A high-resolution modeling study of the Western Iberian Margin mean and seasonal upper ocean circulation

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Abstract:

The mean seasonal hydrography and circulation of the Western Iberian Margin (WIM) are studied by means of a high-resolution configuration of the Regional Oceanic Modeling System. A comparison of 5-year model averages for January and July with climatological datasets shows a general good agreement in the reproduction of the mean water mass properties and hydrographic distribution. We find that there is a prevailing tendency for slope poleward flow at about 80–100 km offshore at all latitudes from the surface to 1,500 m with strong vertical coupling. This northward flow, which is mainly along slope and amounts up to $8\text{--}10\text{ cm s}^{-1}$, exhibits several mean flow recirculation regions on its way and evidences of an offshore pathway of poleward flow. Transports at different zonal sections further confirm the poleward flow tendency with two peaks of poleward transport in summer (3–10 Sv) and winter (2–7 Sv). The transport time series emphasize the seasonal character of the alongshore circulation and the interannual intrinsic variability of the circulation, since the forcing fields are climatological. As a conceptual essay with the purpose of assessing the Mediterranean Water flow influence on the WIM mean circulation, a second model configuration is setup, where the Mediterranean outflow into the study domain is removed. We find that there is an attenuation of the mesoscale field, but the slope poleward flow intensifies and remains as a mean dynamical feature closer to the upper slope.

Keywords: Western Iberian Margin ; Numerical modeling ; Poleward flow ; Upwelling systems

1 1. Introduction

2

3 The Western Iberian Margin (WIM) (see Figure 1, domain SD) is located between
4 two offshore large-scale oceanic systems: to the north, the North Atlantic Current
5 flowing in the NE direction; to the south, the Azores Current flowing eastward with
6 complex interactions with the Gulf of Cadiz. This region is characterized by a weak
7 large-scale circulation, mostly southward, with typical velocities of a few centimeters
8 per second, and frequently named the Portugal Current (Saunders, 1982).

9 At the coastal transition zone, the circulation along the WIM is largely influenced by
10 the presence of the atmospheric anticyclonic system named Azores High, whose
11 position oscillates between northern locations off the Iberian Peninsula during
12 summer and a southern position during winter in front of Morocco, while the Iceland
13 low intensifies leaving room for the eastward passage of low pressure systems, that
14 introduce a strong variability in the circulation along the WIM during wintertime.

15 These atmospheric regimes are the reason why in late spring and summer there is
16 coastal upwelling, that is, deep, cool, nutrient-rich waters upwell and are advected
17 equatorward, resulting in an offshore displacement of surface warm waters and
18 phytoplankton blooms (see review by Relvas et al., 2007). The thermohaline structure
19 of the water masses, consisting in large-scale temperature and salinity meridional
20 gradients, (Peliz et al., 2003) together with wind forcing, generate a system of
21 poleward currents along the Atlantic Iberian Peninsula observed mainly during
22 autumn and winter, as described in the classic references of Frouin et al. (1990) and
23 Haynes and Barton (1990) for the western Iberian Peninsula, and continuing along the
24 Bay of Biscay as described by Pingree and Le Cann (1990). This northward
25 circulation is typical of midlatitude eastern continental margins (Neshyba, 1986). A

1 consensual denomination for this flow is the Iberian Poleward Current (IPC). The
2 classic view of the IPC presents it as an upper-slope baroclinic current, associated
3 with the downwelling of the isopycnal field in a width of about 40 km, and positive
4 temperature and salinity anomalies from the surface down to typically 400 m. It was
5 suggested that poleward flows are not a winter phenomenon only, but are present
6 throughout the year (see reviews by Peliz et al., 2005 and Relvas et al., 2007).

7 Below the surface waters, the Mediterranean Water (MW) is of central importance for
8 the circulation in the region. The MW enters the domain through the Strait of
9 Gibraltar and flows at depth along the Gulf of Cadiz and the WIM slope. The
10 resulting intermediate-level current is called Mediterranean Undercurrent (MU) in the
11 Gulf of Cadiz (Ambar and Howe, 1979) and Mediterranean Water Vein along the
12 WIM (Daniault et al., 1994); it should be noted that the Mediterranean Water Vein is
13 not the only pathway by which MW reaches the Atlantic. Upon exiting the 300 m
14 deep Strait of Gibraltar, the dense MW flows downslope within the Gulf of Cadiz
15 until it reaches neutral buoyancy; afterwards, along the WIM, it flows at mid-depths
16 often in two cores, typically at 800 m (the upper core), where a temperature maximum
17 is reached, and at 1200 m (the lower core), the depth at which the salinity maximum is
18 found. The signature of the hydrographic properties of both cores decreases (notably
19 the upper core) in the poleward direction along the WIM. Associated with this flow,
20 there are mesoscale structures that are formed and which are responsible for a relevant
21 part of this water mass properties transport: the Mediterranean eddies (Meddies),
22 anticyclonic vortices with radii of several tens of kilometers, centered around 700 m
23 and 1300 m deep (Bower et al., 1995). Meddies propagate typically southwestwards
24 into the North Atlantic (Richardson et al., 2000).

1 The numerical study of the circulation of the WIM is a challenge for modelers, for
2 several reasons: i) the presence of a narrow shelf and a steep slope (up to 10%), as
3 well as the presence of numerous canyons and promontories, requiring high resolution
4 to properly resolve the shelf/slope circulation in presence of realistic stratification ; ii)
5 the presence of a unique oceanographic feature, which is the MU, generated at the
6 Strait of Gibraltar and spreading into the NE Atlantic giving rise to the large scale
7 MW anomaly; iii) the influence of the open ocean circulation (the Azores Current
8 system and meridional pressure gradients) on the coastal transition zone, that forces
9 the circulation in this region.

10 In this way, there is a need for high-resolution studies that may deal with the factors
11 mentioned above, with special attention to the different scales that influence the
12 circulation of the WIM, ranging from the few kilometers of the Strait of Gibraltar
13 dynamics, to the hundreds to thousands kilometers scale of the Azores Current
14 system. These features compelled us to perform a multi-scale simulation based on
15 nesting techniques, as detailed below.

16 In what concerns numerical modeling regional studies at the full scale of the region,
17 some previous works contributed to the study of this particular circulation. Some of
18 them were carried out at a low resolution of about 9 km (Stevens et al., 2000; Coelho
19 et al., 2002) and focused on spring and winter circulation; others were idealized
20 studies at a high resolution (Røed and Shi, 1999; Peliz et al., 2003). Batteen et al.
21 (2000, 2007) carried out sensitivity studies in order to assess the different mechanisms
22 controlling the circulation in the region. The study by Serra et al. (2010) resolves the
23 circulation in this region with realistic forcing and high resolution, focusing on the
24 study of the influence of the MW vein on the surface circulation, with emphasis on
25 the mesoscale phenomena. On the other hand, from the point of view of operational

1 oceanography of the Western Iberian Margin, Mateus et al. (2012) focused on the
2 application of an operational configuration ($\sim 1/16^\circ$ resolution) to discuss its potential
3 for products and services for scientific and coastal management activities.
4 In the neighborhood of the region, and also at a high resolution, Friocourt et al. (2007)
5 and Peliz et al. (2007) studied the Bay of Biscay and the Gulf of Cadiz circulation,
6 respectively.
7 Although the three major distinctive phenomena of the WIM, referred to above:
8 upwelling, poleward flow and Mediterranean outflow, have an independent dynamical
9 origin, they are intimately linked, and should be studied in an integrated manner. This
10 is the main challenge for the present work.
11 The present configuration includes the WIM region, the Western Bay of Biscay and
12 the Gulf of Cadiz. It is intended to study the equilibrium hydrography and circulation
13 in the surface layers, the underlying Central Waters, and the intermediate
14 Mediterranean Waters.
15 In order to do that, numerical simulations for the WIM region forced by
16 climatological heat and momentum fluxes were set up. In the present work we intend
17 to discuss the mean circulation that arises from the intrinsic variability of the model in
18 the absence of any external interannual atmospheric or oceanic forcing.
19 Using climatological forcings the model results achieve a realistic circulation,
20 reflected by a mesoscale field (not discussed in this work), and a mean field, which is
21 described and discussed in the light of the observations. The results discussed in this
22 work intend to contribute to the understanding of the IPC, not only during winter as is
23 classically studied, but also in the summertime, when the dominant poleward flow on
24 the outer slope coexists with equatorward flow, in agreement with the observations. It
25 is also intended to discuss the generation of vorticity structures of the mean flow

1 along the WIM, and its relation to the topography, and the large scale circulation. The
2 links between the surface and the intermediate circulation are also discussed, as is the
3 resulting circulation in the hypothetical case of suppression of the MW overflow.
4 The outline of this work is as follows: in section 2, we describe the methodology and
5 the ocean model configurations set up for this study; in section 3 we present a
6 comparison of model results with observational data and we characterize the mean
7 annual and seasonal hydrography and circulation of the WIM region; in section 4 we
8 study the influence of the MW in the WIM circulation; and in section 5 we discuss the
9 results.

10

11 2. Methods and Data

12

13 2.1 Ocean Model

14

15 A high-resolution nested configuration of the Regional Ocean Modeling System
16 (ROMS) (Shchepetkin and McWilliams, 2003) is used to simulate the seasonal
17 dynamics of the Iberian System. ROMS is a split-explicit, free-surface, topography
18 following coordinate model, designed to resolve regional problems (Shchepetkin and
19 McWilliams, 2005). ROMS solves the incompressible primitive equations based on
20 the Boussinesq and hydrostatic approximations, and is coupled with
21 advection/diffusion schemes for potential temperature and salinity as well as a
22 nonlinear equation of state. The advection scheme is based on the work done by
23 Marchesiello et al. (2009), in order to reduce spurious diapycnal mixing in sigma-
24 coordinate models characteristic of higher-order diffusive advection schemes. This
25 scheme involves the split of advection and diffusion, as a biharmonic operator.

