are bounded at the south by a jet that enters in the Gulf of Cádiz moving around the second eddy and eastward to feed the Atlantic inflow through the Strait of Gibraltar.

Keywords: Cape Santa Maria; Cape San Vicente; Upwelling filaments; Huelva front; Wind-driven flow

1. Introduction

The continental shelf in the Gulf of Cádiz is bounded by the isoline of 100 m depth. Off Santa Maria Cape (Fig. 1) the continental shelf is very narrow (less than 5 km wide) and ends in a continental slope that descends steeply to more than 600 m depth in less than 4 km. To the west, between Cape San Vicente and Cape Santa Maria (CSV and CSM hereinafter, respectively), the shelf widens to 15–20 km and it is cut by the Portimao submarine canyon, which extends down to 2000 m depth. East of CSM the continental shelf quickly widens to more than 40 km off Guadalquivir River (Fig. 1). Therefore, CSM divides the continental shelf into two halves, which may make the circulation in each half independent from the other.

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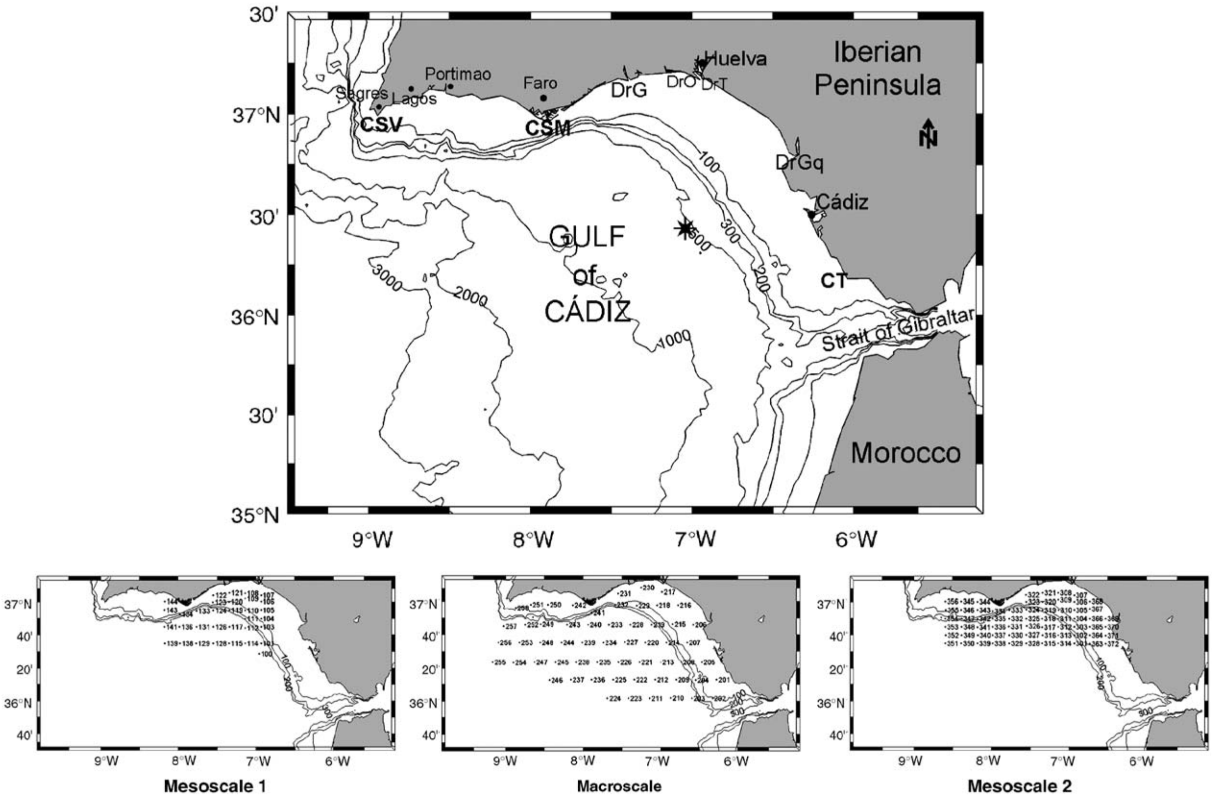

Fig. 1. Upper panel: map of the Gulf of Cádiz showing the position of locations and other geographical features mentioned in the text. CT, CSM and CSV stand for Cape Trafalgar, Cape Santa Maria and Cape San Vicente, respectively. DrG, DrTO and DRGq stand for the mouths of Guadiana, Tinto-Odiel, and Guadalquivir Rivers, respectively. The star marks the position of the *Red de Aguas Profundas* (RAP) oceanographic buoy mentioned in the text. Lower panels show the grid of stations of the three legs (Mesoscale1, Macroscale and Mesoscale2) of GOLFO 2001 survey.

Historically, studies on physical oceanography of the Gulf of Cádiz have focused on the Mediterranean water outflow more than the surface circulation. The reason is the importance that this warm and salty plume has for the large scale circulation of the Atlantic Ocean and, probably, in the global ocean (Ambar and Howe, 1979; Baringer and Price, 1997; Mauritzen et al., 2001; Ambar et al., 2002; Potter and Lozier, 2004). The surface circulation has not been investigated so extensively and the scientific literature is rather limited. Most of the studies deal with remotely sensed sea-surface temperature (SST), or climatological data (Fiúza et al., 1982; Fiúza, 1983; Folkard et al., 1997; Vargas et al., 2003; Sánchez and Relvas, 2003) and only a few papers analyse quasi-synoptic in situ data to depict the three-dimensional (3-D) structure of the mass field and its dynamics.

One of the first studies on the Gulf of Cádiz surface circulation was the descriptive work by Stevenson (1977) who, using infrared SST and visible satellite imagery, identified and named a series of surface thermal features. Stevenson (1977) identified the ''Portuguese upwelling zone'', the ''Huelva front'', the ''Tarifa eddy'', and the accumulation of warm surface water over the Spanish continental shelf. Nowadays, these features are better known but they are still far from being well understood.

The ''Portuguese upwelling zone'' corresponds to the noticeably colder signature of the upwelling area off CSV. According to Mazé et al. (1997) and Sánchez and Relvas (2003), its origin is related to the prevailing positive z-component of the wind-stress curl in the zone, which produces Ekman pumping and, consequently, a cyclonic circulation in the interior to match the isopycnal upraise. The extension of the upwelling area changes following the direction of winds. Westerlies induce coastal upwelling that enhances the former open-sea process and increases the upwelling area, whereas easterlies reduce its extension. Westerlies originate a second and noticeable core of upwelling just east of CSM that, apparently, merges with the more permanent upwelling off CSV (Fiúza, 1983; Folkard et al., 1997; García et al., 2002). In contrary, easterlies favour a westward warm coastal counter current that invades the zone off CSV and displaces the cold upwelled water offshore. If easterly winds persist long enough, the warm counterflow goes around CSV and northwards for some tens of km (Fiúza, 1983; Relvas and Barton, 2002).

The "Huelva front" in Stevenson (1977) is a vaguely defined surface feature identified by the SST contrast between continental shelf and open-ocean waters. Apparently, Stevenson (1977) was referring to a cold filament that extends from CSM towards the southeast along the continental shelf break/slope of the eastern Gulf of Cádiz. Subsequent papers by Fiúza (1983), Folkard et al. (1997), and Vargas et al. (2003), among others, related this cold surface feature to the wind regime in the Gulf of Cádiz, showing that westerly (easterly) generates and enhances (diminishes, even destroys) the surface signature. There is not much information about the 3-D structure of this front. Furthermore, there is no reported evidence of this front in the subsurface density field. The question then arises as to whether the Huelva front is just the surface footprint of upwelling episodes off CSM or whether it has a subsurface presence.

