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**VARIABILIDAD HIDROGRÁFICA Y CIRCULACIÓN DEL
OCÉANO PROFUNDO EN EL NOROESTE DE LA
PENÍNSULA IBÉRICA**

**DEEP OCEAN HYDROGRAPHICAL VARIABILITY AND
CIRCULATION IN THE NORTH-WEST IBERIA**

TESIS DOCTORAL
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“Si he logrado ver más lejos, ha sido porque he subido a hombros de gigantes.”
I. Newton.

Resumen

El muestreo hidrográfico desde 2003 a lo largo de tres secciones oceanográficas profundas perpendiculares a la costa, 200 millas náuticas al oeste de cabo Finisterre en 43°N , 100 millas al norte de cabo Ortegal (8°O) y 100 al norte de Santander ($3^{\circ}47'\text{O}$), ha permitido el análisis más detallado hasta la fecha de la variabilidad estacional e interanual de las propiedades termohalinas de las masas de agua en el noroeste de la península ibérica y sur del golfo de Vizcaya.

El muestreo semianual de la sección de Finisterre durante el periodo 2003-2010 ha revelado una fuerte estacionalidad en las propiedades termohalinas de todas las masas de agua de la termoclina permanente (hasta los 2000 dbar), caracterizada por una notable asimetría entre la región del talud y el océano abierto. En verano la vena de agua mediterránea aparece ceñida al talud y menor profundidad, reforzando su huella termohalina que a su vez disminuye en océano abierto. En invierno la situación se invierte y el agua mediterránea parece desprenderse del talud y propagarse hacia el oeste observándose incluso una rama secundaria alrededor del banco de Galicia. La estacionalidad en aguas intermedias alcanza valores de hasta 0.4°C y 0.08 en salinidad en la base del agua mediterránea, del orden del 20% de la variabilidad interanual observada durante todo el periodo. La descomposición de los cambios isobáricos en cambios isopícnos y cambios debidos al desplazamiento vertical de las isoneutras se muestra útil a la hora de analizar los patrones observados en función de (1) la estacionalidad a gran escala del giro subtropical como respuesta a la migración del sistema de altas presiones en el Atlántico Norte, (2) la dinámica del talud continental, caracterizada por el afloramiento en verano, el desarrollo de la Corriente Ibérica hacia el Polo en invierno y la propagación del agua mediterránea, y (3) la posible influencia de los cambios estacionales de las propiedades de las masas de agua adquiridas en sus regiones de formación.

Una vez cuantificada y extraída la estacionalidad se analizaron las series temporales de las propiedades de las diferentes masas de agua a lo largo de las tres

secciones. La correlación entre cambios en la columna de agua, secciones y la propia autocorrelación de cada serie indican que las señales observadas son cambios robustos de las masas de agua a escala regional. Las series no muestran tendencias suaves sino que se caracterizan por transiciones que una vez producidas persisten durante campañas consecutivas. Como observaciones destacadas se incluye un aumento relativamente brusco de la salinidad de las aguas centrales en 2005 que permanece hasta la actualidad y una disminución de las propiedades termohalinas del agua con origen en el mar de Labrador desde el otoño de 2008 a 2010. Se observa una respuesta integrada a estados fuertes de la Oscilación del Atlántico Norte, destacando el calentamiento/salinificación de la columna de agua tras la fuerte caída del índice NAO en 2010. Este patrón es consistente con la respuesta esperada de la circulación a gran escala del Atlántico Norte, caracterizada por la propagación de aguas cálidas y salinas hacia el norte en la cuenca oriental durante un índice NAO negativo.

Las campañas de febrero y septiembre, caracterizadas por la estructura invierno/verano típica, disponen además de un conjunto de medidas directas de velocidad de corriente —perfilador acústico en roseta (LADCP), ADCP de casco y una línea de fondeo— que han servido para estimar transportes a través de la sección y comparar los diferentes métodos posibles. Se observan tres regiones bien diferenciadas, plataforma-talud caracterizada por transporte hacia el norte, el paso hasta el banco de Galicia con transporte sur y el oeste del banco de nuevo con transporte norte. Las corrientes en el paso hasta el banco de Galicia son predominantemente barotrópicas por lo que las estructuras que se observan en los campos gesotróficos se ven advectadas.

Summary

Oceanic hydrography of the north-easternmost region of the North Atlantic subtropical gyre has been monitored since 2003 for the whole water column (> 5000 m) by three sections extending zonally 200 nautical miles (nm) westwards of Cape Finisterre (43°N , western Iberia margin), and meridionally 100 nm off Cape Ortegal (8°W , southwestern Bay of Biscay) and 100 nm off Santander ($3^{\circ}47'\text{W}$, southeastern Bay of Biscay), allowing for the analysis of the deep ocean variability from seasonal to interannual scales.

Semiannual time series of the Finisterre oceanographic section from 2003 to 2010 have shown that all water masses down to the permanent thermocline (2000 dbar) exhibit a consistent seasonal signature in their thermohaline properties with notable asymmetry between the slope region and the outer ocean. In summertime, Mediterranean Water (MW) gets tightly attached against the slope and is uplifted, reinforcing its thermohaline signature and diminishing its presence at the outer ocean. In wintertime the situation reverses, MW seems to detach from the slope and spreads out to the open ocean, even being observed a secondary branch around the Galicia Bank. Thermohaline seasonality at depth shows values up to 0.4°C and 0.08 in salinity at the lower MW, of the order of 20% of the overall interannual variability observed during the whole period. Decomposition of thermohaline changes at isobaric levels to changes along isoneutral surfaces and changes due to vertical displacements helped analyze the physical processes behind the observed seasonality in terms of (1) the large-scale seasonality of the subtropical gyre in response to the seasonal migration of the subtropical high pressure system, (2) the continental slope dynamics, characterized by summer upwelling, winter development of the Iberian Poleward Current and Mediterranean water spreading, and (3) the possible influence of seasonal changes of water mass properties at their formation sources.

Once signals were de-seasonalized, time series of property interannual changes of the different water masses were analyzed. Correlation of series in the verti-

cal and among sections, autocorrelation and estimates of the effect of the noise induced by the mesoscale field, all indicate that observed signatures are robust changes of water masses at the regional scale. The hydrographic timeseries are not characterized by smooth trends but instead by shifts that persist through consecutive cruises. The most notable features include a shift to more saline central waters around 2005 after which they remained stable, and a decrease in thermohaline properties of the Labrador Sea Water from autumn 2008 to 2010. Years with a strong winter North Atlantic Oscillation (NAO) index are characterized by shifts in thermohaline properties across most of the intermediate levels, with the most notable event being the warming and increasing salinity that followed the large NAO index drop of 2010. These observations are consistent with current understanding of the large-scale functioning of the North Atlantic, which predicts a northeastwards expansion of subtropical temperate waters in the eastern boundary as a response to negative NAO forcing.

February and September cruises in 2008, exhibiting the typical winter/summer thermohaline distinct structures and also having a quite complete set of direct velocity measurements — Lowered Acoustic Current Profiler (LADCP), ship hull ADCP and an active mooring line — were explored to estimate transports across the section and compare different methods available. Three regions were evidenced, the continental shelf break-slope characterized by northwards transport, southwards circulation across most of the passage to the Galician Bank summit and then northwards flow again west of the seamount. Currents across the passage from Finisterre to the Galician Bank were markedly barotropic so mesoscale structures evidenced by geostrophy were being advected across the passage.

Glossary

AC Azores Current

BT Bottom-Track

CC Canary Current

CTD Conductivity-Temperature-Depth

ENACW Eastern North Atlantic Central Water

ENADW Eastern North Atlantic Deep Water

IEO Instituto Español de Oceanografía / Spanish Institute of Oceanography

IPC Iberian Poleward Current

IPCC Intergovernmental Panel on Climate Change

LADCP Lowered Acoustic Doppler Current Profiler

LDW Lower Deep Water

LSW Labrador Sea Water

MOCHA Meridional Overturning Circulation and Heat Transport Array

MW Mediterranean Water

NAC North Atlantic Current

NAO North Atlantic Oscillation

OCCAM Ocean Circulation and Climate Advanced Modelling

PC Portugal Current

RadProf Radiales Profundas

RAPID Rapid Climate Change

SADCP Ship-mounted Acoustic Doppler Current Profiler

SCOW Scatterometer Climatology of Ocean Winds

SPG Subpolar Gyre

SPGI Subpolar Gyre Index

STG Subtropical Gyre

VACLAN/COVACLAN Variabilidad Climática del Atlántico Noreste // Coordinación y Optimización de VACLAN

WOCE World Ocean Circulation Experiment

CHAPTER 1

General Introduction

1.1 The importance of the long-term ocean monitoring programs

The ocean is a crucial component of Earth's climate, playing a key role in the regulation of global energy budget by transporting heat and salt from low to high latitudes and storing large amounts of heat and carbon dioxide. The ocean is assumed to be absorbing large amounts of heat in response to ongoing climate change ([Barnett *et al.*, 2005](#)) but, due to the non-linear responses of complex ocean-atmosphere system, this does not imply sustained long-term trends in hydrographic properties at all basins or sites but rather a more heterogeneous response. Moreover, the different time scales of variability require sufficiently long and regularly sampled time series to detect long-term trends ([Wunsch, 1999](#)). Oscillations on decadal time scales appear naturally in the climate system ([Latif & Barnett, 1996](#)) and so caution is needed when treating any observed change in water mass properties or ocean circulation as an indicator of climate change (e.g. [Bryden *et al.*, 2003](#)). For these reasons oceanographic panels currently urge the establishment of systematic and sustained programs of ocean observation ([Baker *et al.*, 2007](#); [Hughes *et al.*, 2012](#)).