1 Vertical mixing consists in the KPP (K-profile parameterization) scheme (Large et al.,
2 1994).

3

4 2.1.1 Model Configuration

5

6 a. Large domain

7 The strategy to manage a large range of scales consists in the implementation of
8 a two-domain approach, as shown in Figure 1. A large-scale first domain (FD) is
9 run independently in order to provide initial and boundary conditions to our
10 second domain (denoted SD hereafter) through an offline nesting. The first
11 domain horizontal resolution is $1/10^\circ$ (~ 9.5 km in longitude), and the main aim
12 for this domain is to resolve the large-scale circulation features such as the
13 Azores Current, and their interaction with the Atlantic margin of the Iberian
14 Peninsula.

15 For this domain, 30 sigma vertical levels are used, with a stretching factor of $\theta_s=7$
16 and $\theta_b=0$ to conserve a good near-surface resolution over the entire domain. The
17 bathymetry is based on ETOPO1 (Amante and Eakins, 2009), with corrections
18 near the slope and a smoothing filter to fulfill the $r=\Delta h/2h$ criterion (Haidvogel
19 and Beckmann, 1999), $r<0.2$.

20 The World Ocean Atlas 2009 (WOA-2009) climatology is used as the initial value
21 for the temperature (Locarnini et al., 2010) and salinity (Antonov et al., 2010)
22 fields, and also to recycle these fields along the nudging bands, providing open
23 boundary conditions. Surface fluxes are obtained from the Comprehensive
24 Ocean-Atmosphere dataset (COADS) (da Silva et al., 1994), interpolated to the
25 grid with the Roms_tools (Penven et al., 2008) package. This climatology derived

1 from the NOAA global data retrieval program (<http://icoads.noaa.gov>) and
2 consist in one value per month for each parameter, computed for the time period
3 1945-1989. That is, COADS has no interannual variability, only seasonal. Its
4 spatial resolution is of 0.5 degrees for all parameters except for sea surface
5 salinity, for which it is of 1 degree. Initial velocities are zero, and monthly
6 geostrophic velocities (with level of reference 1200 m) and Ekman velocities are
7 computed from the climatology and applied along the open boundaries.

8 The Mediterranean undercurrent is introduced by nudging, applied to the Gulf of
9 Cadiz and Western Iberia as described in Peliz et al. (2007), in order to restore
10 the hydrographic properties (T and S) of the Mediterranean levels, so as to
11 generate realistic Mediterranean water hydrographic properties.

12 The open boundary conditions were based on Marchesiello et al. (2001). These
13 conditions are suitable for long time integration, and are characterized by inflow
14 (outflow) nudging timescales of 1 (360) days for tracers, and 10 (360) days for
15 momentum respectively. Sponge layers 120 km wide were applied along the
16 edges, with a lateral viscosity coefficient ranging from 600 m²/s at the boundary
17 to zero in the interior. Explicit viscosity and diffusivity is null, and a linear drag
18 formulation with coefficient $r = 3 \times 10^{-4} \text{ m.s}^{-1}$ is applied at the bottom. The
19 objective of the large grid is to achieve a good representation of the Azores
20 Current transport and the connection of the Azores Current with the Gulf of
21 Cadiz where the western and southern boundary of the high-resolution grid is
22 located.

23

24 b. High-resolution domain

1 The target domain, SD, (second domain, see Figure 1), has a horizontal resolution
2 of $1/27^\circ$, (~ 3 km in longitude), and includes the Gulf of Cadiz, the WIM, and part
3 of the western Bay of Biscay, extending for ~ 1300 km in the meridional
4 direction, from 34°N to 46°N . In the zonal direction, the domain extends from the
5 Strait of Gibraltar, located at 5.5°W to 12.5°W , representing a width of about
6 ~ 600 km.

7 Sixty sigma vertical levels with $\theta_s=4$ and $\theta_b=0$ are used to properly resolve the
8 Mediterranean undercurrent with enough near-bottom resolution. In this way,
9 the grid has $60 \times 188 \times 389$ cells.

10 The topography by Sibuet et al. (2004) was used, it has a resolution of ~ 1 km and
11 was smoothed in order to fulfill the same r-factor criterion (r-factor less than
12 0.2) of the large-scale domain.

13 The forcing for this high-resolution configuration domain is the same as the one
14 used for the large scale FD domain. The initialization and the boundary
15 conditions are obtained using year 5 from FD, with average data stored every 3
16 days. To test the sensitivity of the model results to this choice, an additional
17 experiment to the main simulation, was run using year 7 of FD simulation,
18 instead of the year 5, showing results similar to the main run ones described
19 below.

20 Also, like for the large-scale simulation, a nudging sponge layer is introduced.
21 Concerning the boundary conditions of the SD domain, radiation conditions of
22 Marchesiello et al. (2001) are also used. The nudging coefficients at the edges are
23 the same as the ones used for FD configuration. However, the sponge layers at
24 the edges are applied to a band of 40 km, with a lateral viscosity coefficient
25 ranging from $200 \text{ m}^2/\text{s}$ at the boundary to zero at the interior of the domain. A

1 quadratic drag coefficient of 5×10^{-3} is used for the bottom friction
2 parameterization. At the Strait of Gibraltar, at the southeastern boundary, the
3 water exchange with the Mediterranean basin is explicitly represented in the
4 domain, through a fixed boundary (as in Peliz et al., 2007): it consists in the
5 imposition of vertical profiles of temperature, salinity and zonal velocity at the
6 boundary located at the Strait. This condition is designed to setup a two-layer
7 system consisting in a transport of 0.8 Sv of Atlantic water leaving the domain
8 through the surface layer, and 0.7 Sv of Mediterranean water entering the
9 domain through the lower layer. No attempt to introduce seasonality or
10 interannual variability in these fluxes was done. The process of entrainment of
11 Atlantic Central Waters with the MU is also parameterized by increasing the
12 viscosity and diffusivity coefficients in a region in which the MW is strongly
13 mixed with the overlying Atlantic waters, until the MW vein forms along the
14 northern slope of the Gulf of Cadiz.

15 The inflow of freshwater in the ocean, originating from the main rivers of the
16 region (the most important rivers are identified in Figure 1), is included. For the
17 Portuguese rivers, climatological values were provided by INAG (Water Institute
18 of Portugal); the information on the Spanish rivers was given by Barja and
19 Lestegás (1992).

20 The spin-up time for this domain, during which the Kinetic Energy stabilizes, is 5
21 years, as the adjustment time of the MU along the western and northern Iberian
22 Margin is quite a slow process. Thus, the model is run for 14 years, and the
23 analysis presented here is done for years 10 to 14, in order to show the seasonal
24 and interannual variability during that period.

25

c. Solution without Mediterranean Water

A sensitivity experiment is performed in order to study the resulting circulation without the MW spreading through the Strait of Gibraltar. For that, the large scale domain FD is run for 15 years, with initial conditions from the World Ocean Atlas 2009. The condition to maintain MW outflow was relaxed by turning off the nudging condition applied at the Gulf of Cadiz, until the effect of the MW salinity and temperature anomalies disappears as much as possible by diffusion. Boundary conditions for SD are built from the results of this FD configuration, with a closed boundary condition at the Strait of Gibraltar. SD is then run following the same methodology described in the previous section for 14 years.

2.2 Climatology and Satellite Products

In the following section, model output is compared to climatological data. Two datasets were chosen to carry out this comparison: the recent climatological GHER atlas for the Northeast Atlantic (Troupin et al., 2010) (hereafter GHER-NEA climatology) and satellite products. The GHER-NEA climatology was chosen for its high resolution (0.1 degree), which properly resolves the slope and shelf regions, unlike most available climatologies, and both temperature and salinity are compared to model results. For the sea surface temperature (SST), the comparison is carried out with data from the Advanced Very High Resolution Radiometer (AVHRR) of the National Oceanic and Atmospheric Administration (NOAA). The data was extracted from the EUMETSAT Ocean & Sea Ice Satellite Application Facility (OSI-SAF) (www.osi-saf.org) and was made available by CERSAT (IFREMER, France). The product has an approximate resolution of 2 km.

1 The final data consists in 7-year averages, corresponding to years 2002 to 2008,
2 for January and July, of the night satellite sweep (hereafter satellite climatology).

3

4 3. Results

5

6 The first part of this section is dedicated to introducing the main features of the
7 hydrography of the region, assessing the reliability of the model in resolving the
8 hydrography of WIM and its seasonal variability, and comparing model output
9 for SD with the climatological data. This comparison is carried out by means of
10 surface horizontal field maps, θ - S diagrams for water mass characterization and
11 vertical sections of hydrographic fields along different latitudes. The second part
12 of this section focuses on studying the mean seasonal circulation of the region
13 and the vertical structure of the flow from model output, through velocity fields
14 and the analysis of meridional transport. A second solution without MW as
15 described above is also analyzed at the end of the section.

16

17 3.1 First domain model configuration.

18

19 The FD domain was designed to solve the large scale circulation in order to
20 provide boundary conditions to the target domain SD. It is not intended to
21 discuss the resulting circulation from this domain, but to show an example of
22 how the large scale circulation is solved for this domain. For that purpose, a EKE
23 comparison between altimetry and FD output is done in Figure 2. Altimetry
24 weekly data was obtained from AVISO for the period October 1992 to July 2011.
25 The signature of the Azores Current is clear between 33°N and 36°N, and the

1 signal extends eastwards to $\sim 12^\circ\text{W}$ in both. There is also evidence of the North
2 Atlantic Current at the northwestern corner of the FD domain. In terms of
3 magnitude, EKE is more intense in ROMS for the Azores Current, and more
4 intense in AVISO for the North Atlantic Current. Nevertheless, we find that the
5 comparison is acceptable and we conclude that FD is reproducing correctly the
6 regional dynamics. Other diagnostics (not shown) were done in order to check
7 that the large scale hydrography and circulation of FD domain are suitable for
8 use as boundary conditions for the target domain, SD.