The "80 × 180 nautical miles Tarifa eddy" represents the large-scale anticyclonic circulation in the central part of the Gulf of Cádiz. Actually, different studies carried out in the Gulf of Cádiz confirm the presence of an antyciclonic surface circulation in the middle of the basin with a north boundary current flowing over the continental shelf-break and slope (Pérez-Rubín et al., 1999; García et al., 2002). This circulation prevails during most of the year, and definitely during spring–summer (Vargas et al., 2003; Sánchez and Relvas, 2003). Most studies have a limited offshore extension and do not provide a satisfactory answer to the question of whether this anticyclonic circulation closes a gyre or it is just a large meander of the Portugal–Canary eastern boundary current that intrudes in the Gulf of Cádiz. Criado-Aldeanueva et al. (2006) show that, at least during May 2001, the surface circulation pattern is more likely a meander than a closed gyre. A considerable fraction of the surface volume transport in the meander is diverted towards the Strait of Gibraltar to feed the Atlantic Jet into the Mediterranean Sea (Criado-Aldeanueva, 2004), and the remaining veers southwards and then southwestwards to rejoin the Canary Current.

The clear bias of oceanographic surveys towards spring–summer leads to a permanent anticyclonic circulation in the Gulf of Cádiz. Long current-meter observations collected at 36° 28.8′N, 6° 57.8′W by the *Red de Aguas Profundas* (RAP) network of Puertos del Estado, Spain (see http://www.puertos.es/index.jsp) confirm a prevailing anticyclonic circulation throughout the year (southeastwards currents at the monitoring station) but they also reported northwestwards velocities in wintertime. Mauritzen et al. (2001) suggested a mid depth (100–180 m depth) cyclonic circulation in the Gulf of Cádiz after examining a set of hydrological data acquired in October 1995. Vargas et al. (2003), analysing an 8-year time series of SST images, showed that the spatial pattern of the first empirical mode (explaining up to 60% of the variance) indicates accumulation of warm (and, thus, light) water in the middle of the basin compatible with anticyclonic geostrophic circulation. But the time coefficients of the mode showed important seasonal variability with minimum values in winter, which weakens the horizontal thermal gradient (and, hence, the pressure gradient) and facilitates the reversal of the circulation during this season. These seasonal changes would follow those of the wind regime off the Iberian Peninsula, which veers from northerly (upwelling season) to westerly or southwesterly in winter (Fiúza et al., 1982; Fiúza, 1983; Relvas and Barton, 2002) following the seasonal displacement of the Azores high. Such displacement drives seasonal fluctuations in the circulation of the whole Subtropical Gyre, which in turn, and according to Machín et al. (2006), would induce noticeable seasonal changes in the surface circulation of the Gulf of Cádiz in the manner described above.

Another interesting feature of the water circulation is the warm coastal counter current mentioned above. It flows westward near the shore and, eventually, may reach CSV and proceed northwards. The current responds to wind stress in the sense that westerlies hamper its westward progression while easterlies push it in that direction (Fiúza, 1983; Folkard et al., 1997; Relvas and Barton, 2002). Its origin is not clear. Relvas and Barton (2002) suggested the presence of an alongshore pressure gradient that drives the counterflow. They supported their conclusion by the fact that observations of coastal sea level indicated a sea level slope along the southern coast of Portugal, the surface sloping down to the west, with an increased slope in summer. Using historical data, they computed monthly dynamic heights relative to 500 db in some selected points off the Portuguese coasts with the aim of providing further evidence of the claimed sea level slope and they found a reasonable agreement between both estimations. Mauritzen et al. (2001) invoked the need of continuity of mass to explain this counter current. They argued that the volume rate of Atlantic water that enters the Gulf of Cádiz by its western boundary (9°W) exceeds the flow that proceeds eastward into the Mediterranean Sea through the Gibraltar Strait. Volume conservation implies recirculation of water with modified T–S characteristics back to the Atlantic Ocean. The needed recirculation would take place along the northern side of the Gulf of Cádiz while the T–S modification was basically a salinity gain by interaction with the Mediterranean outflow. Mauritzen et al. (2001) argued that this near-surface net salt export is necessary to explain the fact that, at any depth, the North Atlantic Central Water is saltier in the eastern North Atlantic than in the western North Atlantic.

In this paper many of these features are revised and commented in the light of the new data acquired in the frame of the Spanish-funded MAR99-0643 research project.

2. Data and methods

Hydrological, currentmeter, meteorological and remote sensing data have been processed and used in this work. The in situ data were collected during the GOLFO-01 oceanographic survey carried out in May 2001.

Conductivity temperature-depth (CTD) data in the grid-stations shown in Fig. 1 were recorded by means of a Hydronaut MK 137 CTD probe onboard the R.V. *Hespérides* for the three legs of the GOLFO-01 survey. The different legs will be referred as to Mesoscale1 (from 14 to 16 May, 2001), Macroscale (from 17 to 24 May, 2001), and Mesoscale2 (from 28 May to 2 June, 2001). The spatial coverage and the code of the different stations of each leg are presented in Fig. 1. In addition to CTD data, a Seabird SBE21 thermo-salinographer was continuously recording subsurface (3 m depth) temperature, salinity and fluorescence along the ship track, providing a useful high-spatial resolution data set.

A vessel-mounted RDI-150-NB acoustic Doppler current profiler (ADCP) measured the velocity profile along the ship track. Vertical bins were 8 m wide and the first bin was centred at 16 m depth (12–20 m). A total of 50 vertical cells were resolved (400 m range) though, in general, velocity below 300 m was very noisy and has not been considered here. In the Gulf of Cádiz and a few tens km west of the Strait of Gibraltar barotropic tidal currents are very weak (M2 current is less than 3 cm/s amplitude at the RAP buoy position, see Deep Sea Network data reports in http://www.puertos.es). The same is true for the continental shelf area located far from Gibraltar Strait's influence, that is, the region from Guadalquivir River to the northwest (Baldó et al., 2005). The lack of reliable tidal current predictions in this area and, more importantly the fact that these currents are comparable to the measurement error of the instrument, justify the decision of not removing tidal currents from ADCP data. On the other hand, the region near the Strait of Gibraltar is strongly influenced by tides, a fact that is of particular concern for ADCP observations taken in the southeast area surveyed during Macroscale (Fig. 1). For this reason, ADCP transects within this area have not been considered in this study. The processed ADCP data include an adjustment of the "bottom tracking" and "navigation" modes and an exponential smoothing by the factor $\exp(-ar^2)$, where r is the normalised distance of each observation to the central point of the filtering grid and a is a smoothing filter parameter that has been set to $a = 0.5$ considering the correlation between the measured velocities both in the horizontal and vertical directions (Criado-Aldeanueva, 2004).

Air pressure and wind velocity was recorded by the RAP oceanographic buoy (see location in Fig. 1). Due to its offshore location, these data were considered as representative of the mean wind and pressure fields over the Gulf of Cádiz.

Remote sensing SST data acquired by the advanced very high-resolution radiometer (AVHRR) sensor onboard NOAA-14 satellite were transmitted to the Terascan station of the R.V. *Hespérides*. At least one SST image per day was transmitted and processed. However, cloudy weather prevented the acquisition of good-quality images during most of Mesoscale1.

Residual sea-level predictions provided by a barotropic version of the Hamburg Shelf Circulation Model (HAMSOM) have been used to assess the importance of wind set-up on the shelf circulation. The model is routinely operated by Puertos del Estado as part of the Nivmar sea-level forecast application (Álvarez Fanjul et al., 2001) and gives a satisfactory prediction of the tidal-free component of sea level. The correlation of HAMSOM computed and observed residuals in different ports around Spain is very satisfactory ($r\sim0.87$–0.82; RMSE$\sim$5.2–6.4 cm in a region where mean tidal amplitude is 1.1 m) and increases for subinertial frequencies. These data have been used to monitor changes of the along-shore sea-level slope, which in turn can drive the observed variability of the shelf circulation.

3. Circulation pattern in the continental shelf

The geostrophic approximation is not well justified in shallow continental shelves where friction and other forces have enhanced importance. This is one of the reasons to use the processed ADCP data as the main data source for the velocity field, which, on the other hand, shows good agreement with the geostrophic field in the open ocean where geostrophy is satisfied. Moreover, the spatial scale of mesoscale processes that usually dominate the continental shelf circulation requires sampling with a spatial resolution higher than the typical one provided by conventional CTD grids in order to resolve adequately the structures. The vessel-mounted ADCP yields the needed high spatial resolution along the ship track, which provides a second reason to use the ADCP velocity field. In this Section the high-resolution ADCP transects are used to depict and analyse the main characteristics of the circulation on the continental shelf and slope and in their vicinity.