There exists a large number of studies dealing with trends and interannual variability of water mass properties, stating widespread warming since at least the second half of the 20th century (e.g. [Solomon *et al.*, 2007](#)). However, information about seasonal fluctuations in the ocean interior is less robust and more scarce, due to the lack of subannual monitoring programs, adding a potential source of noise when trying to solve climatic trends. Seasonality in water mass properties

and current fields are expected as a response of the yearly heating/cooling cycle and the large-scale response in the wind curl, and evidence of seasonal variability below the seasonal thermocline has long ago been anticipated theoretically (e.g. [Krauss & Wuebbler, 1982](#), for the case of the North Atlantic). Recently [Chidichimo et al. \(2010\)](#) and [Kanzow et al. \(2010\)](#) reported a strong contribution of the eastern boundary density variations in the subtropical Atlantic to subseasonal and seasonal anomalies in the strength and vertical structure of the Atlantic Meridional Overturning Circulation (AMOC). The AMOC plays a key role in the meridional heat transport in the North Atlantic transporting ~ 19 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) of warm, saline waters northward above ~ 1000 m depth and the same amount of cold water southward below 1000 m. The strength of the AMOC was estimated from sparse hydrographic observations with insufficient temporal resolution ([Good et al., 2013](#); [Dickson et al., 1988](#); [Nolan et al., 2012](#); [Bryden et al., 2005](#); [Lohmann et al., 2009](#)) which could bias the detection of AMOC trends. Aiming for the continuous monitor of the temporal evolution of the AMOC, the RAPID (Rapid Climate Change)/MOCHA (Meridional Overturning Circulation and Heat Transport Array) transatlantic currentmeter array along 26.5°N became operative in 2004. Such array deployment has recently revealed that the 8 Sv decline in the AMOC transport observed by [Bryden et al. \(2005\)](#) was biased by a seasonal cycle of 6.7 Sv in the AMOC transport ([Kanzow et al., 2010](#)). They stated that seasonal geostrophic AMOC anomalies might represent an important and previously underestimated component of meridional transport and storage of heat in the subtropical North Atlantic. Moreover, 5.2 Sv of such seasonal cycle arise from seasonal density variations of the North Atlantic eastern boundary ([Chidichimo et al., 2010](#)). The North Atlantic circulation is also subject to strong interannual, decadal and multidecadal variability. In particular, hydrographic sections in the sub-polar gyre from the coast of the Labrador Sea to Europe show a multi-decadal variability in temperature, while convective events occur on decadal or shorter time scales ([van Aken et al., 2011](#)). Such events were shown to homogenize the properties of intermediate layers for significant periods of time, emphasizing the leading role of Labrador Sea convection as a driving mechanism of the variability of intermediate water layers in the North Atlantic. Recent warming over the upper 3000 m of the North Atlantic has been attributed to both long-term climate change (including anthropogenic) and natural multidecadal variability, the latter accounting up to a 60% of the warming since 1970 ([Polyakov et al., 2010](#)). Embedded in this positive phase of multidecadal variability characterized by warmer than normal conditions both in the land and sea, 2000-2010 was the warmest decade on record, reaching the highest global average temperatures in 2010 ([Hansen et al.,](#)

2010). Changes in the subpolar North Atlantic hydrographic properties are tightly related to large-scale atmospheric forcing induced by the North Atlantic Oscillation, NAO (e.g. [Sarafanov, 2009](#)); however, an overall weak and variable NAO during the first decade of the present century made the relationship less robust ([Hughes et al., 2012](#)). The North Atlantic Subpolar Gyre (SPG) index was recently proposed as a further indicator of ocean variability that should be considered in climate change studies ([Häkkinen & Rhines, 2004](#); [Hátún et al., 2005](#)). There are a number of well established monitoring programs in the subpolar regions that have led to long timeseries ([Sarafanov et al., 2010](#); [Hughes et al., 2012](#)). However, the temperate mid-latitude areas (except the repetition of the WOCE A5 26.5°N transatlantic section which has been sampled on regular basis since 2004 ([Cunningham et al., 2007](#))), have been observed less frequently.

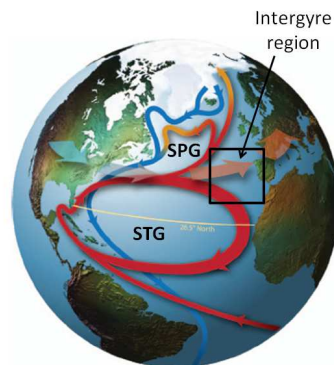


Figure 1.1: Schematic picture of the North Atlantic large scale circulation modified from [Srokosz et al. \(2012\)](#), showing the intergyre region between the subpolar (SPG) and subtropical (STG) gyres.

In this work, we exploit a dataset of semiannual occupations of an hydrographic section located west of the Iberian Peninsula (intergyre region, see Fig. 1.1) at 43° N, 009–014° W (off Cape Finisterre) to explore the seasonal variability of water masses below the mixed layer in the midlatitude eastern boundary of the North Atlantic. We also use 11-year time series of property variations along this section and another two deep sections located at the southern Bay of Biscay (off Cape Ortegal, 8° W and Santander, 3°47' W), to analyze the interannual thermo-

haline changes observed during 2003-2013. Local regional changes are related to changes observed in other distant regions and outcomes are interpreted in the context of main driving mechanisms of North Atlantic ocean variability. Estimates of water mass transports by directly measured ocean currents are necessary to understand the role that the ocean plays in the world's climate but also, on shorter scales, to infer the connectivity between different marine ecosystems. Direct velocities (LADCP) are measured and geostrophic velocity currents are computed to provide the best representation of circulation patterns in the region. A method based on referencing the geostrophic fields using LADCP bottom-tracking is also introduced following the study carried by *Comas-Rodriguez et al. (2010)* along the transatlantic section 24.5°N. The LADCP bottom-tracking provides accurate velocity measurements near the bottom, permitting high-resolution spatial studies that can reveal circulation patterns of deep waters not observable by other way.

1.2 Objectives

The main thesis objectives are:

1. To process, to ensemble and to analyze the hydrographic database associated with the VACLAN/COVACLAN projects of the Spanish Institute of Oceanography (IEO) running since 2003, aiming for the standarization of data format and the organization of data by a common structure of folders, making easier to access the information.
2. To provide novel and relevant findings to the scientific community, deepening on the knowledge of the driving mechanisms behind the deep ocean thermohaline properties variability and circulation in the north-west Iberia. The regular repetition of perpendicular to the coast transects on board oceanographic research vessels provides the necessary hydrographical time series for the study of the regional climate variability.
3. To determine the different spatial and temporal scales of variability, as well as the correlation of changes observed in the themohaline properties of water masses (vertically in the water column, temporally during the sampling period 2003-2013 and geographically between different sections).
4. To decompose the isobaric changes observed in temperature and salinity fields on their two natural components of change, following the method developed by *Bindoff & McDougall (1994)*: 1) the isopycnal change, related to

the intrinsic variation of thermohaline properties without change in density; this term is related to changes in the air-sea fluxes at the formation region or to advection, which means that the sampled water mass is essentially the same (no density change) but with different properties, and 2) the change by *heave*, due to the shoaling or deepening of isopycnal surfaces in the water column, due to changes in the wind-induced circulation or renewal rates of water masses.

5. To determine the relevance of the seasonal cycle at intermediate depths of the North Atlantic eastern boundary mid-latitudes, extending results from southern latitudes and supporting the importance that the eastern boundary mid-ocean seasonality has in regional and global climate change.
6. To analyze the time series of interannual variability in the study region between 2003-2013, in terms of the North Atlantic large-scale atmospheric forcing, including the deep ocean response to strong shifts in the North Atlantic Oscillation.
7. To establish similarities and differences between the three methods available to estimate the oceanic water mass transports in the region: 1) directly measured velocity fields, 2) indirect velocities calculated from thermohaline properties (geostrophic approximation) and 3) geostrophic fields corrected using near-bottom velocities.