9

10 3.2 Hydrography and comparison to climatological data

11

12 3.2.1 Surface Fields

13

14 Figure 3 (4) presents the January (July) mean sea surface temperature (SST,
15 upper row) and salinity (SSS, lower row) of both SD and climatology, as well as
16 the difference field between the modeled and the climatological fields.

17

18 a. Winter

19 During winter (Figure 3), both SD and climatologies show the existence of a
20 meridional gradient of sea surface temperature and salinity. Moreover, a tongue
21 of saltier and warmer water centered at approximately 9.5°W - 10°W is associated
22 with the poleward advection of waters of subtropical origin, frequently referred
23 to in the literature as the Iberian Poleward Current (IPC, Álvarez-Salgado et al.,
24 2003). Although this tongue is visible in both SST and SSS, it is less evident in the
25 SST satellite field. Furthermore, the poleward penetration of the climatological

1 tongue reaches farther north than the SD one, causing a difference of about -0.1
2 in salinity between 39°N and 42°N.
3 Along the continental shelf region, north of the Estremadura Promontory (39°N),
4 we observe the presence of a band of buoyant fresher and colder waters, in the
5 SST and SSS climatologies, as well as in the SD averages. This band is associated
6 with the presence of winter river inflow along the western coast, mainly
7 contributed to by Tagus, Mondego, Douro, Minho and the Galician Rias rivers
8 (see Figure 1). This fresh water signature remains trapped on the continental
9 shelf, generating the Western Iberian Buoyant Plume (WIBP) (Peliz et al., 2002,
10 Otero et al. 2008).

11

12 b. Summer

13 During summer (Figure 4), the most important phenomenon observed is coastal
14 upwelling induced by the predominant equatorward winds along the WIM,
15 which result in the presence of a coastal band of cold and fresh waters.

16 The main patterns of the July average temperature and salinity for the offshore
17 region are well reproduced in SD, resulting in small temperature and salinity
18 differences (Figure 4 c, f). However, SD tends to overestimate the intensity of the
19 upwelling phenomenon, and therefore exhibits a coastal water band that is
20 colder than the satellite climatology. As discussed by Veitch et al. (2010), and
21 references herein, the climatological winds do not account for the drop-off of the
22 wind stress in the neighborhood of the coast.

23 There are also regions in which SD, on the contrary, overestimates the surface
24 temperature. Along the Galician coast, north of Cape Finisterre, unlike SD results,

1 filaments like the Finisterre filament (43°N), are frequently observed by satellite
2 in this area (Torres et al., 2003).

3 Other regions where SD temperatures are warmer than observed are the
4 northern Gulf of Cadiz (Portuguese southern coast) and the northern coast of
5 Morocco. These regions are influenced by gap winds, that consist in strong
6 easterly winds associated to pressure differences between both sides of the
7 strait of Gibraltar (Peliz et al., 2009), not represented in the climatological winds.
8 The summer SSS field (Figure 4d) is characterized by the upwelling of Eastern
9 North Atlantic Central Waters of northern origin (ENACW_{sp} , see section 3.1.2 for
10 details), which is advected southward. This upwelled water has a typical salinity
11 of about 35.8. Overlying the upwelled Central Water, there is a low salinity
12 surface water (lower than 35.7) observed in the GHER-NEA climatology (Figure
13 4e), which is associated with the remnant of the WIBP advected offshore by
14 Ekman transport, giving rise to a low salinity plume at the surface.

15

16 3.2.2 Water Masses

17

18 The main water masses of the upper layers observed in the region of study along
19 the WIM originate from the subpolar or the subtropical branches of the Eastern
20 North Atlantic Central Waters, ENACW_{sp} and ENACW_{st} , respectively, as defined
21 by Ríos et al. (1992). Below the Central Waters lies the Mediterranean Water
22 (MW) that flows through the Gulf of Cadiz and the entire basin with a typical
23 tongue structure, reaching the Bay of Biscay as described in the literature and as
24 observed in the available climatologies (WOA-2009; Reynaud et al., 1998; Iorga
25 and Lozier, 1999; Troupin et al., 2010). Below 1500 m, there is the signature of

1 the Labrador Sea Water, described by Paillet et al. (1998). Although its
2 hydrographic properties are reasonably well represented in SD (not shown), this
3 water mass, as well as the deeper layer North Atlantic Deep Water, is not
4 discussed in this work.

5 To compare the SD water mass properties to the climatological values, θ - S
6 diagrams were elaborated for two boxes (1° latitude x 1° longitude) centered at
7 10°W and 38°N and 42°N (see Figure 1 for location). These diagrams are
8 displayed in Figure 5. For each box, monthly (January, April, July and October)
9 mean profiles of potential temperature and salinity were calculated for years 10
10 to 14 of the simulation (one dashed line per year, in order to show the
11 interannual variability), and are superimposed to the GHER-NEA climatology
12 (solid line).

13 The seasonal evolution of the thermohaline properties of the Central and surface
14 waters is reproduced by SD, when compared to the climatological values. As for
15 the Central Water properties of both SD and climatology, by comparison with the
16 standard straight lines in the θ - S space for ENACW_{st} and ENACW_{sp} , it is clear that
17 the northern (southern) region is more influenced by the presence of the
18 subpolar (subtropical) branch of the ENACW.

19 For the southern region, in January (Figure 5a), the Central Waters have their
20 thermohaline properties superimposed with the ENACW_{st} line, typical of the
21 waters that give origin to the IPC. Below the surface, salinity decreases linearly
22 to a minimum of 35.7, and increases downwards because of the influence of the
23 underlying MW mass. In April (Figure 5b), a decrease of the surface salinity
24 maximum is observed, associated with the start of the upwelling season. During
25 summer (Figure 5c), the seasonal heating modifies the θ - S structure at the

1 surface, increasing the temperature of the surface layer. Furthermore,
2 interannual variability of the temperature in the surface layers is observed. The
3 MW mass is reproduced in SD, ranging from 36.2 to 36.35, which includes the
4 climatological value. In October (Figure 5d), after the upwelling season, the
5 signature of saltier surface waters associated to the IPC is visible both in the SD
6 and the climatology, although SD underestimates the surface salinity by about -
7 0.1 units.

8 Regarding the northern region (centered at 42°N, 10°W, Figure 5 e-h), the
9 seasonal evolution of the θ -S field follows a pattern similar to the one described
10 for the southern region, the main differences being observed in the salinity field
11 at the surface levels, with lower maximum values of salinity both in winter and in
12 summer. At the MW levels the salinity signature decreases in the poleward
13 direction as expected.

14 Overall, the SD configuration not only follows the seasonal cycle of temperature
15 and salinity for the surface and central waters, but also reproduces the MW
16 hydrographic properties, although some negative biases are observed (as well as
17 in the MW core depths, as discussed later).

18

19 3.2.3 Vertical Structure

20

21 As a complementary analysis of the water masses in the two regions discussed
22 above, we now study the vertical distribution of the hydrographic properties,
23 and compare the modeled values with the GHER-NEA climatology. The aim is to
24 show and discuss the modeled temperature and salinity fields and to compare
25 them to climatological values across two latitudes, 42°N and 38°N,

1 representative of the northern and southern WIM, respectively. The averages for
2 SD were computed for January and July of simulation years 10 to 14, down to
3 1500 m in depth.

4

5 a. Winter Hydrography

6 At 42°N (Figure 6 a, c), the winter distribution of temperature and salinity shows
7 a general downward sloping of the isolines in the upper slope region (down to
8 about 400 m), between 9.5°W and 11°W. This downward tendency is associated
9 with the presence of the IPC during wintertime, with salinities higher than 35.9
10 and temperatures around 14.5°C (Figure 6c). The SD IPC (Figure 6a) shows
11 evidence of this poleward advection at the surface; however, there is a bias of
12 about -1°C in temperature and about -0.1 in salinity. On the other hand, the
13 salinity minimum of ENACW_{sp} of 35.65 (see Figure 5e) is reached at depths
14 around 400 m, shallower than the climatological minimum (450 m). Below that
15 depth lies the MW and thus the salinity field values increase accordingly to 36
16 near the slope at an approximate depth of 900 m, which is -0.1 fresher and 100 m
17 shallower than the climatological value.

18 For the southern latitude, 38°N (Figure 6 b, d), the surface layers are
19 characterized by the presence of saltier and warmer waters when compared to
20 the northern section. In January, for the surface and central water range, the
21 hydrographic properties reflect the straight line observed in the θ - S diagrams
22 (Figure 5a). Concerning the MW, below the ENACW, the SD temperature
23 maximum ($\sim 12^\circ\text{C}$) associated with the upper core of MW does not reach as far
24 offshore as the climatological one (extending to 10.5°W whereas in SD the
25 isotherm is close to the slope), but the depth at which they are found (800 m) is

1 the same. On the other hand, the lower core SD MW salinity maximum, 36.2, is
2 underestimated (-0.1 bias in S), as it is the depth at which this maximum is found
3 (about 1000 m in SD and 1200 m in the climatology). Below the MW, the
4 transition to LSW in SD is located at depths lower than the climatological value.

5 6 b. Summer Hydrography

7 Like the winter situation, the summer vertical structure is analyzed for zonal
8 sections at 42°N and 38°N. At the northern section (Figure 7 a, c) the summer
9 hydrography is well characterized by the presence of coastal upwelling, which
10 affects clearly the near-surface temperature field distribution on the continental
11 shelf. The isotherms warmer than 14°C outcrop on the shelf, generating a front in
12 the numerical configuration. Note that the GHER-NEA climatology shows
13 evidence of upwelling, although the frontal structure is smoothed, as expected
14 from an observational climatological dataset.