3.1. *Mesoscale1*

Fig. 2 shows the ADCP velocities and the density field at each north–south transect of mesoscale1. The fields of density and ADCP velocity show a clear tendency to be coupled geostrophically to each other. For instance, Figs. 2E–G show a marked offshore density gradient over the continental shelf-break in the place where a relatively swift jet flows eastward. This tendency, which is also present during the two other legs, will be useful to analyse some features of the shelf circulation.

There are two distinguishable cores of eastward flowing water. The first one, named "N2" in panels of Fig. 2, is clearly seen in all transects. It is linked to the large-scale circulation and plays the role of a boundary current for the basically geostrophic flow in the open sea side of the Gulf of Cádiz. The second one, named "N1" in panels of Fig. 2, appears weakly in transect 5 (Fig. 2D) for the first time and it is clearly visible in all transects eastwards of transect 5. It is more related to the shelf and slope circulation than to the open-sea dynamics. Near the eastern end of the surveyed region, the cores approach each other, to finally merging at the easternmost transect (Fig. 2H). In transects 5 to 2, however, the two cores are distinguishable.

The arguable distinction of the cores based on the visual inspection of ADCP velocity is confirmed by the T–S characteristics of the water. Fig. 3 shows that the surface water in N2 is always warmer and saltier than in N1. This suggests that core N1 contains upwelled water since the source of cold and fresh water is the NACW beneath the surface layer, whereas water in N2 has been made saltier and warmer because of a longer exposure to air-sea fluxes. Volume transport within N1 progressively increases from around 0.12 Sv through transect 4 to 0.28 Sv through transect 2, suggesting water entrainment from below and/or from the side, which is partially supported by the evolution of the T–S characteristics in N1 through the successive transects (Fig. 3).

Water-mass circulation in or near the shelf has distinctive patterns east and west of CSM. Figs. 2F–H show a coastal counter current that flows westward over the eastern shelf of the Gulf. There are no traces of this counter current in the western shelf, which suggests that the current recirculates eastwards as it approaches CSM, thereby closing a cyclonic cell over the shelf. This recirculation probably contributes to the increase of the volume transport of N1. The counter current occupies the inner continental shelf in depths less than 70 m, which are clearly insufficient for the geostrophic balance to stand. However, the doming of isopycnals in Fig. 4C (see also Fig. 2F) supports some degree of geostrophic adjustment. The thermal–wind relationship applied to the density field of Fig. 4C gives a geostrophic westward velocity near the shore of 13 cm/s at 16 m depth relative to the bottom. The ADCP velocity at this depth is 10–12 cm/s (Fig. 4B), slightly less than (but quite comparable to) the geostrophic counterpart, the discrepancy being attributed to frictional effects. The accumulation of light water near the shore along with the shoaling of isopycnals (Fig. 4C) gives

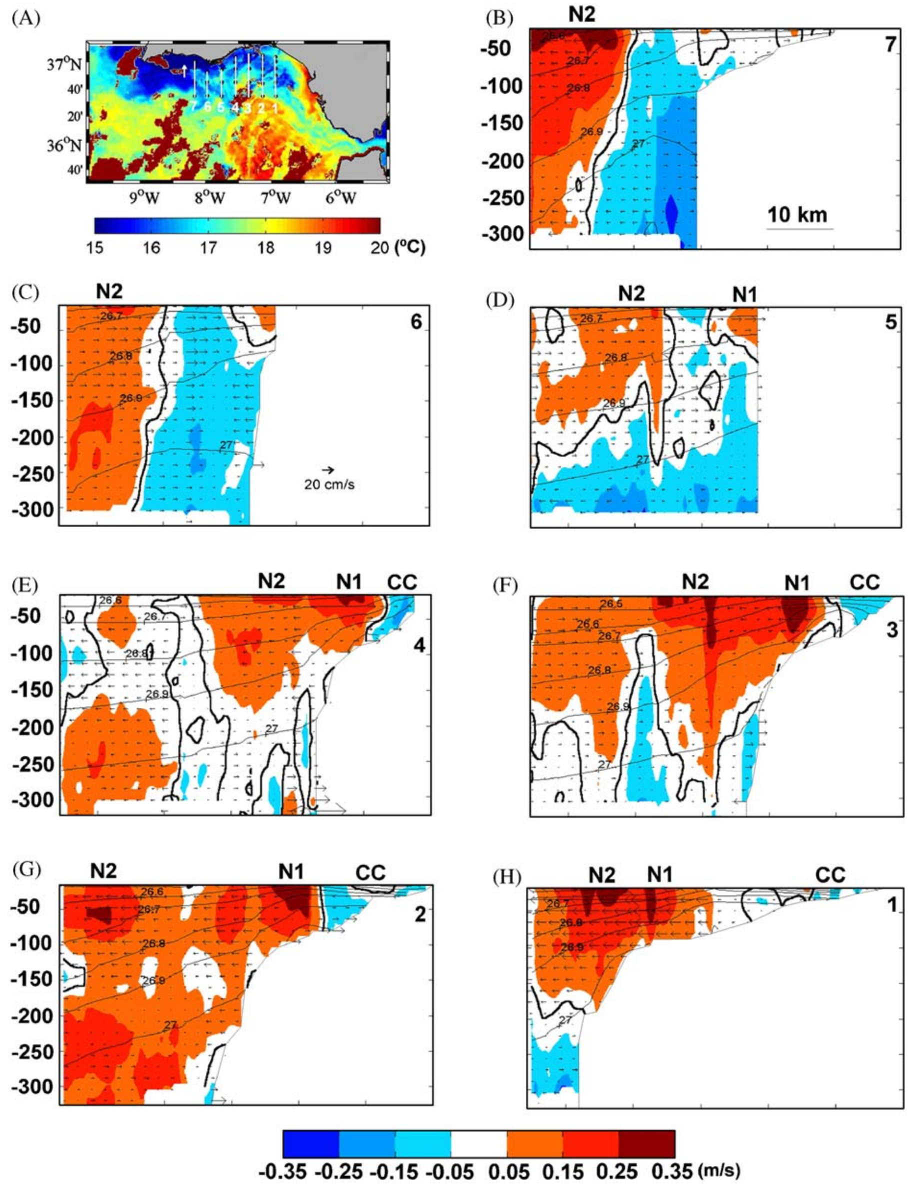

Fig. 2. (A) SST image of 13 May 2001 showing the different ADCP transects (white lines) and the code assigned to each of them during Mesoscale1. Panels (B)–(H) show the ADCP velocity in the following way: colour-shaded contours indicate the eastward (normal-to-transect) component of the velocity with positive values to the east and arrows represent the along-transect component. Black lines are contours of sigma-t. The transects have different lengths but they start at the same latitude (as indicated by the white lines in panel A) and have the same horizontal scale, which is given by the scale in panel (B). Cores N1 and N2 and the coastal counter-current (CC), when present, are indicated in each panel. Numbers in the lower right corner refer to transect number. The SST image of panel (A) is the day before Mesoscale1 survey started.

support to the existence of a cell of cyclonic circulation. Notice that temperature makes the density diminish shoreward (Fig. 4B), which agrees with previous observations in the zone that report accumulation of warm water in the inner part of the shelf (García et al., 2002; Relvas and Barton, 2002; Sánchez and Relvas, 2003). A rough estimation of the volume transport in the counter current gives a moderate 0.03 Sv with typical velocities of 10–15 cm/s to the west. These values are rather constant through the different transects in the eastern shelf during mesoscale1.