1.3 Thesis organization

The thesis is structured as follows: first, the regional oceanography and main water masses west of the Iberian Peninsula are described in Chapter 2, and the database provided by the series of VACLAN/COVACLAN RadProf oceanographic cruises are introduced. Methods applied for the decomposition of thermohaline changes of water masses, both at seasonal and interannual scales, and the methodology used for the computation of associated transports, both from geostrophy and direct velocity measurements are described in Chapter 3. Results obtained are presented in Chapter 4 and discussed in Chapter 5 through the corresponding subsections. Conclusions are summarized in Chapter 6 and Chapter 7, in English and Spanish, respectively. Future work is proposed in Chapter 8 and References used throughout the thesis are included at the end of the document.

CHAPTER 2

Regional oceanography and data set

2.1 Water masses and circulation in the northwest Iberian Peninsula

The western Iberian margin is a portion of the North Atlantic eastern boundary located at the north-eastern edge of the subtropical gyre. Upper ocean circulation is weak, exhibiting a mean southward flow of few cm s^{-1} (*Mazé et al.*, 1997; *Paillet & Mercier*, 1997), but dominated by mesoscale activity (*Memery et al.*, 2005).

The continental slope is characterized by the development of a density driven countercurrent (poleward) known as the Iberian Poleward Current, IPC (e.g. *Frouin et al.*, 1990; *Peliz et al.*, 2005). A detailed review of the modal, intermediate and deep water masses of the midlatitude north-east Atlantic Ocean was performed by *van Aken* (2000a,b, 2001) and pathways of such waters masses from their source region towards western Iberia are shown in Fig. 2.1b.

Briefly, these are as follows:

- Eastern North Atlantic Central Water (ENACW) constitutes the upper permanent thermocline of the European and north-west African basins. It is characterized by a narrow and nearly straight line in the θ/S with its core located at ~ 350 dbar. ENACW is formed by winter mixing in a wide region from the Azores to the European boundary bounded by the North Atlantic Current (NAC) and the Azores Current (AC) (*Pollard & Pu*, 1985; *Pollard et al.*, 1996). In the western Iberia margin, warmer ENACW of subtropical origin created near the Azores flows eastward and mixes with the colder

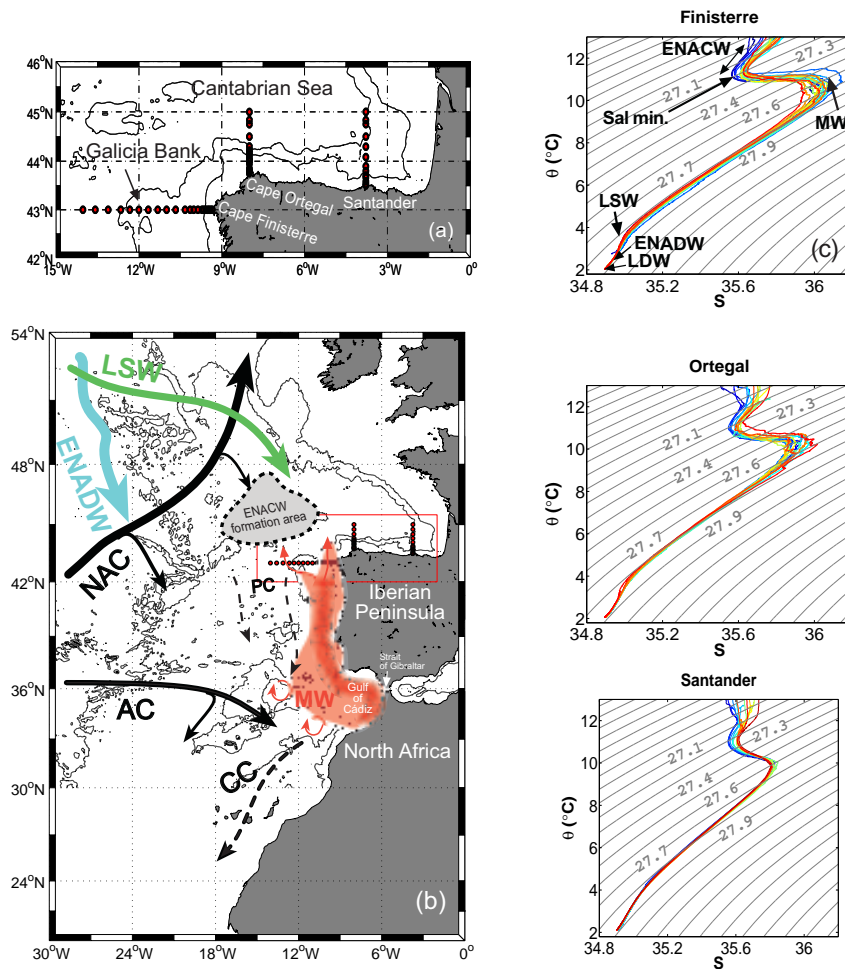


Figure 2.1: **(a)** Zoom of the VACLAN/COVACLAN sampling region including the Finisterre (western Iberia margin), Ortegal (southwestern Bay of Biscay) and Santander (southeastern Bay of Biscay) sections. **(b)** Main circulation patterns of waters in the area. Gyres bounding North Atlantic Current (NAC) and Azores Current (AC) (black bold), southward flowing Portugal Current (PC) and Canary Current (CC) (black dashed). Central water: ENACW formation area is shown as a dashed grey region north of Finisterre, from where it flows southwards with the PC. Intermediate waters: Mediterranean water (MW, red), spreading from the Strait of Gibraltar, flows northwards along the continental slope sometimes showing a detachment contouring the west Galicia Bank, and southward flowing water from the Labrador Sea (LSW, green). Eastern North Atlantic Deep Water (ENADW, blue). **(c)** From top to bottom, θ - S diagram of along-section profile averages in Finisterre, Ortegal and Santander, between 2003 (darkest blue) and 2013 (darkest red), overlying σ_t contours in grey.

southward flowing ENACW of subpolar origin (*Fiúza et al., 1998*). The lower bound of ENACW is characterized by a salinity minimum.

- Mediterranean Water (MW) is formed at the Gulf of Cádiz from the intense mixing of Atlantic central waters and the warm and salty overflow from the Mediterranean Sea through the Strait of Gibraltar. The main core is characterized by a clear salinity maximum spreading from Cape St. Vincent at 37° N as a northward deep boundary current along the European ocean margin, and reaching high latitudes as the Porcupine Bank at 53° N. Along the west Iberian margin, two cores are found: the upper (MW_u) at ~ 750 dbar and the lower (MW_l) at ~ 1250 dbar, becoming indistinguishable north of 42° N. West of Cape Finisterre, an intermediate salinity maximum centered at 1000 dbar is observed between a salinity minimum at about 500 dbar and a minimum connected with the below-lying Labrador Sea Water (LSW) core near 1900 dbar (*Iorga & Lozier, 1999b; van Aken, 2000b*).
- Labrador Sea Water (LSW) is the deeper intermediate water mass (i.e. belonging to the permanent thermocline). It is the last stage of the modification of the subpolar gyre mode waters formed in the centre of the cyclonic circulation in the Labrador Sea by deep winter convection, spreading southwards and eastwards in the North Atlantic from its formation area. In the north-east Atlantic, its core is characterized by a deep salinity minimum near 1900 dbar (*Pingree, 1973*).
- Eastern North Atlantic Deep Water (ENADW), found below the intermediate water masses, is formed by a mixture of different polar source water types including a component of Antarctic origin (*van Aken, 2000a*). The term Lower Deep Water (LDW) is used for the bottom waters found deeper than 4000 m.

Table 2.1 summarizes properties of water masses separated by density layers (following the scheme devised by *van Aken (2000a,b, 2001)*) with corresponding along-section mean values of pressure, potential temperature and practical salinity.