15 The upwelling of the isothermal field extends down to the 13°C isotherm at
16 approximately 200 m in both SD and climatology. In the surface layers, summer
17 salinity is lower than during winter, because the signature of ENACW_{st} is not
18 observed (Figure 5g) as the upwelled waters have a Northern origin. A minimum
19 of salinity is nevertheless observed, like in winter, with the 35.6 isohaline located
20 at approximately 400 m depth.

21 Concerning the MW distribution in July, at 42°N there are no significant
22 differences with respect to wintertime in the climatology: the same salinity
23 maximum is found at approximately 1000 m, enclosed by the 35.9 isohalines
24 between 700-800 m and 1200-1300 m. In SD, the MW vein seems to be squeezed
25 and the temperature field is shallower by about 100 m with respect to the winter

1 equilibrium depth, as observed by Garcia-Lafuente et al. (2008). These authors
2 related the shallowing of the MW vein to the mesoscale field associated with the
3 MW vein, which may originate from the variability of the wind field.

4 The southern section, 38°N (Figure 7b, d), during summer is also affected by
5 upwelling. In this case the outcropped isotherm is 15°C in SD against 17.5°C in
6 the climatology. At the levels of the MW vein, the main difference between SD
7 and climatology is the 200 m depth difference in the location of the salinity
8 maximum, whereas SD reaches the same climatological maximum of 36.3.

9

10 c. Winter vertical meridional section.

11 The surface layers meridional distribution of temperature and salinity fields
12 along the WIM at 11.5°W (Figure 8) shows evidence of a large scale gradient,
13 with southern warmer and saltier waters of subtropical origins, and northern
14 colder and fresher waters of subpolar origins. A frontal region, frequently
15 observed between 38°N and 40°N, and denoted by Western Iberia Winter Front
16 (WIWF) (Peliz et al., 2005), separates both regions.

17 This meridional density gradient is partly at the origin of the Iberian Poleward
18 current system (Peliz et al., 2003). A proper representation of the meridional
19 distribution of hydrographic fields is needed to obtain a realistic poleward
20 current system along the WIM.

21 A comparison of ROMS monthly averages of temperature and salinity and the
22 GHER-NEA climatological fields for January was done (Figure 8).

23 We observe similar meridional gradients of temperature and salinity in the
24 upper levels (0-200 m), although ROMS presents lower salinities north of 43°N
25 and south of 37° N (roughly differences of -0.1) and lower temperatures of about

1 -0.5°C to the north, reaching 1°C to the south. At depth, both the ENACW and the
2 MW signatures are clearly observed, the former at 400-600 m with typical
3 salinities of 35.6-35.7 and the latter centered at 1000-1200 m with maximum
4 salinities of 35.2-35.3 and a temperature of 11°C.

5 6 3.3 Mean Flow Structure

7
8 The main focus of this section is the analysis of the circulation in the WIM
9 resulting from the process of adjustment of the hydrographical fields. Horizontal
10 slices at 50 and 1000 m depth showing monthly means for years 10 to 14 of the
11 SD run of the salinity and velocity fields for January and July are displayed in
12 Figure 9. In addition, vertical slices of the alongshore circulation at 43°N are
13 displayed in Figure 10, also for January and July monthly means.

14 15 a. Winter Circulation

16 One of the main dynamical features at the western Gulf of Cadiz is the presence
17 of a permanent cyclonic circulation (C1 in Figure 9b), which is part of the
18 structure usually referred to a topographic β -plume circulation (Kida et al.,
19 2008). Lamas et al. (2010) provided observational evidence of the structure. This
20 cyclonic vortex is stronger at the surface, but has a clear signature at depth as
21 well (Figure 9 a, b).

22 The circulation in the WIM is affected by the entrance of the MW vein at the
23 gateway between Cape St Vincent and the Gorringe Bank; furthermore, along its
24 path around the WIM, the current sometimes becomes unstable and separates

1 from the slope at different locations, generating recurrent anticyclonic structures
2 (Meddies) (Zenk and Armi, 1990).

3 As consequence of the spreading process of the MW vein, when the flow turns
4 into the Tagus Basin, the resulting mean flow is observed to generate mean
5 vorticity structures, two of them (anticyclonic and cyclonic) trapped on the
6 slope, denoted by A1 and C2 in Figure 9b, and an anticyclonic mean structure
7 offshore, denoted by A2. The resulting mean circulation is rather complex in the
8 Tagus basin, and on its northern side, near the Estremadura promontory in
9 progress in the poleward direction, giving rise to a partial detachment and an
10 anticyclonic structure, denoted A3, which extends from the MW levels up to the
11 surface (Figure 9a).

12 Evidences in the literature of this conspicuous negative vorticity in this region
13 are discussed by Daniault et al. (1994) and Mazé et al. (1997). On the other hand,
14 a region of anticyclonic vorticity is frequently observed near 40°N in the SST
15 field, separating the southern waters (warmer and saltier) that progress in the
16 poleward direction, from the northern waters (fresher and colder). Also, a
17 detailed discussion of this anticyclonic region in the framework of the dynamics
18 of the IPC is presented by Peliz et al. (2003, 2005).

19 On the northern side of anticyclone A3, a flow bifurcation is observed (Figure
20 9b). Part of the flow veers cyclonically and becomes part of a northwestward
21 flow that leaves the domain at 42°N, 12°W south of the Galicia Bank; the other
22 part of the flow reattaches to the slope in the poleward direction, along Cape
23 Finisterre at 43°N. The vertical structure of this poleward flow at this latitude
24 can be seen in Figure 10a, consisting in a poleward flow, associated with a
25 downward slope of the isopycnal field centered at about 10° W and extending in

1 the vertical direction from the levels of MW up to the surface, while offshore of
2 10° W a weak equatorward flow is dominant.

3 The 43°N section in the work of Mazé et al. (1997) shows similar circulation,
4 with poleward tendency maximum at ~9.75°W, and equatorward flow from
5 10.5°W to 12°W, which is attributed in part to the recirculation around the
6 Galicia Bank.

7 The circulation at 50 m (Figure 9a) shows patterns similar to those at 1000 m,
8 showing coupling between the MW and the surface layers as further discussed
9 below.

10

11 b. Summer Circulation

12 The summer circulation at the MW levels (Figure 9d) presents a similar general
13 behavior to that of winter, discussed above. The inflow in the Tagus Basin occurs
14 closer to the slope than in January.

15 The anticyclonic structure centered at 40°N, 11°W exhibits a weaker signature
16 than the one observed in winter, which means a weaker tendency to detach from
17 the northern flank of the Estremadura Promontory, and hence stronger slope-
18 trapped flow at the levels of the MW.

19 One of the most striking features in our modeled mean circulation for summer is
20 the existence of conspicuous poleward flow. This issue will be further discussed
21 below. Concerning the surface layers circulation (Figure 9c), a band of
22 equatorward flow along the continental shelf/upper slope is observed associated
23 with the upwelling front, with particular notice of the velocities between Cape
24 Ortegal and Cape Finisterre. This flow is clearly visible in the vertical section at
25 43°N, showing the structure of the upwelling front and an alongshore jet

1 penetrating to a depth of about 400m surrounded offshore by the poleward flow
2 centered at about 10°W, and again equatorward weak flow offshore in a similar
3 way to the winter situation (Figure 10a)

4

5 c. Meridional Transport

6 In order to get some insight about the seasonal and interannual variability of the
7 circulation, series of alongshore transport were computed across three zonal
8 sections at latitudes 43°N, 40°N and 37.5°N, and integrated in the upper 1500 m
9 depth range.

10 Each zonal section is divided into 3 sub-sections (hereafter called boxes for
11 simplicity): shelf/upper slope (from the coast to the 400 m isobath), lower slope
12 and offshore box. These boxes will hereafter be designated as onshore, central
13 and offshore boxes, respectively. The width of the boxes is defined individually
14 for each latitude, because the location of both the upper and lower slope
15 circulation varies with latitude. The border of the onshore box was designed to
16 accommodate the shelf/upper slope circulation, and was therefore defined by
17 setting the depth at the 400 m isobath. The separation between the central and
18 the offshore boxes was specified by defining the central box, as the area of
19 prevailing poleward flow.

20 Thus, monthly means of meridional net transport, with positive (negative) values
21 corresponding to poleward (equatorward) flow, are shown for 43°N (Figure 11a-
22 c), 40°N (Figure 11 e-g) and 37.5°N (Figure 11 i-k).

23 In the onshore boxes (Figure 11 c, g, k), the main feature observed is the
24 equatorward flow associated with upwelling, which presents one main peak
25 centered in July-August and occasionally one or several minor peaks during

1 spring. In wintertime, a poleward flow associated with the inshore intrusion of
2 the IPC is observed on the upper slope. The equatorward summer transport peak
3 is typically 0.5 Sv, independent of the latitude, but the spring peak varies from
4 less than 0.1 Sv to about 0.3 Sv. Although a clear seasonal cycle of the circulation
5 is seen for both upwelling and poleward flow, there is also some interannual
6 variability in the structure of the peaks associated with the upwelling
7 equatorward transport, both in width and intensity. Regarding the seasonal
8 cycle, some differences can be observed in the equatorward circulation at
9 different latitudes. In the northern and central sections (43°N and 40°N ,
10 respectively – Figure 11c, g), the equatorward flow starts early in the year
11 (February) and lingers on until October, whereas in southern Portugal (37.5°N –
12 Figure 11k) a weak and variable circulation is observed during springtime
13 (February to May), and the clear signal of upwelling appears in June-July.
14 Furthermore, poleward flow along the shelf/upper slope is observed every year
15 for the 43°N and 40°N sections, with typical values of 0.2 to 0.3 Sv, starting at the
16 beginning of autumn and with peaks in December, occasionally with secondary
17 peaks in January. The southern region does not show clear evidences of
18 poleward flow on the upper slope/shelf.