The circulation west of CSM is different from the pattern depicted above. While core N2 is quite clear

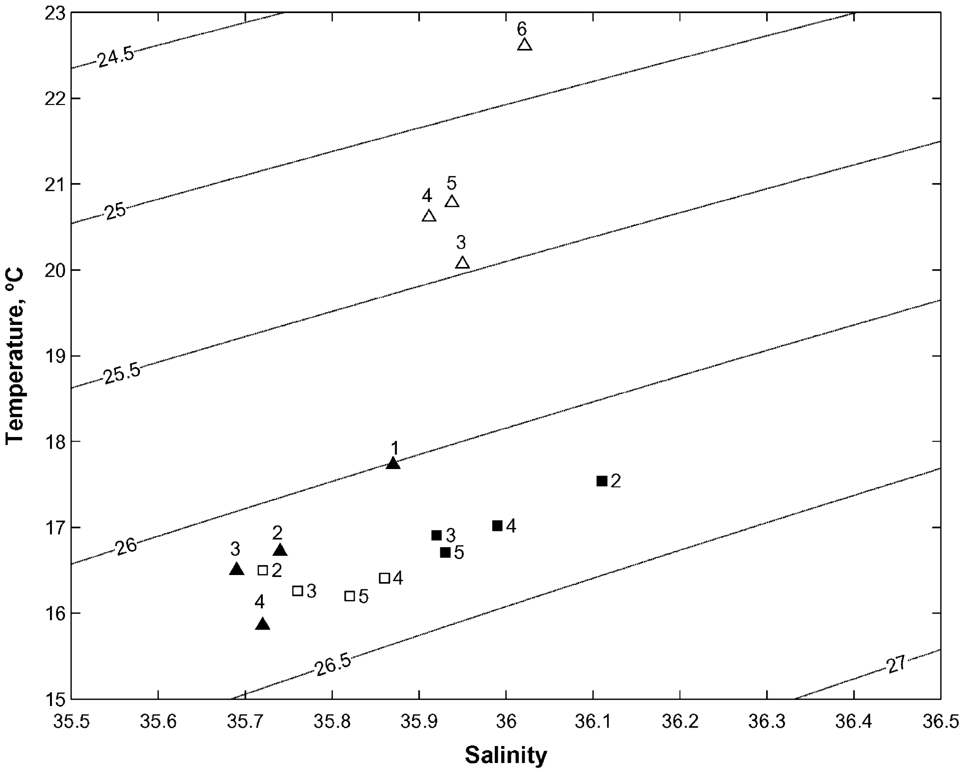

Fig. 3. T–S characteristics of surface water in core N1 (open squares) and N2 (full squares) during Mesoscale1. Numbers beside the symbols indicate the transect number. Triangles indicate T–S characteristics of the surface water in the coastal counter current for Mesoscale1 (full triangles) and Mesoscale2 (open triangles). Numbers above the symbols indicate transect number. Notice that transects 6, 5, 4 and 3 of Mesoscale2 correspond to transects 4, 3, 2 and 1 of Mesoscale1, respectively. All these data are from the continuously recording thermo-salinographer.

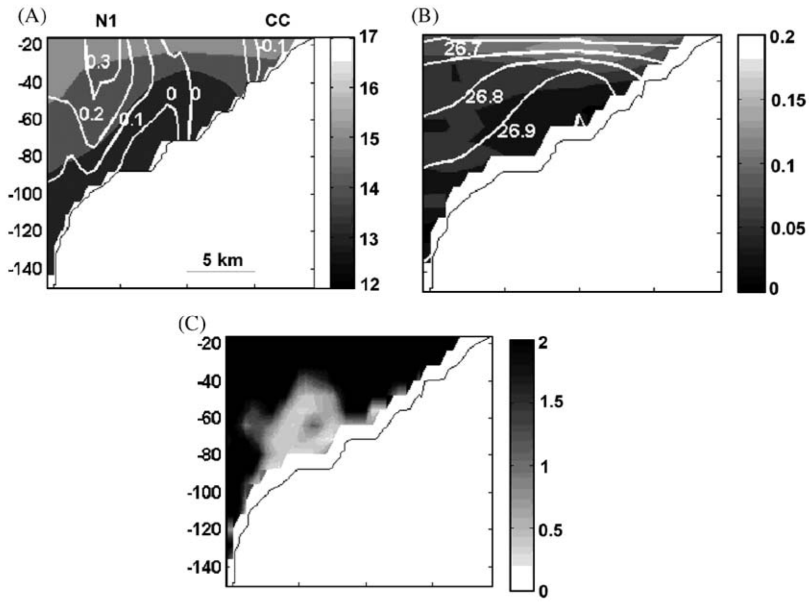

Fig. 4. (A) Contours of temperature (filled) and cross-transect velocity (lines) along the northern half of transect number 3 of Fig. 2A. (B) Contours of Brunt-Väisälä frequency (min^{-1}, filled) and sigma-t (lines); and (C) contours of the gradient Richardson number. Core N1 and the coastal counter-current (CC) are indicated in panel (A). The transect was sampled on 15 May 2001.

in transects 5–7 (Figs. 2B–D) there are neither traces of core N1 west of the Cape, nor of the coastal counter current. There is, however, a clear westward circulation that extends from the surface down to depths greater than 300 m near the steep continental slope. The velocity field in these western transects (not only the east–west component, but also the north–south one—Figs. 2C and D) strongly suggests the presence of a mesoscale cyclonic eddy whose centre would be located west of the westernmost transect, thus avoiding our sampling. Volume transport around the eddy is important: using Fig. 2B as reference, the westward transport in the upper 300 m exceeds 0.45 Sv, while it is more than 0.55 Sv towards the east inside N2. Notice that this last volume transport is hardly indicative of the overall N2 transport since the computation misses an unknown—but necessarily important—amount of eastward transport south of the southernmost station of the transect.

3.2. Macroscale

Fig. 5 shows a selection of transects during Macroscale. Only the area north of 36°20′N, which roughly coincides with the area sampled in Mesoscales1 and 2, has been considered. The two easternmost transects closer to the Strait of Gibraltar have been excluded from the analysis due to the tidal influence. The direct comparison of Fig. 5 with Figs. 2 and 6 below is hampered because of the sparser CTD sampling and the oblique orientation of the stations grid. Nevertheless, many of the features identified in Fig. 2 are recognisable in Fig. 5.

Core N1 appears for the first time in transect 7 (Fig. 5F), which is the counterpart of transect 5 (Fig. 2D) in Mesoscale1, and it is quite visible in transects 6, 5 (Figs. 5G and H, respectively), 4 and 3 (not shown). The coastal counter current is visible in transect 4 (not shown) and appears well developed in transects 5–7 (Figs. 5H, G and F, respectively). The volume transport is greater than in Mesoscale1 and shows important spatial variability. For instance, volume transports through transects 5–7 are 0.06, 0.12 and 0.04 Sv to the west, respectively. Conservation of mass implies an important recirculation between transects 6 and 7 and, in fact, the volume transport to the east in core N1 increases from 0.17 Sv in transect 7 to 0.32 Sv in transect 6. As in Mesoscale1, the doming of isopycnals in the shelf, visible in Fig. 5G, indicates near-geostrophic cyclonic circulation around the doming. The surface geostrophic current relative to the bottom associated to the horizontal density gradient is estimated as 18 cm/s through the thermal–wind relationship. Again, there is a good agreement with the ADCP velocity, which is of the order of 20 cm/s. These velocities are greater than the observed during Mesoscale1. Contrary to the observations in Mesoscale1, in this leg the counter current seems to continue flowing to the west through transect 8. However, the computation of transport here is difficult since the counter current has apparently merged with the much greater west-going flow of the western cyclonic eddy (Fig. 5E). There are traces of the counter current in transect 9 (Fig. 5D) with a volume transport of around 0.01 Sv and, even, in transect 10 (Fig. 5C) with a negligible transport of less than 0.005 Sv. Both transects are west of CSM. The diminution of the volume flow from 0.04 Sv in transect 7, east of CSM, to 0.01 Sv in transect 9, west of CSM, implies eastward recirculation in the vicinity of the Cape, as it happened in Mesoscale1.