2.2 The VACLAN/COVACLAN projects

A monitoring programme of ocean properties has been in place since 2003 in the southern Bay of Biscay and northwestern Iberia margin, under the VACLAN/COVACLAN

Table 2.1: Potential density layers defining water masses according to *van Aken (2000a,b, 2001)*. Corresponding neutral densities (γ^n , *Jackett & McDougall (1997)*) and depth-averaged values of pressure (\bar{P}), potential temperature ($\bar{\theta}$) and salinity (\bar{S}) along Finisterre (subscript Fist), Ortegal (Orte) and Santander (Sant)

Water mass	Density (kg m^{-3})	γ^n (kg m^{-3})	\bar{P}_{Fist}	$\bar{\theta}_{\text{Fist}}$	\bar{S}_{Fist}	\bar{P}_{Orte}	$\bar{\theta}_{\text{Orte}}$	\bar{S}_{Orte}	\bar{P}_{Sant}	$\bar{\theta}_{\text{Sant}}$	\bar{S}_{Sant}
ENACW	$27.00 < \sigma_\theta < 27.20$	27.06–27.26	297	12.16	35.70	278	11.92	35.64	265	11.84	35.61
Sal min	$\sigma_\theta \approx 27.2$	27.26	475	11.30	35.61	463	11.22	35.59	457	11.22	35.59
MW	$31.85 < \sigma_1 < 32.25$	27.46–27.79	1002	10.07	35.85	1033	9.41	35.69	1020	9.18	35.63
LSW (core)	$\sigma_2 \approx 36.88$	27.93	1764	4.64	35.13	1786	4.41	35.08	1778	4.57	35.12
ENADW	$41.42 < \sigma_3 < 41.51$	28.02–28.10	3236	2.60	34.94	3230	2.59	34.94	3226	2.59	34.94
LDW (core)	$\sigma_4 \approx 45.83$	28.09	4500	2.08	34.90	4500	2.09	34.90	>4000	2.12	34.90

projects (Spanish acronym of “Coordination and optimization of the Northeast Atlantic CLimate VAriability Observing System”) of the Spanish Institute of Oceanography. The program aims to maintain a continuous observation programme of climate variability in this region of the North Atlantic eastern boundary. Three deep sections perpendicular to the coast extent to 200 nm off Cape Finisterre (northwestern Iberia, 43°N , > 5000 m), 100 nm off Cape Ortegal (southwestern Bay of Biscay, 8°W , > 4000 m) and 100 nm off Santander (southeastern Bay of Biscay, $3^{\circ} 47'\text{W}$, > 4000 m) as it is shown in Fig. 2.1a,b. The sections were sampled twice a year from 2003 to 2010, with data collected to WOCE standards. In 2011 sections were only sampled in summer, and only sampling along the Finisterre section was performed in summer 2012 and 2013. The cruises consist of standard hydrography performed by SBE911+ CTD provided with oximeter, turbidimeter and fluorometer integrated into an oceanographic rosette (see Fig. 2.2).

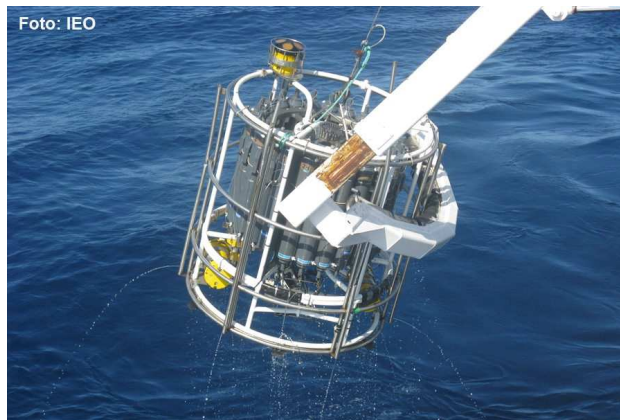


Figure 2.2: Oceanographic rosette leaving the water after the finalization of a hydrographic station. A set of 24 10l Niskin bottles collect water from different depths as they are closed automatically from the vessel. CTD lies horizontally at the bottom of the frame. LADCP upward and downward looking devices (in yellow) are placed at the top and bottom of the structure, with a shared pack of batteries.

The diagrams of Fig. 2.1c, provide a first overview of the year-to-year thermohaline variability in all depths of the water column. A Lowered Acoustic Doppler Current Profiler (LADCP) was mounted on the oceanographic rosette and deployed at each CTD station, providing simultaneous full-depth velocity measurements since September 2004. The LADCP consisted on two 300 kHz RDI Workhorse (WH) devices run in master/slave mode, one upward looking (slave)

and one downward looking (master) with their own two shared batteries. Continuous velocity observations were made at the upper 500-750 m of the ocean along the ship track using a shipboard ADCP (SADCP) RDI Ocean Surveyor 75 kHz installed on the research vessel's hull from 2008. Accurate velocities near the ocean bottom are also provided by the bottom-track (BT) of the LADCP. A mooring line with current-meters at the levels of ENACW, MW and LSW (300, 1000 and 1800 m) was set at 43°N, 11°W in 2003 and operated intermittently until 2010.

The number of sampled stations and the start/end dates of surveys are shown in Table 2.2, where $N_{\text{Fist.}}$, $N_{\text{Orte.}}$ and $N_{\text{Sant.}}$ refers to the number of stations sampled at Finisterre, Ortegal and Santander sections respectively. Note that the Finisterre section was occupied a total of 18 times during the sampling period, becoming the most repeated and the longest and the deepest sampled transect. Ortegal section was not sampled in February or September 2010 (14 occupations) and Santander section was not sampled in February 2010 (15 occupations).

Table 2.2: VACLAN/COVACLAN series of cruises. “RadProf” is an acronym for the Spanish of “Deep Section”. Dates and number of stations sampled at Finisterre ($N_{\text{Fist.}}$, 43°N), Ortegal ($N_{\text{Orte.}}$, 8°W) and Santander ($N_{\text{Sant.}}$, 3° 47'W).

Cruise	dates	N_{Fist}	N_{Orte}	N_{Sant}
RadProf200303	26 March – 17 April, 2003	13	11	11
RadProf200309	10–20 September, 2003	16	8	11
RadProf200402	5–13 February, 2004	18	11	12
RadProf200409	7–13 September, 2004	18	11	11
RadProf200501	27 January – 3 February, 2005	6	10	9
RadProf200508	20 August – 9 September, 2005	20	14	12
RadProf200602	5–13 February, 2006	19	14	12
RadProf200607	12–29 July, 2006	19	13	10
RadProf200702	1–7 February, 2007	21	14	6
RadProf200802	11–19 February, 2008	19	14	12
RadProf200809	3–13 September, 2008	18	14	12
RadProf200902	1–13 February, 2009	15	13	10
RadProf200908	11–21 August, 2009	19	14	12
RadProf201002	9–11 February, 2010	19	–	–
RadProf201009	2–11 September, 2010	19	–	12
RadProf201108	14–24 August, 2011	21	14	12
RadProf201209	12–16 September 2012	23	–	–
RadProf201309	13–17 September 2013	24	–	–

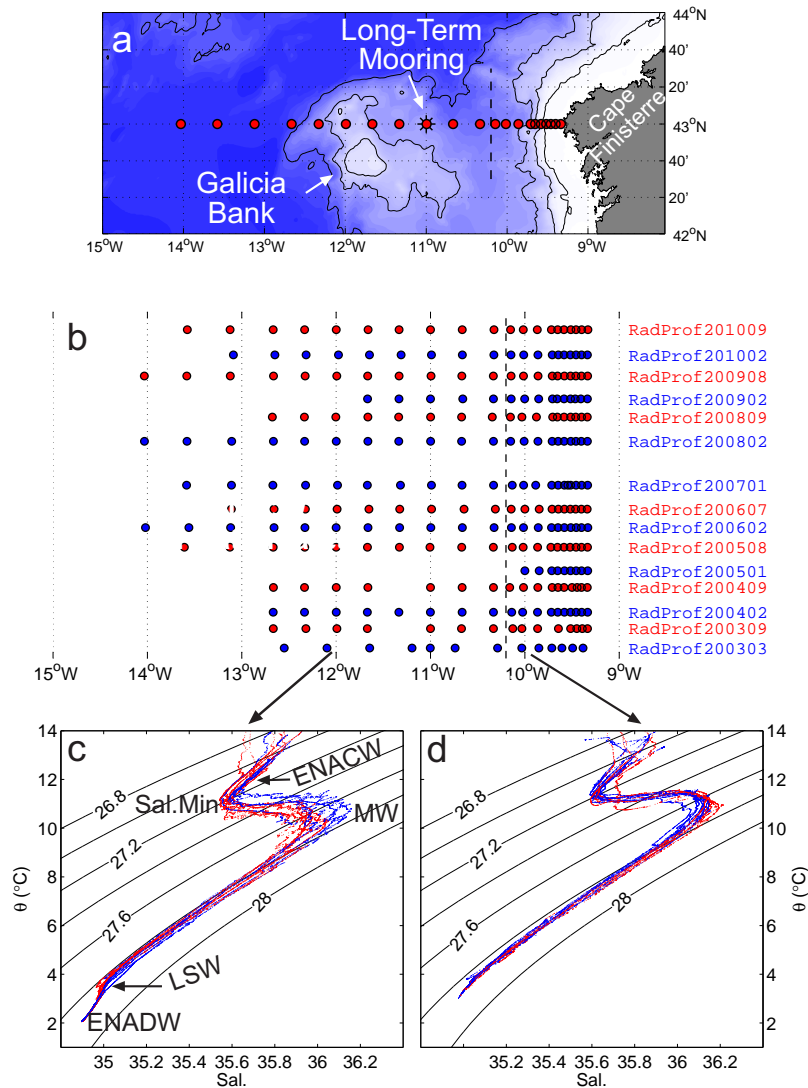


Figure 2.3: **(a)** Zoom of the Finisterre section. **(b)** Graphical representation of the hydrographical database. Dots are plotted at true positions with reference to **(a)**. Blue cruises are for winter and red for summer. **(c)** θ S diagram of all profiles at a station 012°40' W, west of the Galicia Bank (colours keep to represent winter vs summer profiles). **(d)** Same as **(c)** for a slope station 009°43' W