19 With respect to the central (Figure 11b, f, j) and offshore (Figure 11a, e, i) boxes,
20 the transport patterns are not as clearly defined as in the case of the onshore
21 box. For the northern and central latitudes (43°N and 40°N) in the central box
22 (Figure 11b, f), two peaks of poleward flow are observed throughout the year in
23 almost all simulation years: the first one during summertime (July to
24 September), presenting high transport values, from 3 to 10 Sv; the second peak
25 with smaller amplitudes, 2 to 7 Sv, is observed from November to January. Both

1 transport peaks show strong interannual variability. The first peak appears as an
2 offshore poleward flow co-existing with the inshore summer upwelling peak (cf.
3 Figure 11b with Figure 11c and Figure 11f with Figure 11g). This offshore
4 shifting of poleward flow during upwelling season has also been reported in
5 hydrographical cruises (Peliz et al., 2002, and Torres and Barton, 2007). The
6 second peak is associated with the winter IPC. Conversely, every year, recurrent
7 peaks of equatorward transport are observed during the late winter-spring
8 months. The southern region in the central box (Figure 11j) presents a complex
9 pattern, although the tendency for poleward flow with multiple peaks during
10 summer (shaded in Figure 11j), late autumn and January is still discernible. The
11 tendency for equatorward flow observed during late winter and spring is only
12 partially observed. Recall that this box is strongly influenced by the MW vein and
13 the corresponding shedding of meddies, which makes it more difficult to
14 interpret than the patterns described for the northern region.

15 The offshore region (Figure 11a, e, i) is the widest of the three boxes, and does
16 not present a clear seasonal pattern as in the central and onshore boxes. The
17 northern offshore box (Figure 11a) is characterized by a tendency for
18 equatorward flow, with typical values of less than 5 Sv, as previously discussed
19 (Figure 10a, b). This behavior is opposite to the one observed at the central box
20 at the same latitude (Figure 11b), which means that at 43°N the poleward flow is
21 preferentially attached to the slope (central box), which Figure 9b, d also shows.
22 At 40°N (Figure 11e), the offshore box is characterized by high values of
23 transport in both directions, resulting from the anticyclonic vorticity structure of
24 the mean circulation (denoted A3 in Figure 9b). However, the net overall
25 transport in this box (Figure 11e) is poleward during the summer months with

1 strong interannual variability, and on the other hand, presents also peaks of
2 poleward flow in February-March. This late winter poleward transport is
3 delayed by two to three months relative to the peaks observed in the central box
4 at 40°N (Figure 11f), which can be interpreted as the westward IPC migration.
5 On the other hand, the separation of the flow associated with A3 (Figure 9b)
6 occurring north of the Estremadura Promontory also explains the poleward
7 tendency in this offshore box.
8 The southern (37.5°N) offshore region (Figure 11i) is affected by a strong
9 variability, consequence of the spreading of the MW vein and its mesoscale
10 features. However, like for the central offshore box (Figure 11e) some features
11 can be observed, as peaks of poleward flow during summertime (shaded), and
12 during November-January for most of the years, but also equatorward peaks
13 between February and June every year.

14

15 3.4 Influence of the MW in the WIM circulation

16

17 One unique oceanographic feature in this region is the presence of the MW and
18 its associated current patterns. Comparing the obtained circulation from a model
19 configuration where the MW spreading through the Strait of Gibraltar was
20 removed, with the base experiment, it is possible to get some insights into the
21 influence of the MW spreading on the circulation of the region, which is closely
22 linked to the density distribution at depth along WIM upper slope and to the
23 dynamics of Atlantic surface and Central Waters and associated mesoscale
24 activity.

1 The January and July mean circulation and salinity fields at 50 m and 1000 m in
2 depth are displayed in Figure 12, like for Figure 7. Zonal sections of meridional
3 velocity were also computed along latitudes 43°N for January and July, like for
4 Figure 10, and are shown in Figure 13.

5

6 a. Winter circulation

7 The 1000 m current fields (Figure 12b) present lower velocities in this No
8 Mediterranean Water case (hereafter NMW) and there are less vorticity
9 structures in the mean flow when compared to the base case. Poleward flow
10 remains dominant at the surface and extends to ~1000 m depth. The mean
11 cyclonic recirculation cell signature to the south nearly disappears, as it was
12 associated to the MW outflow. In the Tagus Basin there are no longer clear
13 evidences of a region of vorticity structures, as it was in the base case (A1, A2, C2
14 in Figure 9b). North of the Estremadura Promontory, anticyclonic vorticity of the
15 mean flow can still be observed, but is much weaker than in the base case, with
16 velocities decreasing from 10-15 cm.s⁻¹ throughout the water column to 2-4 cm.s⁻¹,
17 barely reaching the surface (Figure 13a). Concerning the salinity field, weak
18 horizontal gradients are observed at 1000 m depth, as expected due to the
19 absence of MW.

20 Near the surface (Figure 12a), the patterns of the mean velocity field are similar
21 to those at 1000 m depth with increased intensity. Poleward current is the
22 prevailing pattern, with a continuous mean flow from the Gulf of Cadiz
23 northwards, around Cape St Vincent and along the WIM. The surface poleward
24 flow proceeds along northwestern Iberia with evidence of two branches: one
25 narrow branch over the continental shelf (not visible in the base experiment),

1 and another wider branch centered at about 10°W , both along the slope and
2 extending eastwards into the Bay of Biscay with stronger intensity than in the
3 base case. This double flow is also visible in the zonal sections (Figure 13a).
4 A computation similar to the one depicted in Figure 11 was carried out (not
5 shown) and it further confirms the prevailing poleward flow in the continental
6 shelf/upper slope (onshore boxes). Additionally, the mean annual balance is
7 positive, that is, it surpasses the equatorward transport of upwelling.
8 Regarding the density distribution in depth near the upper slope (Figures 10 and
9 13), there is a greater tendency for downward slope in the case of NMW than in
10 the base experiment, which explains the strengthening of the poleward flow near
11 the shelf. This increase in downward slope was also observed in other regions
12 along the WIM (not shown). This stronger inshore poleward flow is also visible
13 in the salinity field at 50 m depth (Figure 12a), where the penetration of the 36
14 isohaline (marked in a thick contour) reaches farther north than in the base case,
15 which presents nearly-zonal isohalines.

16

17 b. Summer circulation

18 For July, the mean circulation patterns are similar to those of January, for both
19 50 and 1000 m depths (Figure 12 c-d), that is, there is a general weakening of the
20 recirculation regions and a poleward flow tendency along the slope of WIM, as
21 well as a farther northward penetration of the 36 isohaline. As for the meridional
22 velocity zonal sections (Figure 13b), we find that there is still the strong inshore
23 poleward flow ramification observed in winter, which restricts upwelling to a
24 thinner band on the continental shelf (less than 0.5° from the coast) and

1 therefore also at depth, allowing the associated equatorward flow to spread only
2 for the first 100-150 m.
3 In this way, the main effect of the absence of the MW vein on the mean
4 circulation of WIM is the suppression or strong attenuation of a significant part
5 of the recirculating flow associated with the mesoscale features characteristic of
6 the MW spreading, such as shedding of meddies southwest of the Iberian
7 Peninsula, whose signature in the mean flow are the structures depicted in
8 Figure 9b, β -plume cyclonic circulation in the Gulf of Cadiz, anticyclonic vorticity
9 generation north of the Estremadura Promontory, and meandering along
10 northwestern Iberia. As a consequence, the poleward flow seems to circulate
11 more trapped to the slope, between 10°W and 9.5°W.

12

13 4. Discussion

14

15 The circulation of the Western Iberian Margin was studied by means of a high-
16 resolution primitive-equation numerical model. It was intended to model the
17 seasonal cycle of the hydrographic fields, and to study the associated circulation,
18 resulting from the interaction of the three main phenomena: The upwelling, the
19 Iberian Poleward Current system and the Mediterranean water outflow.

20 The WIM ocean circulation is characterized by a complex oceanic system and it is
21 influenced by large-scale phenomena like the eastward extension of the Azores
22 Current, the structure of the meridional density gradient, or the less energetic
23 Portugal offshore equatorward current. It is also influenced by small-scale
24 localized phenomena like the exchange of waters through the Strait of Gibraltar,

1 which spreads along the NE Atlantic. The river inflow also has influence on the
2 circulation in the surface layers along the continental shelf, and offshore.
3 The main focus of this study is to integrate the above mentioned phenomena into
4 a numerical configuration, that takes into account the large-scale phenomena
5 and the local phenomena referred to above. The main objective is to study the
6 mean hydrography and flow with special emphasis on the shelf/slope mean
7 circulation and transports, and its seasonal evolution. The connection between
8 the intermediate circulation (at the MW levels) with the surface layers is
9 explored, and the prevalence of poleward slope flow along the summer.

10

11 a. Circulation Considerations

12 The forcing that we used lacks nonseasonal transient forcing and contains no
13 interannual or synoptic meteorological scale variability. It was intended to model
14 the hydrographic fields, and to reach a realistic circulation as a result of the
15 adjustment of the hydrographic fields. Concerning this resulting circulation,
16 many of the modeled features have their correspondence in the available
17 observations. The main pattern of the circulation consists in a poleward flow
18 along the slope during winter, which lingers on in summer. In addition to the
19 classic ideas related in the introduction, Torres and Barton (2006) described a
20 complementary view of the initial development of the flow along northern WIP;
21 they estimated a mean transport of 2 Sv from the surface down to 500 m and
22 postulated that the poleward flow could penetrate down to the levels of MW. The
23 authors hypothesize that a poleward flow at the MW levels would arise from
24 different dynamical reasons, but can form a continuous flow inducing
25 topographic guidance of the surface flow. This behavior is well reproduced by

1 the model results, suggesting that the poleward flow extends down to the levels
2 of the MW (Figure 12 a-b).