The circulation on and near the shelf in the western part of the Gulf of Cádiz compares well with the description provided for Mesoscale1. Figs. 5B–E show clear traces of a larger scale cyclonic circulation that extends down to depths greater than 300 m. This would correspond to the large cyclonic eddy associated with the permanent upwelling off CSV. Westward transport within the eddy exhibits important spatial variability, from more than 0.5 Sv through transect 12 (Fig. 5B), to less than 0.2 Sv through transect 10 (Fig. 5C), to increase again above 0.4 Sv through transect 8 (Fig. 5E).

Another difference with Mesoscale1 is the more offshore trajectory of core N2, which is easily identifiable in the western part of the basin. This agrees with the geostrophic description in Criado-Aldeanueva et al. (2006), who show that core N2 bifurcates off CSM. One branch flows along the continental shelf-break joining N1 and the other flows to the east-southeast to veer north further on and joint N1 by transect 5 (Fig. 5H). Consequently, the volume transport in N1 almost duplicates from transect 6 (0.41 Sv) to transect 5 (0.75 Sv) where it has already merged with the second branch.

3.3. Mesoscale2

The features described for Mesoscale1 and Macroscale are identified also in Mesoscale2

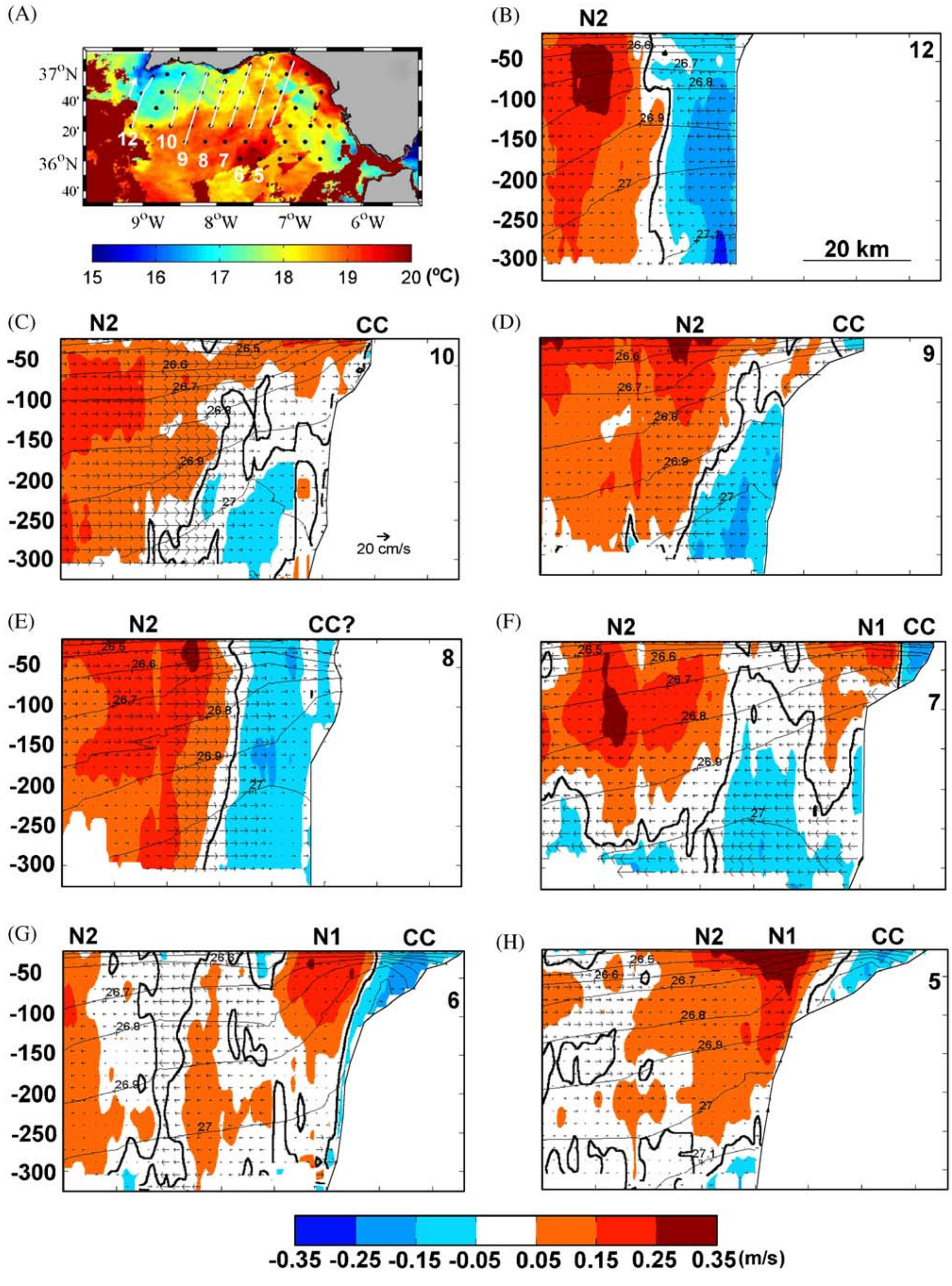

Fig. 5. Same as Fig. 2 for Macroscale. The SST image in panel (A) corresponds to 21 May 2001, when Macroscale sampling was under way. Notice that the oblique orientation of transects means that the normal-to-transect velocity is rotated to the southeast. In other words, it does not coincide with the east–west component of the velocity.

(Fig. 6) and will not be commented upon in detail; but some points are noteworthy. First, it is again in the first transect east of CSM that core N1 appears (Fig. 6F, see caption). Second, the coastal counter current reaches farther west in the western shelf than it did in Macroscale. Third, the volume transport of this counter current in the eastern shelf is noticeably greater than in Mesoscale1 (from 0.03 Sv on average during Mesoscale1 to around 0.1 Sv in Mesoscale2). Fourth, T–S characteristics of the surface water in the counter current are 4 °C warmer on average and saltier than in Mesoscale1