Semiannual occupations of the Finisterre section between 2003-2010 are showed in Fig. 2.3a,b). A seasonal pattern emerges at first view from the representation of

the θ_S properties (Fig. 2.3c) specially at MW levels. The study of water masses circulation presented is based on the hydrography+LADCP data collected during those seasonal surveys performed in winter (February) and summer (September) 2008 along the Finisterre section (43°N), cruises RadProf200802 and RadProf200809 respectively, which represent typical winter and summer conditions, as we will see. In the first cruise a total of 21 stations were sampled (Fig. 2.3b) and the LADCP was used in dual mode as planned. In the second cruise (18 stations) bad weather conditions made necessary the replacement of the oceanographic rosette by a smaller one with only one device (master) running in downward looking mode.

CHAPTER 3

Methodology

3.1 Decomposition of hydrographic changes along neutral density surfaces: seasonality and inter-annual variability.

The first step in order to quantify the seasonality in the different waters masses is to adopt a systematic and objective classification of them. Water masses are distributed as layers with distinct hydrographical properties placed approximately at specific depths. Normally, water masses are labelled by characteristic density levels, while depth is only taken as reference below the permanent thermocline where density gradient is very weak. The most detailed description of the water masses in the midlatitude north-east Atlantic ([van Aken, 2000a,b, 2001](#)) used isopycnal bounds to tag the main water masses referred to in Sect. 2.1, with the exception of LDW identified with deep water located roughly below 4000 dbar. We have followed these works in order to label the water masses, with the exception of ENACW and the Salinity Minimum, which have been identified focusing on the θ/S diagrams of Figs. 2.1c, but using neutral density γ^n ([Jackett & McDougall, 1997](#)) instead of potential density. Figure 3.1 provides a graphical view of the water masses division.

Different behaviour is expected on the slope or further into the basin, and actually a distinct pattern emerges from a first glance of the dataset (see Fig. 2.3c,d). For this reason, the section will be divided into two subsections (see Fig. 2.3a) with the aim of studying the seasonality separately . These are:

1. The *slope*, which extends from the Galicia coast off Cape Finisterre at

$\sim 9.3^\circ$ W to the middle of the trough at 10.2° W.

2. The *outer ocean*, which spans from 10.2° W to 14° W, hence including the Galicia Bank.

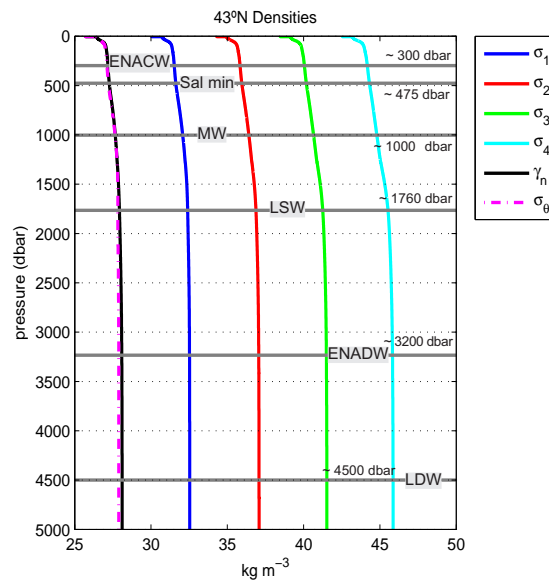


Figure 3.1: Time and zonally averaged potential density relative to the sea surface (σ_θ), 1000 dbar, 2000 dbar, 3000 dbar, 4000 dbar ($\sigma_{1,2,3,4}$) and neutral density (γ^n) as a function of pressure.

In order to analyse water mass changes, it is helpful to consider separately thermohaline changes occurring at fixed density levels (hence causing a true modification of the θS diagram) and vertical displacement of isopycnal levels (a process known as heave). Properties of modal waters formation are set by the balances in atmospheric forcing, so altered heat and freshwater fluxes shaping the mixed layer that is later subducted into the ocean interior will cause subsurface isopycnal changes (warming/freshening). On the other hand, the vertical displacement of density surfaces (which changes the properties at a given depth) can be caused either by changes in the rates of renewal of water masses or by dynamical changes affecting the density field structure. As water masses move further from their sources, the diffusive processes affecting their transformation contribute to blur the specific signatures acquired at the formation stages. *Bindoff & McDougall*

(1994) developed a systematic methodology to characterize water mass changes, showing that observed changes along pressure surfaces can be approximately split into changes of properties along neutral density surfaces and changes due to the displacement of isopycnals through the vertical gradients of properties assumed constant over time. An schematic view of the decomposition of changes is shown in Fig. 3.2.

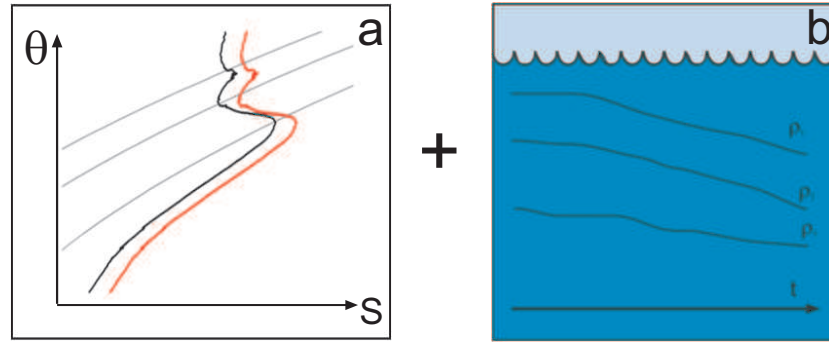


Figure 3.2: Isobaric changes can be decomposed into an isopycnal change causing a modification of the θ/S diagram (a) plus a change by heave due to the variation of the isopycnals pressure with time (b).

Isobaric changes can be mathematically expressed as,

$$\left. \frac{d\theta}{dt} \right|_p \simeq \left. \frac{d\theta}{dt} \right|_n - \left. \frac{dp}{dt} \right|_n \frac{\partial \theta}{\partial p} \quad (3.1)$$

$$\left. \frac{dS}{dt} \right|_p \simeq \left. \frac{dS}{dt} \right|_n - \left. \frac{dp}{dt} \right|_n \frac{\partial S}{\partial p}, \quad (3.2)$$

where p and n indicate, respectively, changes at isobars and isopycnals. The left-hand term is thus called the *isobaric change*, the first right-hand term the *isopycnal change* and the second one the *heave term*. The vertical coordinate pressure p is assimilated to the z -axis so Eqs. (3.1) and (3.2) may be written in compact notation as $\theta'_z \simeq \theta'_n - N' \theta'_z$ and $S'_z \simeq S'_n - N' S'_z$, where N' is the change in height of the neutral surface and the primes indicate temporal change of the scalar quantities. We will apply this methodology in order to characterize the seasonal changes occurring in different water masses in the two regions.

Temporal changes are usually calculated by comparing properties between two cruises, regardless of the time-lag between them and lack of information about variability at shorter timescales. Since we have a true time series of hydrographical sections, we will consider changes referred to an overall average section instead of calculating changes between pairs of cruises. Therefore we will construct the series of anomalies of key properties (isopycnal/isobaric changes and heave) with respect to the overall average section. From these series will be estimated the presence of seasonality and its amplitude.

In order to quantify interannual changes in water properties it is necessary to extract the seasonal cycle. The deseasonalized signal is computed by subtracting each cruise section from the corresponding winter (summer) average field, i.e.

$$\theta_i^{\text{des}} = \theta_i - \bar{\theta}_{\text{winter/summer}}, S_i^{\text{des}} = S_i - \bar{S}_{\text{winter/summer}} \quad (3.3)$$

where i is the number of repeats of a section and subscript “des” refers to the deseasonalized field. Anomalies are calculated at each profile site and then zonally averaged for every cruise i . Next the anomalies are vertically averaged within pressure intervals 0-200, 200-400 and from 400 dbar to the bottom. Then, changes of properties along pressure surfaces are decomposed into changes along neutral density surfaces (isopycnal change) plus changes due to the displacement of isopycnals (heave) following the work of *Bindoff & McDougall (1994)* in the same way as for seasonal changes. We will apply methodology for the decomposition of changes using neutral density surfaces (*Jackett & McDougall, 1997*) every 0.1 kg m^{-3} , in the water column between $27\text{-}28 \text{ kg m}^{-3}$.