3 The winter flow patterns obtained by the SD configuration resemble the typical
4 description of poleward flow in the literature. The model results, as seen by the
5 analyzed boxes (Figure 11), also suggest interannual variability associated with
6 the intrinsic variability, and hence not resulting from external forcing.

7 However, this resulting interannual variability does not explain episodes of
8 strong alongshore current observed in particular years only (January 1990,
9 1996, 1998 as reported in García-Soto et al., 2002, and Peliz et al., 2005). This
10 means that factors other than climatological forcing or large-scale forcing, both
11 imposed seasonally but with no interannual variability, may explain the
12 variability and the years of extreme poleward flow, not described within our
13 simulations.

14 When considering the mean fields of hydrographic properties and alongshore
15 velocity fields for several years, we obtain a picture of a poleward flow extending
16 from the surface down to the levels of the MW, centered at about 10°W, with
17 weak vertical shear and typical velocities of 6 to 10 cm.s⁻¹ for winter and
18 summer, although with some weakening at the surface in this season (Figure 10).

19 This coupling between the MW and the Central Water layers was already
20 suggested by Mazé et al. (1997). Moreover, observations of Huthnance et al.
21 (2002) report maximum poleward flow at moorings near 42°N, 9.5°W in
22 September-October for all depths, a secondary peak in December-January for the
23 uppermost currentmeter, and a maximum of equatorward flow in February-
24 April. These features resemble the transport patterns in the central box for
25 section 43°N (Figure 11b). Likewise, persistent poleward flow at about 250 m

1 depth at a location near 42.3°N, 10.2°W is reported by Peliz et al. (2005) based
2 on datasets of the Instituto Español de Oceanografía (Alonso et al., 1995), where
3 the intensity of the poleward current was about 10 cm.s⁻¹ and no correlation
4 between winds and currents was found at this mooring. This tendency for
5 poleward flow at this longitude is comparable to the mean values presented in
6 Figure 10.

7 Concerning the issue related to the presence of poleward flow during the
8 upwelling season, Peliz et al. (2002) presented evidences of such behavior for
9 central Portugal. Torres and Barton (2007) studied the Galician shelf/slope
10 based on hydrography and ADCP measurements and also showed the
11 coexistence of offshore poleward flow during June 1997 at the start of the
12 upwelling season. The model results suggest that poleward flow is present along
13 the slope and coexist with the upwelling-associated flow along the shelf/upper
14 slope. This is visible not only in the mean alongshore velocities, but also in the
15 transport associated with the central (slope) box (Figure 11).

16

17 b. The influence of the MW flow on WIM

18 Concerning the experiment without explicit Mediterranean Water exchange at
19 the Strait of Gibraltar, the resulting circulation consists in a general tendency for
20 poleward currents in way similar to the SD experiment, which flows mainly
21 trapped on the slope. The tendency for poleward flow regardless of the presence
22 or absence of Mediterranean Water seems to be a robust result. However, the
23 MW vein and its mesoscale features have strong influence on the poleward flow
24 paths along the WIM, namely the enhancement of the separation north of Cape St
25 Vincent, the promotion of strong anticyclonic recirculation north of the

1 Estremadura Promontory and the separation at 41-42°N. The enhancement of
2 the slope-trapped poleward flow is a noteworthy result given that one would
3 expect the reverse by switching off the MW flow. There are a few possible
4 reasons for this effect in the absence of MW spreading: i) the absence of the
5 associated positive thermohaline anomaly of the MW at mid-depths forces the
6 density distribution at depth to re-arrange, so that the geostrophic adjustment at
7 the slope, seen as a downward inflection of isopycnals, takes place at upper
8 levels, as referred to above. As discussed in Huthnance (1984), geostrophic
9 adjustment at western coasts produces the so-called Eastern Boundary Currents,
10 which flow usually in subsurface in a direction opposite to that of the prevailing
11 winds; ii) the almost disappearance of the cyclonic recirculation in the Gulf of
12 Cadiz, the absence of detached anticyclonic structures throughout the WIM
13 (meddies), and the absence of instabilities (meanders) associated with the
14 absence of the MW vein, may allow a stronger slope-trapped poleward flow that
15 would otherwise spread offshore, and at once a more stable flow along the WIM
16 as it is observed in SD domain under NMW (No Mediterranean Water)
17 experiment. Consequently, lower levels of EKE were observed in the case of
18 NMW (not shown); iii) several authors (e.g. Jia, 2000; Kida et al., 2008; Volkov
19 and Fu, 2010) have hypothesized that there is a connection between the MW
20 outflow in the Gulf of Cadiz and the recirculation of the Azores Current
21 southwest of the Iberian Peninsula, and even the very existence of that current.
22 Therefore, the removal of the MW flow into the Atlantic may have implications at
23 a larger spatial scale, which would then produce changes on the structure and
24 intensity of the poleward flow. However, we cannot draw conclusions on this
25 matter since our target domain (SD) does not enclose the Azores Current region.

1 The results suggest that in order to be able to realistically model the WIM
2 circulation, a good representation of the MW spreading should be achieved,
3 including its associated mesoscale features that control the low frequency
4 circulation.

5

6 5. Conclusions

7

8 A numerical configuration was setup to resolve the ocean circulation at the
9 Western Iberian Margin, based on climatological forcing, in order to study the
10 resulting dynamical features of the surface layers down to the Mediterranean
11 Water levels.

12 As a first step, the hydrographic fields resulting from the equilibrium solution
13 were analyzed and compared to climatological values, and we obtained
14 reasonable results that allow exploring the associated circulation and studying of
15 the mean seasonal velocity fields and associated transports.

16 The resulting circulation presents a tendency for poleward flow along the slope
17 of the WIM, but also evidences of equatorward flow mostly in spring. Also, as
18 expected, wind-forced upwelling-driven equatorward flow is observed along the
19 continental shelf in summer, in coexistence with poleward flow offshore. The
20 main paths of poleward flow appear to be linked to the MW vein along the WIM.

21 A schematic view of the annual circulation at 1000 m (which extends its main
22 features up to the surface layers), showing the main paths of the mean
23 circulation, is proposed in Figure 14.

24 The MW vein enters through the Cape St Vincent-Gorringe Bank gateway and
25 flows in the northwestward direction, separating from the southwestern slope of

1 the Iberian Peninsula (near cape St Vincent) through an anticyclonic mean
2 vorticity region (A1). Part of the flow continues poleward along the slope and
3 part recirculates southward through a cyclonic mean vorticity region (C2) and
4 continues poleward offshore due to the presence of another anticyclonic
5 structure (A2). Both poleward branches, at the slope and offshore, give rise to
6 the A3 anticyclonic region north of the Estremadura Promontory, although there
7 is still flow along the slope, getting around the promontory. On the northern
8 flank of this mean vorticity region, results suggest a new separation: part of the
9 flow returns toward the coast and continues as the slope poleward flow, which is
10 linked to the so-called Iberian Poleward Current (IPC); the other part circulates
11 poleward in another offshore branch between 11°W and 12°W. The slope
12 poleward flow seems to separate south of 42°N and joins the offshore branch,
13 exiting afterwards the domain south of the Galicia Bank. The slope branch
14 continues to flow northward along Cape Finisterre, with evidences of a
15 separation at the northwestern tip of the Iberian Peninsula. Observational
16 evidences for many of the features of this scheme can be found in the literature,
17 as discussed above.

18

19

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21

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7 References

8

9 Alonso, J., F.F. de Castillejo, G. Diaz del Rio, D. Marcote, and G. Casas (1995),
10 Preliminary Results of Current Measurements in the North West of the Iberian
11 Peninsula, MORENA Sci. and Tech. Rep. no 15, Instituto Espanol de Oceanografia
12 C.O., La Coruna, Spain.

13

14 Álvarez-Salgado, X.A., F.G. Figueiras, F.F. Pérez, S. Groom, E. Nogueira, A.V.
15 Borges, L. Chou, C.G. Castro, G. Moncoiffé, A.F. Ríos, A.E.J. Miller, M. Frankignoulle,
16 G. Savidge, and R. Wollast (2003), The Portugal coastal counter current off NW
17 Spain: new insights on its biogeochemical variability, Prog. in Oceanog., 56, 2,
18 281-321, doi: 10.1016/S0079-6611(03)00007-7.

19

20 Amante, C., and B.W. Eakins (2009), ETOPO1 1 Arc-Minute Global Relief Model:
21 Procedures, Data Sources and Analysis. NOAA Tech. Memorandum NESDIS
22 NGDC-24, 19 pp.

23

24 Ambar, I., and M.R. Howe (1979), Observations of the Mediterranean outflow I –
25 Mixing in the Mediterranean outflow, Deep-Sea Res. Part A, 26, 5, 535-554.

1

2 Antonov, J. I., D. Seidov, T. P. Boyer, R. A. Locarnini, A. V. Mishonov, H. E. Garcia, O.
3 K. Baranova, M. M. Zweng, and D. R. Johnson, 2010. World Ocean Atlas 2009,
4 Volume 2: Salinity. S. Levitus, Ed. NOAA Atlas NESDIS 69, U.S. Government
5 Printing Office, Washington, D.C., 184 pp

6

7 Barja, F.J.R, and F.R. Lestegás (1992), Os Ríos Galegos – Morfoloxía e Réxime,
8 Consello da Cultura Galega, Ponencia de Patrimonio Natural, Santiago de
9 Compostela, Spain.