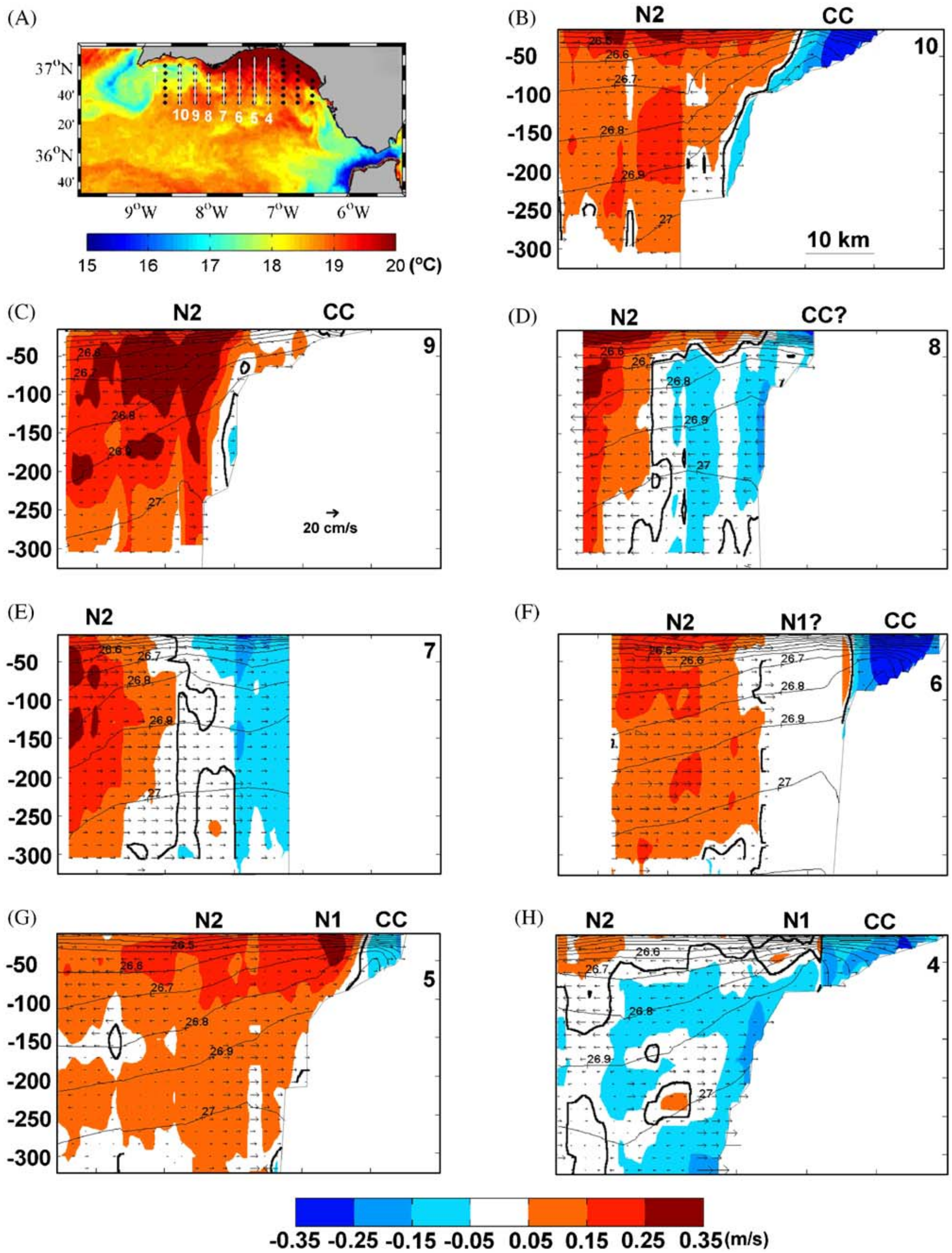

Fig. 6. Same as Fig. 2 for Mesoscale2. The SST image in panel (A) corresponds to 29 May 2001. ADCP failed in part of transect 6 (panel F) just at the expected location of core N1. The change of sign of the alongshore velocity over the shelf-break not only defines the offshore extension of the coastal counter current but also suggests the presence of the shelf-break jet N1.

(Fig. 3). Fifth, there is again significant recirculation of this counter current near CSM as evidenced by the diminution of the volume transport from 0.2 Sv through transect 6 (Fig. 6F) east of the Cape to 0.05 Sv through transect 8 (Fig. 6D), west of the Cape. The coastal counter current invades the western shelf, but its volume flow diminishes when passing by the Cape. SST in Fig. 6A strongly suggests the intrusion of this warm water attached to the Portuguese coast. Finally, the doming of isopycnals in the eastern shelf is found again (Figs. 6F–H). The thermal–wind relationship indicates a geostrophic surface counter current relative to the bottom of around 16 cm/s at 16 m depth in transect 5, in good agreement with ADCP observations (Fig. 6G). As suggested by the SST image of Fig. 6A, the origin of the isopycnal sinking near the shore is the higher temperature of the water near the shore in this case.

4. Discussion

4.1. Schematic of open ocean and shelf circulation

The open-sea (depths greater than 200 m) surface circulation in the Gulf of Cádiz during GOLFO-2001 survey is anticyclonic with short-term, meteorologically induced variability that changes geostrophic streamlines and volume transport noticeably in less than two weeks (Criado-Aldeanueva, 2004). In spite of this variability, the current that flows over the continental shelf break and slope in the eastern half of the basin is a permanent, yet fluctuating, feature in all three legs. This current is both the northeastern boundary of the ocean side circulation of the Gulf of Cádiz and the southern boundary of the continental shelf circulation.

Our analysis indicates that this current is made up of two contributions, named cores N2 and N1, respectively, the last one only being detected in the eastern part of the basin. Core N2 seems to be part of the Portugal Current system that flows southward off the western Iberian coast during spring and summer, the upwelling season (Mazé et al., 1997). According to Relvas and Barton (2002), the most preferred direction for the spreading of this current at the latitude of CSV is eastward along the southern shelf and slope of Portugal. Fig. 7, which summarizes the surface circulation inferred from our analysis, outlines this trajectory and the motion of core N2 around the cyclonic eddy off Cape San Vincent (SVE) and eastwards in the Gulf of Cádiz. This core may bifurcate off CSM to generate identifiable filaments such as that depicted in Criado-Aldeanueva et al. (2006). East of CSM, the N2 core approaches the continental shelf break and merges with core N1 tens of km downstream of the Cape.

Core N1 and the coastal counter current are another rather stable features identified in the eastern continental shelf/slope throughout the three legs. They appear to be part of a mesoscale (and shallow) cyclonic circulation cell that occupies the eastern shelf between CSM and Guadalquivir River mouth in the east. The isopycnic doming in the centre of the structure below the incipient thermocline (Fig. 4C) suggests a tendency for geostrophy, a fact further confirmed by the direct application of thermal–wind equation to the observed density fields. Fig. 4D suggests a likely relationship between the isopycnic uprising and the shear instability at

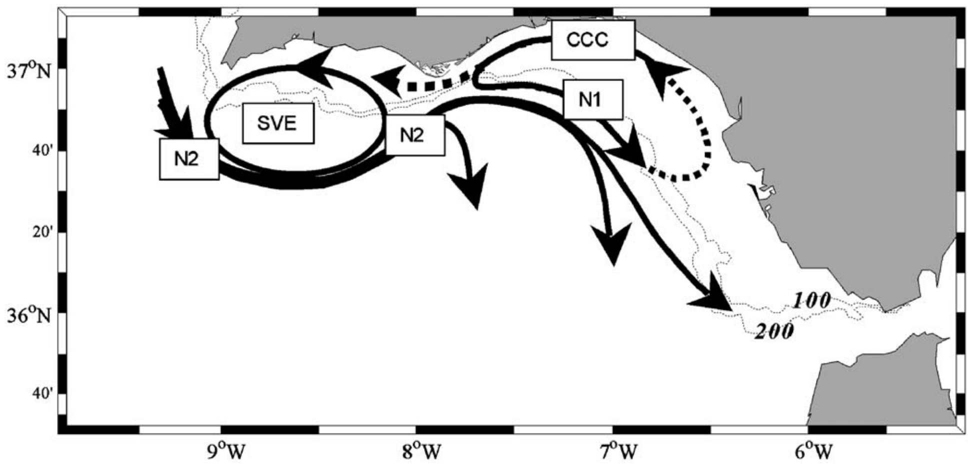

Fig. 7. Sketch of the surface circulation in the Gulf of Cádiz as deduced from our analysis. Core N2 is a branch of the larger-scale Portuguese–Canary eastern boundary current that veers eastward into the Gulf of Cádiz. It moves around a cyclonic eddy off Cape San Vicente (SVE), which is a quasi-permanent feature of the circulation in the Gulf associated with positive wind-stress curl. The core moves further east toward the Strait of Gibraltar to feed in part the Atlantic inflow into the Mediterranean and veers southwards to re-join the Canary current (Criado-Aldeanueva et al., 2006). The eastern shelf is dominated by a cyclonic circulation bounded by a shelf-break front (core N1, identified with the Huelva front) at the south and a—warmer—coastal counter-current (CC). The presence of Cape Santa Maria is important to close the cyclonic cell at the west. The dashed line at the southeast indicates the likely closure of the cell, although there are no data to support this hypothesis. With easterlies, the coastal counter current bifurcates off Cape Santa María and a branch invades the western shelf (dashed arrow) making the SVE drift to the south. The spatial extension of SVE is variable and it is exaggerated in the sketch.

the base of N1 core, where the flow has Richardson numbers not far from the critical value 0.25, developing Kelvin–Helmoltz instability. This instability, which is greatly favoured by the very low values of the static stability at the base of core (Fig. 4C), converts part of the kinetic energy of the shelf break jet into potential energy of the flow through vertical mixing. This is one likely contributing mechanism to the formation of shelf-break fronts that are often found over the outer continental shelf and/or the shelf break (Gawarkiewicz and Chapman, 1992; Matsuno et al., 1997). The isopycnic doming in turn creates horizontal density gradients that could partially drive the alongshore counter current on the inner shelf. In other words, the coastal counter current would have a geostrophic component that would be ultimately related to Kelvin–Helmoltz instability of the shelf-break current.