An important issue is the magnitude of the high frequency variability along the section. We expect that our series are representative of the slow varying changes in water masses properties at a regional scale. Without a higher sampling rate of the section it is not possible to quantify the bias induced by mesoscale and/or short-term shifts in circulation fields, but some complementary approaches may be considered. First, if the slow-varying background properties dominate over the short-term mesoscale induced changes the series should exhibit autocorrelation. Second, if the series are representative of the regional background hydrographical variability there should be correlation (synchronous variation) between the three different sections at the same levels and across large portions of the water column. Finally, an external estimate of the uncertainty caused by the sampling rate can be evaluated by taking a low-passed version of a higher frequency float-based product. The procedure will be explained in detail in Sect. 4.2.2.

3.2 Ocean velocities and water mass transport estimates

Referencing geostrophic velocities to a level of no motion in the ocean has always been a problematic task; however, the most objective way of estimating this level is by direct measurement of the flow (*Pickart & Lindstrom, 1994; Rubio et al., 2009*). *Oliveira et al. (2004)* suggested that “the best approach to study the circulation in the West Iberia Coastal Transition Zone is to use direct current measurements in conjunction with the hydrographic sampling”. Here, we will compute the transports directly from LADCP record, from the classic geostrophy approach and we will use hybrid method based on LADCP bottom-tracking to reference geostrophic fields following the study carried by *Comas-Rodriguez et al. (2010)* along the transatlantic section 24.5°N. Bottom tracking provides accurate velocity measurements near the bottom, permitting high-resolution spatial studies that can reveal circulation patterns of deep waters not observable by other way.

3.2.1 Direct currents from LADCP

LADCP data was processed using a re-implementation software of the velocity inversion method originally developed at the Columbia University - Lamont-Doherty Earth Observatory (LDEO) by Martin Visbeck (*Fischer & Visbeck, 1993*) and primarily maintained by Gerd Krahnman (IFM/GEOMAR, Kiel). Near ocean surface velocities provided by the SADCP and near ocean bottom velocities provided by the LADCP bottom-track were included when available as boundary conditions for data processing at every hydrographic station reducing velocity errors (*Visbeck, 2002*). Transport from LADCP measurement was calculated integrating the product of (in-situ velocities) × (the area bounded by the bin length) × (the sum of half the distance to the two neighbour stations), from top to bottom in the water column.

Tidal currents need to be removed from LADCP measurements in order to get transports. The Oregon State University (OSU) Tidal Inversion Software (OTIS) was used to predict the tidal barotropic component, calculated using the OSU TOPEX/POSEIDON global tidal model TPXO¹ (*Egbert et al., 1994; Egbert & Erofeeva, 2002*), which fits the Laplace Tidal Equations and the along track averaged data from TOPEX/POSEIDON and Jason satellites by a least-square adjustment. Time used for the prediction was the mid-time of the station. Tidal

¹<http://volkov.oce.orst.edu/tides/global.html>

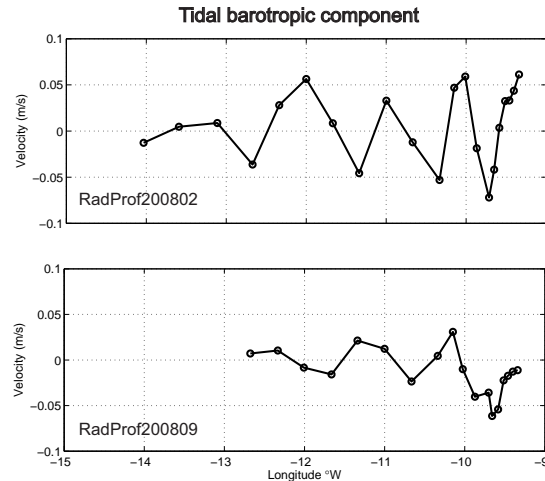


Figure 3.3: Meridional (v) barotropic component of tide predicted by the OTIS model for the Finisterre section during (a) February and (b) September 2008

barotropic component was calculated by averaging the predicted values for all times and locations provided by the ship navigational data during each station (Fig. 3.3).

Meridional barotropic tidal component was subtracted from the corresponding LADCP, constrained by SADC and BT profiles in order to remove the effects that tides have on-situ measurements. Predicted tide values range from -0.05 to 0.05 ms^{-1} , that means of the order of half the current velocities measured in the region, suggesting tides as an important source of noise when calculating transports in a region of a relatively slow circulation.

3.2.2 Geostrophic fields

CTD measurements were used for computing the geostrophic velocity profiles, mass, heat and salt transports between station pairs, using the Matlab software developed by Paul E. Robbins (1995) at the Woods Hole Oceanographic Institution (WHOI). The level of no motion used was $\gamma_n \sim 28.072 \text{ kg m}^{-3}$ ($\sim 3000 \text{ m}$) following the study performed by [Ganachaud \(2003\)](#) in the North Atlantic. When one or both stations of each station pair is shallower than this level, the deepest common level was used as the no movement reference level (e.g. for station pairs near the slope region, st. pairs 2–1 to 9–8, and for station pairs surrounding the Galicia Bank, st. pairs 13–12 to 18–17). Table 3.1 shows reference levels used for

geostrophic profile calculation between each station pair. Configuration of software is such that geostrophic velocity is set to zero below the deepest common level, that is presumably the deepest sampled level from which we can get reliable information.

Mass transports were calculated across the Finisterre section and separately across the slope region (from the coast to 10.2°W) and outer ocean region (from 10.2°W to 12.7°W , see Fig. 2.3). The three outermost stations were removed in RadProf200802 in order to have the same sampled stations than in RadProf200809. The water column was divided into 12 neutral density layers defining water masses following *Fraile-Nuez et al. (2008)*; associated pressure ranges were obtained specifically for our data set and shown in Table 3.2).

Table 3.2: Neutral density (γ_n) layers defining water masses with corresponding pressure ranges (P)

Layer	$\gamma_n \text{kg m}^{-3}$	Water mass	P (dbar)
1	surface – 27.162	Upper layer	surface– 260
2	27.162 – 27.271	ENACW	260–480
3	27.271 – 27.38	ENACW	480–610
4	27.38 – 27.62	UMW	610–930
5	27.62 – 27.82	MW	930–1360
6	27.82 – 27.922	DMW/LSW	1360–1720
7	27.922 – 27.975	LSW	1720–2050
8	27.975 – 28.008	ENADW	2050–2290
9	28.008 – 28.044	ENADW	2290–2610
10	28.044 – 28.072	ENADW	2610–2970
11	28.072 – 28.0986	ENADW	2970–3780
12	28.0986 – bottom	LDW	3780–bottom

The first layer corresponds to the upper ocean layer, from surface to 27.162kg m^{-3} , reaching approximately down to maximum winter mixed layer depths; the next two layers correspond to the upper thermocline waters (from 27.162 to 27.38kg m^{-3}), including the Eastern North Atlantic Central Water (ENACW core centered at 350 dbar) and the Salinity Minimum at ~ 500 m. Next four layers correspond to the lower thermocline waters (intermediate waters): the first three layers are associated with the Upper Mediterranean Water (UMW, 27.38 - 27.62kg m^{-3}), Mediterranean Water (MW) core at 1000 dbar, 27.62 - 27.82kg m^{-3}), and the Deep Mediterranean Water - Labrador Sea Water interface (DMW/LSW, 27.82 - 27.922kg m^{-3}).

Table 3.1: Neutral density (γ_n) reference used as the level of no motion for geostrophic calculations between each station pair during RadProf200802 and RadProf200809 cruises.

Region	Station pair	$\gamma_n(\text{kg m}^{-3})$	
		RadProf200802	RadProf200809
Open ocean	21–20	28.072	–
	20–19	28.072	–
	19–18	28.072	–
	18–17	28.007	28.024
	17–16	27.915	27.930
	16–15	27.910	27.925
	15–14	27.966	27.974
	14–13	28.012	28.029
	13–12	28.015	28.037
	12–11	28.072	28.072
Slope	11–10	28.072	28.072
	10–9	28.072	28.072
	9–8	28.004	28.026
	8–7	28.012	27.996
	7–6	27.860	27.834
	6–5	27.689	27.725
	5–4	26.976	27.148
	4–3	26.963	27.143
	3–2	26.932	27.100
	2–1	26.857	27.016

Below, LSW core (~ 1800 dbar) is associated to the $27.922\text{-}27.975\text{kgm}^{-3}$ layer. Next four layers (from 27.975 to 28.0986kgm^{-3}) associated with the Eastern North Atlantic Deep Waters and Lower Deep Water (LDW, bottom water denser than >28.0986).