10

11 Batteen, M.L., J.R. Martinez, D.W. Bryan, and E.J. Buch (2000), A modeling study of
12 the coastal eastern boundary current system off Iberia and Morocco, J Geophys
13 Res-Oceans, 105 (C6), 14173-14195.

14

15 Batteen, M.L., A.S. Martinho, H.A. Miller, and J.L. McClean (2007), A process-
16 oriented modelling study of the coastal Canary and Iberian Current system, Oc.
17 Mod., 18, 1-36.

18

19 Bower, A.S., L. Armi, and I. Ambar (1995), Direct evidence of meddy formation off
20 the southwestern coast of Portugal, Deep-Sea Res. I, 42, 9, pp. 1621-1630.

21

22 Coelho, H.S., R.J.J. Neves, M. White, P.C. Leitão, and A.J. Santos (2002), A model for
23 ocean circulation on the Iberian coast, J. Mar. Sys., 32, 1-3, 153-179,
24 doi:10.1016/S0924-7963(02)00032-5.

25

1 da Silva, A.M., C.C. Young, and S. Levitus (1994), Atlas of Surface Marine Data
2 1994. NOAA Atlas NESDIS 10. U.S. Department of Commerce, NOAA, NESDIS.
3

4 Daniault, N., J.P. Mazé, and M. Arhan (1994), Circulation and Mixing of
5 Mediterranean Water West of the Iberian Peninsula, Deep-Sea Res. I, 41, 11-12,
6 1685-1714, doi:10.1016/0967-0637(94)90068-X.
7

8 Fiúza A.F.G. (1984), Hidrologia e dinâmica das águas costeiras de Portugal, PhD
9 thesis, University of Lisbon, 294 pp.
10

11 Friocourt Y., B. Levier, S. Speich, B. Blanke, S. Drijfhout (2007). A regional
12 numerical ocean model of the circulation in the Bay of Biscay. J Geophys Res-
13 Oceans, 112, C09008, doi:10.1029/2006JC003935
14

15 Frouin, R., A.F.G. Fiúza, I. Ambar, and T.J. Boyd (1990), Observations of a
16 Poleward Surface Current off the coasts of Portugal and Spain during winter, J
17 Geophys Res-Oceans, 95 (C1), 679-691, doi:10.1029/JC095iC01p00679
18

19 García-Lafuente, J., J.C. Sánchez Garrido, G. Díaz del Río, F. Criado Aldeanueva, D.
20 Marcote, A. Sánchez Román, (2008), Low-frequency variability of the
21 Mediterranean undercurrent off Galicia, northwestern Iberian Peninsula, Journal
22 of Marine Systems, 74, 1-2, 351-363, doi:10.1016/j.jmarsys.2008.02.007
23

24 Garcia-Soto, C., Pingree, R.D., Valdés, L., 2002. Navidad development in the
25 southern Bay of Biscay: climate change and swoddy structure from remote

1 sensing and in situ measurements. J Geophys Res-Oceans 107 (C8)
2 doi:10.1029/2001JC001012

3

4 Haidvogel, D.B., and A. Beckmann (1999), Numerical ocean circulation modeling,
5 Series on Environm. Sci. and Management, Vol.2, Imperial College Press.

6

7 Haynes, R., and E.D. Barton (1990), A poleward flow along the Atlantic coast of
8 the Iberian Peninsula, J Geophys Res-Oceans, 95 (C7), 11425-11441,
9 doi:0.1029/JC095iC07p11425.

10

11 Huthnance, J.M. (1984), Slope currents and "JEBAR", J Phys Oceanogr, 14, 795-
12 810.

13

14 Huthnance, J.M., H.M. Van Aken, M. White, E.D. Barton, B. Le Cann, E. Ferreira
15 Coelho, E. Alvarez Fanjul, P. Miller, and J. Vitorino (2002), Ocean margin
16 exchange — water flux estimates, J Marine Syst, 32, 107-137.

17

18 Iorga, M.C., and M.S. Lozier (1999), Signatures of the Mediterranean outflow from
19 a North Atlantic climatology 1. Salinity and density fields, J Geophys Res-Oceans,
20 104 (C11), 25985-26009 doi:10.1029/1999JC900115.

21

22 Jia, Y., 2000: Formation of an Azores Current Due to Mediterranean Overflow in a
23 Modeling Study of the North Atlantic. J. Phys. Oceanogr., 30, 2342–2358.
24 doi: [http://dx.doi.org/10.1175/1520-0485\(2000\)030<2342:FOAACD>2.0.CO;2](http://dx.doi.org/10.1175/1520-0485(2000)030<2342:FOAACD>2.0.CO;2)

25

1 Kida, S., J.F. Price, and J. Yang (2008), The Upper-Oceanic Response to Overflows:
2 A Mechanism for the Azores Current. *J. Phys. Oceanogr.*, 38, 880–895, doi:
3 10.1175/2007JPO3750.1.

4

5 Lamas, L., Á. Peliz, I. Ambar, A. Barbosa Aguiar, N. Maximenko, and A. Teles-
6 Machado (2010), Evidence of time-mean cyclonic cell southwest of Iberian
7 Peninsula: The Mediterranean Outflow-driven β -plume?, *Geophys Res Lett*, 37,
8 L12606, doi:10.1029/2010GL043339.

9

10 Large, W.G., J.C. McWilliams, and S.C. Doney (1994), Oceanic vertical mixing: a
11 review and a model with a nonlocal boundary-layer parameterization, *Rev.*
12 *Geophys.*, 32, 4, 363-403, doi:0.1029/94RG01872.

13

14 Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, H. E. Garcia, O. K.
15 Baranova, M. M. Zweng, and D. R. Johnson, 2010. *World Ocean Atlas 2009*,
16 Volume 1: Temperature. S. Levitus, Ed. NOAA Atlas NESDIS 68, U.S. Government
17 Printing Office, Washington, D.C., 184 pp.

18

19 Marchesiello, P., J.C. McWilliams, and A. Shchepetkin (2001), Open boundary
20 conditions for long-term integration of regional oceanic models, *Oc. Mod.*, 3, 1-2,
21 1-20, doi:10.1016/S1463-5003(00)00013-5.

22

23 Marchesiello P., L. Debreu, and X. Couvelard (2009), Spurious diapycnal mixing in
24 terrain-following coordinate models: the problem and a solution, *Oc. Mod.*, 26, 3-
25 4, 156-169, doi:10.1016/j.ocemod.2008.09.004.

1

2 Mateus, M., Riflet, G., Chambel, P., Fernandes, L., Fernandes, R., Juliano, M.,
3 Campuzano, F., de Pablo, H., and Neves, R. 2012, An operational model for the
4 West Iberian coast: products and services, *Ocean Sci.*, 8, 713-732.

5

6 Mazé, J.P., M. Arhan, and H. Mercier (1997), Volume budget of the eastern
7 boundary layer off the Iberian Peninsula, *Deep-Sea Res. I*, 44, 9-10, 1543-1574,
8 doi:10.1016/S0967-0637(97)00038-1.

9

10 Neshyba, S. (1986), *Poleward flows along eastern ocean boundaries* (ed.),
11 Springer-Verlag Publisher, ISBN 0387971750, 374 p.

12

13 Otero P., M. Ruiz-Villarreal, and A. Peliz (2008), Variability of river plumes off
14 Northwest Iberia in response to wind events, *J. Mar. Sys.*, 72, 1-4, 238-255,
15 doi:10.1016/j.jmarsys.2007.05.016.

16

17 Paillet, J., M. Arhan, and M.S. McCartney (1998), Spreading of Labrador Sea Water
18 in the eastern North Atlantic, *J Geophys Res-Oceans*, 103 (C5), 10223-10239,
19 doi:10.1029/98JC00262.

20

21 Peliz, A., T.L. Rosa, A.M.P. Santos, and J.L. Pissarra (2002), Fronts, jets, and
22 counter-flows in the Western Iberian upwelling system, *J. Mar. Sys.*, 35, 61-77,
23 doi:10.1016/S0924-7963(02)00076-3.

24

1 Peliz, A., J. Dubert, D.B. Haidvogel, and B. Le Cann (2003), Generation and
2 unstable evolution of a density-driven Eastern Poleward Current: The Iberian
3 Poleward Current, *J Geophys Res-Oceans*, 108 (C8), 3268,
4 doi:10.1029/2002JC001443.

5

6 Peliz, A., J. Dubert, A.M.P. Santos, P.B. Oliveira, and B. Le Cann (2005), Winter
7 upper ocean circulation in the Western Iberian Basin—Fronts, Eddies and
8 Poleward Flows: an overview, *Deep-Sea Res. I*, 52, 4, 621-646,
9 doi:10.1016/j.dsr.2004.11.005.

10

11 Peliz, A., J. Dubert, P. Marchesiello, and A. Teles-Machado (2007), Surface
12 circulation in the Gulf of Cadiz: Model and mean flow structure, *J Geophys Res-*
13 *Oceans*, 112 (C11015), doi:10.1029/2007JC004159.

14

15 Peliz, A. Teles-Machado, P. Marchesiello, J. Dubert, J. Garcia Lafuente, 2009:
16 Filament generation off the Strait of Gibraltar in response to Gap winds,
17 *Dynamics of Ocean and Atmosphere*, 46, 36-45, doi: j.dynatmoce.2008.08.002

18

19 Penven, P., P. Marchesiello, L. Debreu, and J. Lefevre (2008), Software tools for
20 pre- and post-processing of oceanic regional simulations, *Environm. Mod.*
21 *Software*, 23, 5, 660-662, doi:10.1016/j.envsoft.2007.07.004.

22

23 Pingree, R.D., and B. Le Cann (1990), Structure, strength and seasonality of the
24 slope currents in the Bay of Biscay region, *J. Mar. Biol. Ass. of the UK*, 70, 857-
25 885, doi:10.1017/S0025315400059117.