The presence of CSM makes the coastal counter current recirculate (partially at least) towards the east, thus feeding core N1. However, under favourable wind conditions (easterlies), this counter current may surpass CSM and progress westwards as it did during Mesoscale2. This is, otherwise, a well-known situation mentioned in Fiúza et al. (1982), Relvas and Barton (2002), and Sánchez and Relvas (2003) among others. In this case, the coastal counter current would meet the west-going part of SVE, and even displace this cold eddy southwards, as happened during Mesoscale2 (Fig. 6A). A similar behaviour has been reported in Relvas and Barton (2002) from SST imagery as a response of the coastal counter current to favourable downwelling winds.

4.2. Short-term variability

The comparison of SST maps of Figs. 2A and 6A suggests that the surface circulation in the Gulf of Cádiz underwent noticeable changes from Mesoscale1 (14–16 May) to Mesoscale2 (28 May–2 June). The driving force for this change was wind stress that changed from westerly during the first leg to an important episode of easterly during most of the last leg (Fig. 8B). However, the subsurface circulation a few tens of metres below the sea surface is rather insensitive to this meteorological change (Criado-Aldeanueva, 2004). Specifically, the shallow cyclonic circulation cell over the eastern continental shelf (including core N1 and the coastal counter current), the deep cyclonic eddy off CSV, and the meandering N2 jet that apparently feeds the Atlantic inflow into the Mediterranean Sea are easily identified throughout the three legs. But there are also discrepancies. Criado-Aldeanueva et al. (2006) show that the geostrophic volume transport towards the Strait of Gibraltar is largely reduced during Mesoscale2, which agrees with the observed decrease of Atlantic inflow through the Strait of Gibraltar under easterlies (García-Lafuente et al., 2002).

Regarding the shelf circulation, one obvious change is the progression of the costal counter current beyond CSM, driven by the moderate to strong easterly that prevailed during Mesoscale2 (Fig. 8). The time evolution of this coastal current under such wind-stress has been identified and widely discussed through the analysis of SST sequences, using surface temperature as a tracer of the flow (Fiúza, 1983; Folkard et al., 1997; Relvas and Barton, 2002). ADCP observations collected during Mesoscale2 provide strong support to the conclusions obtained from those analyses, and confirm, in fact, that surface temperature is a good indicator of this density coastal current. The agreement is not only qualitative but also quantitative: Relvas and Barton (2002) computed an averaged flow speed of 16–19 cm/s after analysing several sets of SST sequences depicting the west-going progression of the warm plume during an episode of persistent easterlies. These values are fully compatible with the approximately 20 cm/s recorded by the ADCP in the western shelf during Mesoscale2 (Fig. 6B).

A second and more significant difference between Mesoscales1 and 2 is the change of SST pattern that took place in only 15 days (Figs. 2A and 6A). The spatially averaged surface temperature over the shelf region increased in more than 3 °C between both surveys, and more than 5 °C in specific places such as the Guadiana mouth near 7°W. This large increase of surface temperature cannot be explained by local heat gain, but by at least two other concomitant facts: the interruption of coastal wind-induced upwelling that follows the halting of westerlies and the release and subsequent westward spreading of the warm-water pool that westerlies had piled up against the Guadalquivir River mouth and Cádiz embayment (compare the sequence of SST images in Figs. 2A, 5A and 6A). Apparently, the Huelva front (Stevenson, 1977) faded out during Mesoscale2 (Fig. 6A) following the change of the wind regime from westerlies (Mesoscale1 and Macroscale, Fig. 8) to easterlies (Mesoscale2). The

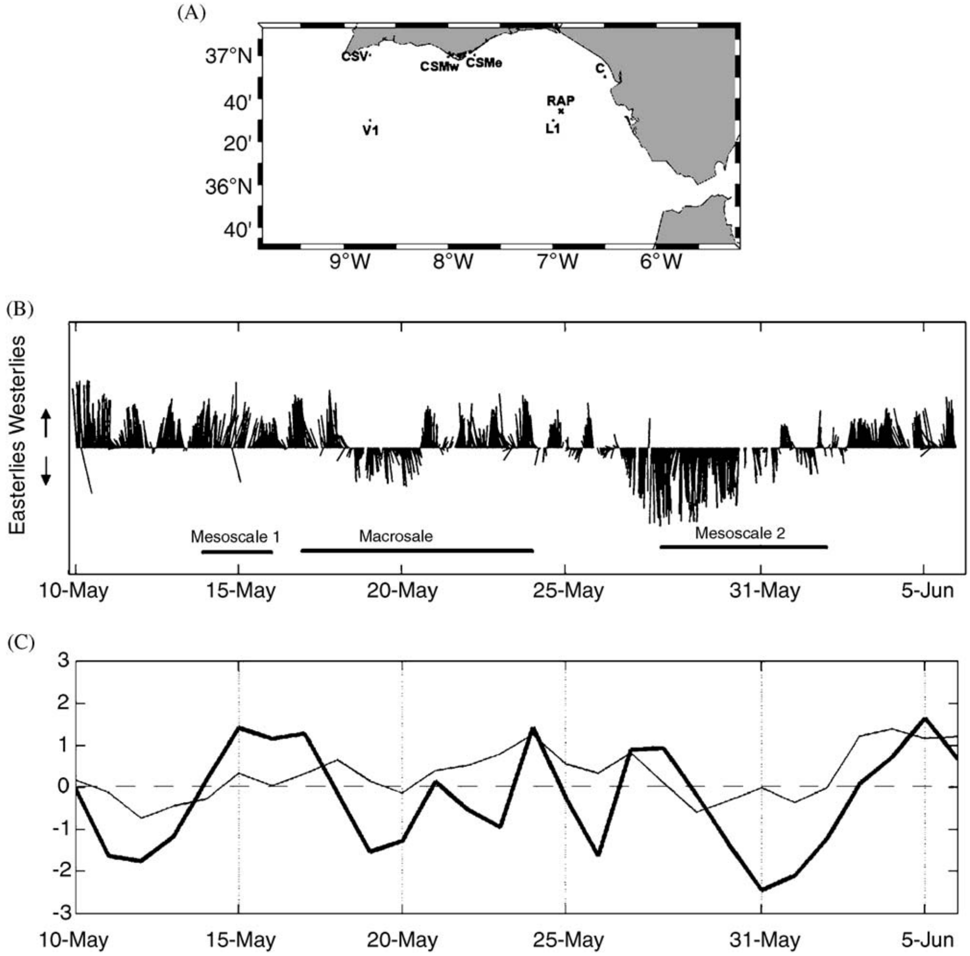

Fig. 8. (A) Location of points in the Gulf of Cádiz mentioned in this Figure and in the text. L1 and V1 indicate the location of points in Relvas and Barton (2002); RAP is the position of the oceanographic buoy of the Puertos del Estado network. (B) Wind velocity recorded at RAP position. (C) Daily values of sea-level difference (cm) between points C and CSMe (C-CSMe; thick line) and CSMw and CSV (CSMw–CSV; thin line) computed from the output of the barotropic model Nivmar.

evolution agrees with the wind-dependent nature of this surface feature (Fiúza, 1983; Folkard et al., 1997; Relvas and Barton, 2002; Vargas et al., 2003). However, ADCP data indicate that the subsurface along slope current (core N1) continued flowing to the southeast (Figs. 6F–H). Should the so-called Huelva front have a subsurface signature, it would coincide with core N1, in which case it would be a density front, as panels B and C in Fig. 4 suggest. Whether or not the front exhibits thermal signature depends on the wind regime. Upwelling favourable westerly wind brings cold water to the surface near CSM either by vertical advection from the depth in the proximity of the Cape, or by horizontal advection from the CSV upwelling area. The subsurface front will carry this cold water downstream, making the cold surface signature historically identified with the Huelva front appear. If winds are from the east, the cold-water supply near CSM stops, as will the cold signature, suggesting the disappearance of the front. It is only the surface footprint that disappears, the front still existing few tens of metres below the sea surface. Curiously, temperature contributes more than salinity to create the subsurface horizontal density gradient, even under easterly conditions (figure not shown).