Ekman transports were computed across the Finisterre section from climatological wind stress fields of the Scatterometer Climatology of Ocean Winds (SCOW) of the QuikSCAT scatterometer data (Oregon State University ², and added to the Layer 1 mass transport (Fig. 3.2). Climatological meridional Ekman transports resulted -0.047 Sv for February (open ocean: -0.049 Sv; slope: 0.002 Sv) and -0.044 Sv for September (open ocean: -0.045 Sv; slope: 0.001 Sv). That is, two orders of magnitude less than meridional mass transports computed for the 0-500 m layer; this means that the main contribution to the Ekman transport seasonality in the Finisterre region is provided by its zonal component as expected, in response to the prevalence of southwesterly/northerly winds during winter/summer.

3.2.3 Hybrid approach: Geostrophic velocities corrected by LADCP bottom-tracking

Every geostrophic velocity profile was corrected using bottom track data from the two neighbour stations following the method previously applied by *Cunningham et al.* (2003) at the Drake Passage and *Comas-Rodriguez et al.* (2010) along the 24.5°N transect. The motivation for doing so is that LADCP at mid-depths is less reliable at oceanic areas where backscatter signal is weak. Differences between geostrophic and bottom-track profiles were obtained at the corresponding depths and averaged for both stations to obtain a final mean. In the case of station pairs where LADCP bottom-track data is only available for one of the two stations, this was taken for correction. An *offset* is thus available for every geostrophic profile providing the corresponding *reference velocity*. Offsets applied for referencing geostrophy are shown in Fig. 3.4. Generally, larger correction will be applied at the slope region (eastward from 10°W), reaching up to ~ 0.1 m/s.

A comparison can be made between pairs of LADCP profiles (including SADCP and BT), the corresponding geostrophic velocity profile initially calculated between them, and the final BT referenced geostrophic profile. Results for a pair of stations located between the coast and the east flank of Galicia Bank (st. 11, 43°N , 10.33°W and st. 10, 43°N , 10.15°W) during RadProf200802 are shown in

²<http://cioss.coas.oregonstate.edu/scow/index.html>

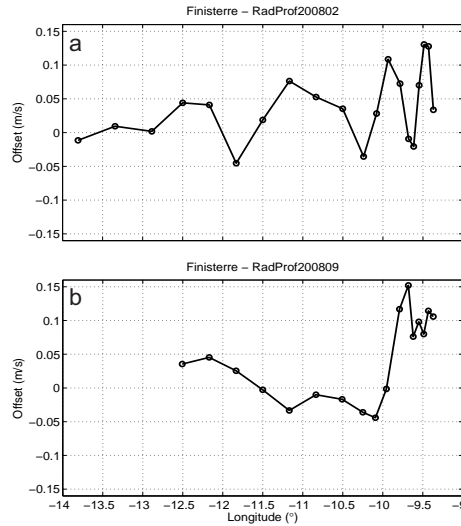


Figure 3.4: Velocity offsets applied for geostrophy correction during a) Rad-Prof200802 and b) RadProf200809.

Figure 3.5. In this case, the level of no motion is set at $\gamma_n=28.072$ (~ 3000 dbar). Note that the geostrophic profile is constrained by the two neighbour bottom-track adding a *different-from-zero* reference level for velocity of ~ -0.04 m/s .

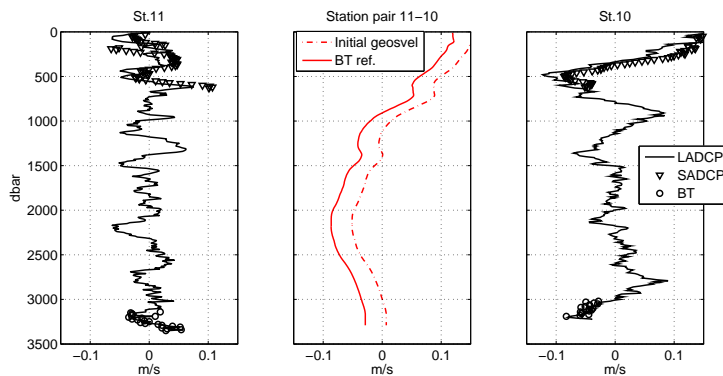


Figure 3.5: Comparison between LADCP velocity profiles (including SADCP and bottom-track (BT)) at stations 10 (43°N , 10.15°W) and 11 (43°N , 10.33°W), initial calculated geostrophic profile between them and the corresponding BT referenced geostrophic profile.

CHAPTER 4

Results

4.1 Seasonality of the deep ocean off the northwest Iberia

Vertical sections of salinity from the cruises series (Fig. 4.1) provide a first insight on the presence of a recurrent seasonal pattern in the structure of hydrographic fields, specially evident at the levels of Mediterranean water characterized by the salinity maximum (core c.a. ~ 1000 m). As expected from Fig. 2.3, a saltier MW vein ($S \sim 36.2$) is tight against the slope year-round, but most occupations (though not all) show that this signature of MW on the slope tends to be stronger in summer, while in wintertime the MW core tends to be broader and displaced offshore, even concentrating in the surroundings of the Galicia Bank. The recurrence of structures in winter and in summer points to the existence of a seasonal cycle at depth. Results about the processes behind this seasonality will be discussed later in Sect. 5.1.

4.1.1 Quantification, zonal structure and nature of seasonal changes

There are basically two methods for extracting the seasonal cycle from a time series (see for instance *Chelton, 1982*). The first consists of carrying out a harmonic analysis of the data and reconstructing the seasonal cycle from the amplitudes and phases of the annual and higher order constituents (e.g. *Bray, 1982*). The second consists of computing the long-term mean value through the calendar year at the sampling interval of the time series (e.g. *Thomson et al., 1985*). The latter ap-

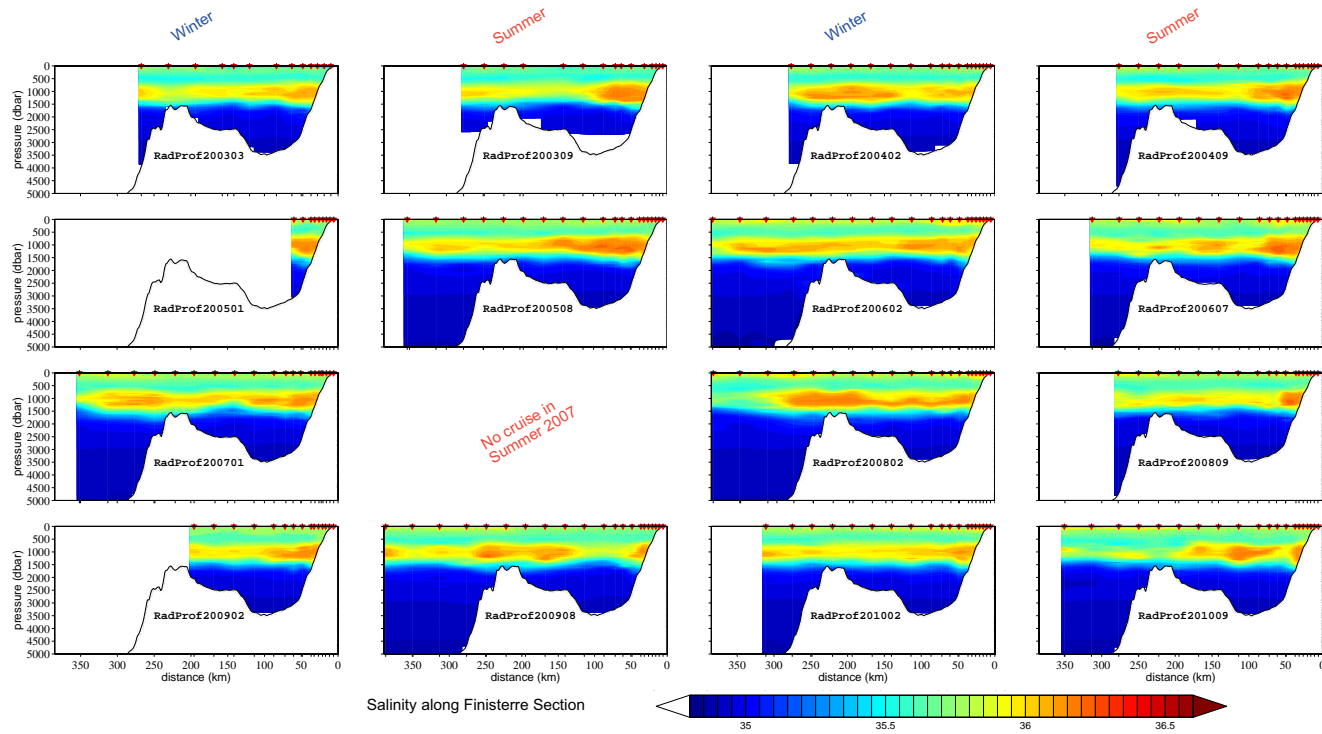


Figure 4.1: Plate showing the salinity field structure for the 15 repetitions of the Finisterre section. Sections are showed in chronological order from left to right, with the first and third columns showing the winter cruises and the second and fourth columns showing the summer cruises, except that of summer 2007.