1

2 Relvas, P., E.D. Barton, J. Dubert, P.B. Oliveira, A. Peliz, J.C.B. da Silva, and A.M.P.
3 Santos (2007), Physical oceanography of the western Iberia ecosystem: Latest
4 views and challenges, Prog Oceanogr, 74, 2-3, 149-173,
5 doi:10.1016/j.pocean.2007.04.021.

6

7 Reynaud, T., P. Legrand, H. Mercier, and B. Barnier (1998), A new analysis of
8 hydrographic data in the Atlantic and its application to an inverse modeling
9 study, International WOCE Newsletters, 32, 29-31.

10

11 Richardson, P.L, A.S. Bower, W. Zenk, (2000) A census of Meddies tracked by
12 floats, Prog Oceanogr, 45, 2, 209-250.

13

14 Ríos, A.F., F.F. Pérez, and F. Fraga (1992), Water masses in the upper and middle
15 North Atlantic Ocean east of the Azores, Deep-Sea Res. I, 39, 3-4, 645-658,
16 doi:10.1016/0198-0149(92)90093-9.

17

18 Røed, L.P., and X.B. Shi (1999), A numerical study of the dynamics and energetics
19 of cool filaments, jets and eddies of the Iberian Peninsula, J Geophys Res-Oceans,
20 104 (C12), 29817-29841, doi:10.1029/1999JC900175.

21

22 Saunders, P.M. (1982), Circulation in the eastern North Atlantic, J. Mar. Res., 40,
23 641-657.

24

- 1 Serra, N., I. Ambar, and D. Boutov (2010), Surface expression of Mediterranean
2 Water dipoles and their contribution to the shelf/slope – open ocean exchange,
3 Ocean Sci., 6, 191–209.
- 4
- 5 Shchepetkin, A.F., and J.C. McWilliams (2003), A method for computing
6 horizontal pressure-gradient force in an oceanic model with a non-aligned
7 vertical coordinate, J Geophys Res-Oceans, 108 (C3), 3090,
8 doi:10.1029/2001JC001047.
- 9
- 10 Shchepetkin, A.F., and J.C. McWilliams (2005), The regional oceanic modeling
11 system (ROMS): a split-explicit, free-surface, topography-following-coordinate
12 oceanic model, Oc. Mod., 9, 4, 347-404, doi:10.1016/j.ocemod.2004.08.002.
- 13
- 14 Sibuet, J.-C., S. Monti, B. Loubrieu, J.-P. Mazé, and S. Srivastava (2004), Carte
15 bathymétrique de l'Atlantique nord-est et du golfe de Gascogne: implications
16 cinématiques, Bulletin de la Société Géologique de France, 175, 5, 429-442.
- 17
- 18 Stevens, I., M. Hamann, J.A. Johnson, and A.F.G. Fiúza (2000), Comparisons
19 between a fine resolution model and observations in the Iberian shelf-slope
20 region, J. Mar. Sys., 26, 1, 53-74, doi:10.1016/S0924-7963(00)00038-5.
- 21
- 22 Torres, R., E.D. Barton, P. Miller, and E. Fanjul (2003), Spatial patterns of wind
23 and sea surface temperature in the Galician upwelling region, J Geophys Res-
24 Oceans, 108 (C4), 3130. doi:10.1029/2002JC001361

25

1 Torres, R., and E.D. Barton (2006), Onset and development of the Iberian
2 poleward flow along the Galician coast, *Continental Shelf Res.*, 26, 10, 1134-1153,
3 doi:10.1016/j.csr.2006.03.009.
4
5 Torres, R., and E.D. Barton (2007), Onset of the Iberian upwelling along the
6 Galician coast, *Continental Shelf Res.*, 27, 13, 1759-1778,
7 doi:10.1016/j.csr.2007.02.005.
8
9 Troupin, C., F. Machín, M. Ouberdous, D. Sirjacobs, A. Barth, and J.-M. Beckers
10 (2010), High-resolution climatology of the northeast Atlantic using Data-
11 Interpolating Variational Analysis (Diva), *J Geophys Res-Oceans*, 115, C08005,
12 doi:10.1029/2009JC005512.
13
14 Veitch, J., P. Penven, and F. Shillington (2010), Modeling Equilibrium Dynamics of
15 the Benguela Current System. *J. Phys. Oceanogr.*, 40, 1942–1964, doi:
16 10.1175/2010JPO4382.1.
17
18 Volkov, Denis L., Lee-Lueng Fu, 2010: On the Reasons for the Formation and
19 Variability of the Azores Current. *J. Phys. Oceanogr.*, 40, 2197–2220.
20
21 Zenk, W., and L. Armi (1990), The complex spreading pattern of Mediterranean
22 Water off the Portuguese continental slope, *Deep-Sea Res. I*, 37, 12, 1805-1823,
23 doi:10.1016/0198-0149(90)90079-B.

1 Figure Captions:

2

3 Figure 1 – Map of the region under study: the first domain (FD), comprising the
4 Iberian Peninsula, the north of Africa (Morocco) and the Azores Islands; the local
5 domain (SD) marked with the main bathymetric and topographic features: Strait
6 of Gibraltar (SG), Cape St Vincent (CSV), Gorringe Bank (GoB), Estremadura
7 Promontory (EP), Iberian Abyssal Plain (IAP), Galicia Bank (GB), Cape Finisterre
8 (CF) and Cape Ortegal (CO). The stars mark the locations of analysis for section
9 3.2.2.

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11 Figure 2 – A EKE comparison between altimetry (from AVISO) (top) and FD
12 output (bottom).

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14 Figure 3 – Sea Surface Temperature (SST) in °C (a-c) and Sea Surface Salinity
15 (SSS) (d-f) for January. (a, d) 5-year mean of ROMS output; (b, e) 7-year mean of
16 AVHRR at 02h and GHER-NEA climatology, respectively; (c, f) Difference
17 between (a) and (b) and between (d) and (e), respectively. Thick contours
18 outline null difference. The 200m isobath is shown in thin black contour.

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20 Figure 4 – Same as Figure 3 but for July.

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22 Figure 5 – θ -S diagrams monthly means of ROMS output (dashed lines, one per
23 year, for years 10 to 14) and GHER-NEA climatology (solid line) for January (a,
24 e), April (b, f), July (c, g) and October (d, h) for two regions: one centered at 38°N,
25 10°W (a-d), the other centered at 42°N, 10°W (e-h). Potential density anomaly

1 (σ_t) is superimposed in dotted lines every 0.2 kg m⁻³; the shaded line
2 corresponds to the standard definitions of ENACW_{st} and ENACW_{sp}, defined as
3 $\theta = 10 + 8.462 \cdot (S - 35.4)$ for the subpolar branch (34.8 < S < 35.7) and
4 $\theta = 13.13 + 5.653 \cdot (S - 35.8)$ for the subtropical branch (35.8 < S < 36.2), as defined by
5 *Fiúza* (1984).

6

7 Figure 6 – Zonal sections of temperature and salinity for January: (a, b) 5-year
8 means of ROMS output for 42°N and 38°N, respectively; (c, d) GHER-NEA
9 climatology for 42°N and 38°N, respectively. Sections are down to 1500 m and
10 the first 200 m are enlarged. Temperature is represented in black contours every
11 0.5°C in the upper layers and every 1°C in the deeper layers. Salinity is depicted
12 in white contours every 0.1 and shades of gray for specific values: MW (35.9 to
13 36.3) is represented in light gray and ENACW (35.5 to 35.7) is represented in
14 dark gray. Bathymetry is colored in black.

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16 Figure 7 – Same as Figure 6 but for July, except temperature in the upper layers
17 is represented in black contours every 1°C.

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19 Figure 8 – Meridional section of temperature and salinity at 11.5°W, for January:
20 (a) 5-year means of ROMS output, (b) GHER-NEA climatology.

21

22 Figure 9 – Salinity and velocity fields for January (a, b) and July (c, d) at 50 m (a,
23 c) and at 1000 m (b, d). Isohalines are depicted every 0.2 (36 in thick black
24 contours) and the vector scale is indicated on the map. For information
25 regarding the pointing arrows, please refer to the text.

1

2 Figure 10 – Zonal sections, at 43°N, of 5-year means of potential density
3 anomalies and meridional velocity for January (a) and July (b) down to 1500 m
4 deep. σ_θ is depicted every 0.2 kg m⁻³. The light shades of gray mark negative
5 values below -0.02 m.s⁻¹ and the dark shades of gray positive values above 0.02
6 m.s⁻¹. The thick contour is zero and the first 200 m are enlarged.

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8 Figure 11 – Time series of meridional transport monthly means along: (a-d)
9 43°N; (e-h) 40°N; and (i-l) 37.5°N. Each column corresponds to a sub-section
10 represented in the rightmost column (d, h, l). From left to right: (a, e, i) offshore;
11 (b, f, j) lower slope and adjacent upper layers; (c, g, k) shelf/upper slope (down
12 to 400 m). Seasonal averages (July, August, September – JAS – and November,
13 December, January – NDJ) are plotted in shades of gray for northward and
14 southward transports. To the left of each plot, total northward and southward
15 averages are in black bars and the net average is in white bars.

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17 Figure 12 – Salinity and velocity fields for January (a, b) and July (c, d) at 50 m (a,
18 c) and at 1000 m (b, d) for the NMW run. Isohalines are depicted every 0.2 (36 in
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1 values above 0.02 m.s^{-1} . The thick contour is zero and the first 200 m are
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4 Figure 14 – Schematic representation of the mean circulation at 1000 m as
5 obtained from the present numerical modeling study, showing the principal
6 paths of the obtained circulation structures, and the recirculation vorticity.

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