4.3. A possible origin of the eastern shelf circulation

The observations analysed in Section 3 suggest that the coastal counter current and the core N1 form part of the same circulation cell. The origin of that current has been linked to the presence of an along-shore pressure gradient (Relvas and Barton, 2002), which is stronger in summer. It would be a convincing driving mechanism if it were not that core N1 is flowing against this sloping sea surface just a few tens of km south of the counterflow without any apparent deceleration. The alleged along-shore dynamic height slope between points V1 (higher) and L1 (lower) in figure 15 of Relvas and Barton (2002) (see also Fig. 8) would drive the coastal current but also prevent the large scale circulation shown by Criado-Aldeanueva et al. (2006). This dynamic height slope should be more related to geostrophy (cyclonic circulation around point L1) than to a density current.

The along-shore pressure gradient necessary to drive the counterflow must be confined to the inner part of the shelf and vanish some tens of km offshore. It is likely related to the pool of warm water that systematically appears off the Guadalquivir River mouth and Cádiz embayment in the SST images from spring (April) to autumn (October) (Folkard et al., 1997; Vargas et al., 2003; Navarro, 2004; particularly illustrative is the monthly SST climatology presented in this last work). Let us consider the origin of this warm water.

The presence of localised spots of water warmer than average needs a heat supply to maintain the surface thermal signature against loss via heat advection and diffusion. Off Guadalquivir River mouth and Cádiz embayment, the heat source is apparently the land. This hypothesis would be confirmed by the SST pattern of Fig. 2A, which shows the spot of warm water deformed by the eastward advection forced by westerlies and, also, by the pattern in Fig. 6A, which now shows the spot deformed by the westward advection forced by easterlies (see Fig. 8 for the wind directions during the days of the images). In this area, M_2 amplitude is more than 1 m and the tide progresses inland through the different arms of the Guadalquivir River around 100 km, flooding a few km^2 of marsh. The same happens in the neighbourhood of Cádiz embayment. The vertical mixing driven by tidal propagation, the greenish colour of water in the river along with its shallowness, the increased length of daylight during spring/summer and, specially, the flooding of marshes that have been heated by sun radiation during the previous low tide (notice that, at least, a low tide occurs during daylight), all these facts lead to a greater absorption of energy per mass unit in the river than offshore during spring–summer daylight. Most of this energy is brought back to the sea during ebb tide. During night, there must be no difference between the amount of heat that a parcel of water in the river or offshore gains or loses. Therefore, a daily average gives a positive net heat export (buoyancy input) from the land to the sea during spring–summer. This heat export is not dependent on whether or not the river has net freshwater discharge (discharge is negligible during spring/summer, the dry season) but rather on the tidal dynamics and the heating of land relative to water. Notice also that this mechanism is extensible to nutrient pumping from land into the sea, a mechanism invoked by Reul et al. (2006) to explain some unusual features in coastal phytoplankton distribution. In winter/autumn, the process will be the opposite and the land imports heat from the sea. Monthly SST climatology (Navarro, 2004) shows that this area becomes cooler than average from November to March.

The accumulation of warmer (and therefore, lighter) water in this part of the shelf creates an east-to-west along-shore sea-surface slope that, if not baroclinically compensated, produces a pressure gradient in the interior. Wind stress plays an important role in this dynamics. Westerlies increase this sea-level slope as they pile up water against the coast. Wind drag cancels the excess of sea-level gradient force, thus preventing a very surface counterflow to the west. Below the surface layer, wind stress cannot cancel the pressure gradient and water moves down this pressure gradient towards the west (at a very different spatial scale, this mechanism is reminiscent of the accumulation of warm water in the western Pacific and of the Pacific equatorial undercurrent during non El Niño years). Fig. 8C shows that Nivmar application during Mesoscale1, which was accomplished under westerlies (Fig. 8B), predicts this wind set-up and therefore the excess of westward sea-level slope (Nivmar model cannot predict the steric slope because it does not include any heat forcing). This wind set-up disappeared in Mesoscale2, when wind was mainly from the east (according to Nivmar model, the wind created the opposite sea-level slope). The warm water pool was no longer retained near Guadalquivir River mouth, but was released, invading most of the upper surface layer of the eastern and (partially) western continental shelves (compare Figs. 2A and 6A). It is interesting that Nivmar model predicts much larger sea-level slopes along the eastern continental shelf than along the western one, indicating the importance that the shoreline orientation has on the wind set-up.

5. Conclusions

The surface circulation in the nearshore area of the Gulf of Cádiz during May 2001 was dominated by two cells of cyclonic circulation located over the eastern and western continental shelves. These cells, which are separated from each other by CSM, are coupled to the open sea-surface circulation. Fig. 7 outlines the circulation pattern inferred from our analysis. It is worth-mentioning that the general circulation in the Gulf of Cádiz is part of the more complex Portuguese–Canary (eastern boundary) current system and therefore it is largely wind-driven.

The two cyclonic cells are of different nature. The cyclonic circulation over the eastern shelf seems to be linked to coastal processes such as the presence of an important pool of warm water off the Guadalquivir River mouth and Cádiz embayment, which could produce the necessary sea-level slope to drive the coastal counter current that closes the cyclonic circulation at the north of the cell. Typical volume transports in this cell are around 0.05 Sv. The cyclonic cell leaves a geostrophic footprint in the density field (doming of isopycnals on the shelf). It is speculated that dynamic instability at the bottom of the shelf-break jet that closes the cyclonic cell at the south edge (core N1, identified with the Huelva front) contributes to the homogenisation of waters on the middle shelf, and hence to isopycnic doming.

The cyclonic eddy found in the western part of the basin would be related to open-sea processes, specifically the sea response to a positive *z*-component of the wind stress curl off CSV, as suggested by Mazé et al. (1997) and Sánchez and Relvas (2003). This hypothesis is supported by (i) the vertical coherence and deep reach of the currents recorded by the ADCP, (ii) by the large volume transport (over 0.5 Sv) that recirculates within this cell, and (iii) by the clear signature that produces in the larger-scale geostrophic velocity field (Criado-Aldeanueva et al., 2006).

Easterly winds, if persistent enough, may connect both cells by pushing warm water of the coastal counter current in the eastern cell to the west of CSM, a process repeatedly illustrated in SST images (Fiúza, 1983; Relvas and Barton, 2002). In these circumstances, the large cyclonic eddy on the western shelf is displaced seawards by the warm tongue of water coming from the east, which remains attached to the shore while progressing westwards (Fig. 6A).

The clear bias of oceanographic surveys towards spring–summer suggests a permanent anticyclonic circulation in the open sea and a more or less complex circulation on the shelves in the manner depicted in this paper. Surface currents observed by the oceanographic RAP buoy (see Fig. 1 for location) show a clear tendency to flow towards the southeast during most of the year. This is the expected direction of the flow at the RAP position for an anticyclonic circulation in the ocean side of the Gulf of Cádiz. But they often flow towards the northwest in winter, when the wind regime off the Iberian Peninsula changes from northerly during the upwelling season to westerly or southwesterly. The change of the mainly wind-driven, large-scale, surface circulation in the Gulf of Cádiz must be echoed by the shelf circulation. Actually, near-shore current observations during November–December 2001 show inversions that are not correlated with local wind stress (Sánchez et al., 2006), indicating remote forcing probably linked to seasonal fluctuations of the basin-scale wind driven circulation, as suggested in Machín et al. (2006). Such winter changes in the coastal counter current would be supported by two reasons. First, core N1, which is part of the antyciclonic circulation of the ocean side in spring–summer, the upwelling season, would disappear if this circulation changes to cyclonic. This in turn would have an impact on the circulation over the eastern shelf, which could not hold the closed cyclonic cell depicted in Fig. 8 any longer, and therefore could not favour the maintenance of the coastal counterflow. Secondly, the buoyancy input at Guadalquivir River mouth, whose origin and dynamic role has been discussed in Section 4.3, stops in winter. Therefore, the alongshore sea-level slope that drives the coastal current, and hence the coastal current itself also will disappear.

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