proach is generally preferred as it filters the slow-varying interannual variability and a seasonal cycle of arbitrary shape is not forced to accommodate to a reconstruction by few harmonics. On the other hand, this method may incorporate a strong bias if any specific section were strongly anomalous. In the present section we take the summer-average vs the winter-average of properties as the measure of their seasonal amplitude. Actually, our semiannual time series does not provide information to characterize the seasonality signature further. We will apply in Sect. 4.1.2 a simple harmonic analysis to key time series of the record in order to get a statistical measure of the robustness of the seasonality signal.

Figure 4.2 shows the seasonal changes of temperature, salinity, and density along the sections. Note that the anomalies are representing the summer minus winter condition, i.e. positive anomalies (red shades) indicate warmer, saltier or denser water along isobars or deeper isoneutrals during summertime. Here, the pattern previously inferred from Fig. 4.1 is clearly perceptible, with warmer and saltier waters from 1000 to 1750 m tight against the slope in summertime (from coast to about 10° W), versus colder and fresher from 11.5° W to 13.5° W over the Galicia Bank. This implies a warmer and saltier intermediate water broadening and spreading above the Galicia Bank in wintertime. Also clearly identifiable is the seasonal effect on the upper layers with warmer waters down to 100 dbar in summertime except over the shelf, where the effect of summer upwelling is notable. There is also a change in the density field structure with lighter surface waters (down to ~ 100 dbar) in summer than in winter, and denser water below; the near-surface (~ 100 dbar) zero contour separating depths where the sign of the anomaly changes, shoals near the coast suggesting the presence of denser (and cooler) waters that are uplifted from the thermocline during summer upwelling. Changes in the density field (hence in pressure of density surfaces) accounts for the heave component of the isobaric changes in the thermohaline properties, implying changes in the geostrophic flow patterns in the ocean interior, as we will see later.

Right column of Fig. 4.2 shows the summer minus winter mean differences of θ , S , and pressure of γ^t surfaces, providing a graphical view of where properties on isopycnals change and where changes due to heave are of importance. Main results that can be drawn are (i) the strong isopycnal character of changes at the lower bound of Mediterranean waters, tight on the slope and west of the Galicia Bank. (ii) The high seasonal differences on isopycnal pressure levels above isopycnal 27.3 (~ 500 dbar) all along the section, implying an isopycnal shoaling up to 50 dbar in summer with respect to the winter season. Such change is much pronounced on the slope and extends deeper. (iii) An apparent change in the struc-

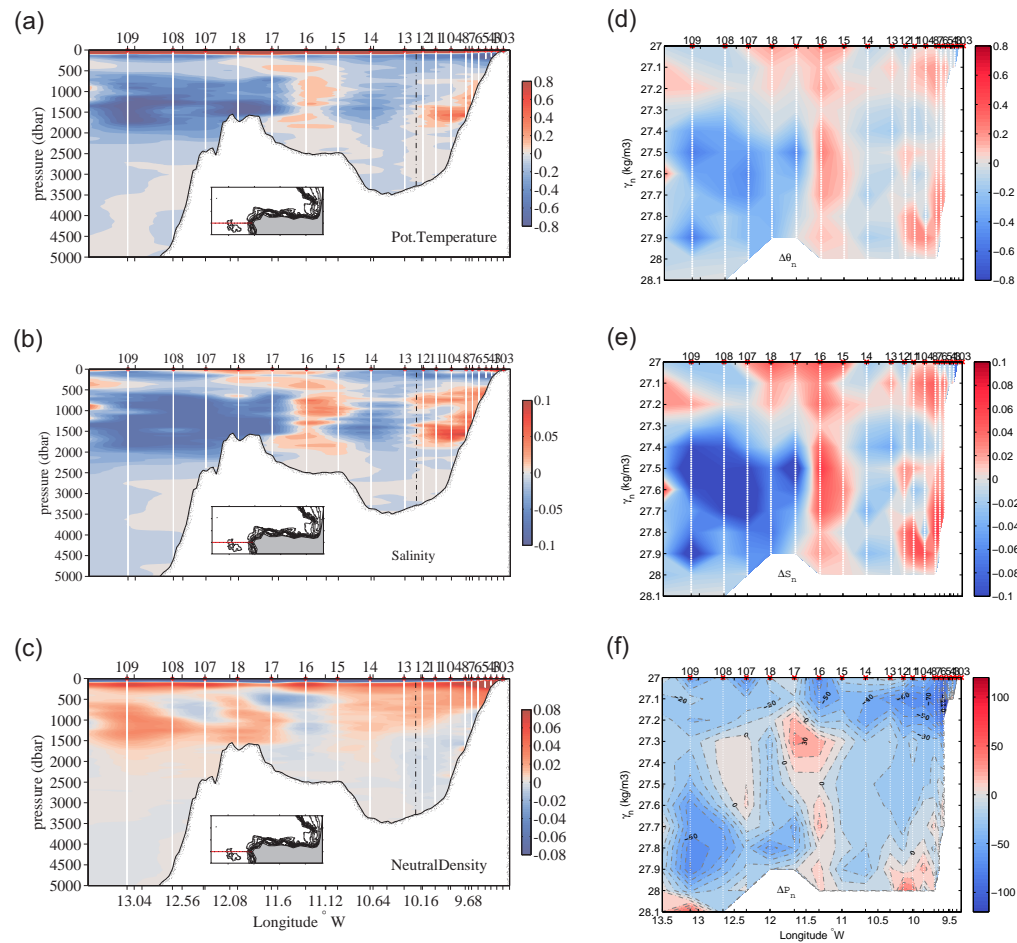


Figure 4.2: Left column: summer–winter differences of **(a)** potential temperature, **(b)** salinity and **(c)** neutral density along isobaric surfaces. Right column: summer–winter differences of **(d)** potential temperature and **(e)** salinity along isoneutral surfaces with associated summer–winter differences of **(f)** isoneutral pressure.

ture of the density field around the Galicia Bank, with the sinking of isopycnals at its eastern flank in summertime at intermediate levels. This readjustment of the density field would imply an enhancement of geostrophic southwards flow in the passage from the Galicia Bank to the shelf during this season. Finally, (iv) there is an enhancement of a recirculation structure associated to the Galicia Bank in summertime.

It is interesting to get overall estimates of the amplitudes of annual cycles at depth. By inspection of Figs. 4.2a, b and even e, it looks rather like there are four distinct regions, one on the slope, one in the channel between the continent and the Galicia Bank, one east of the Bank and one west of it. These four regions appear to be linked to the recirculation system that develops in the surroundings of the Galicia Bank, which seems to yield a differentiated response to the seasonally varying background flows. We could easily apply the methodology of splitting isobaric changes that latter generate Fig. 4.3 to the four regions instead of the two proposed. The reason for keeping only two regions is that some of the subregions are narrow and involve very few stations. For instance, the anomaly at the eastern flank associated with warmer/saltier waters in summertime is confined to stations 15–17. Therefore, we prefer to keep a larger number of stations per region in order to get more reliable results from a statistical point of view, while keeping in mind that there are biases in the thermohaline seasonal signature in the outer and inner regions of the Bank due to its influence on the dynamics. Our approach assumes that, for the purpose of providing an overall view of the hydrographical variability of the outer ocean, the local anomalies caused by the Bank circulation should cancel each other out.

Thus, we consider separately the two regions (slope and outer ocean) and the methodology for splitting the changes, as in Eqs. (3.1) and (3.2), is applied at isopycnal levels from 27 to 28 kg m⁻³. Then, amplitudes of the annual cycle for each of the terms in the decomposition relationship are obtained for every neutral density layer and seasonal anomalies are zonally averaged for each region.

A graphical view of the computation is shown in Fig. 4.3, which provides a continuous profile of seasonal amplitudes, while the numerical values are shown in Table 4.1 as anomalies at the pressure levels of key isopycnals.

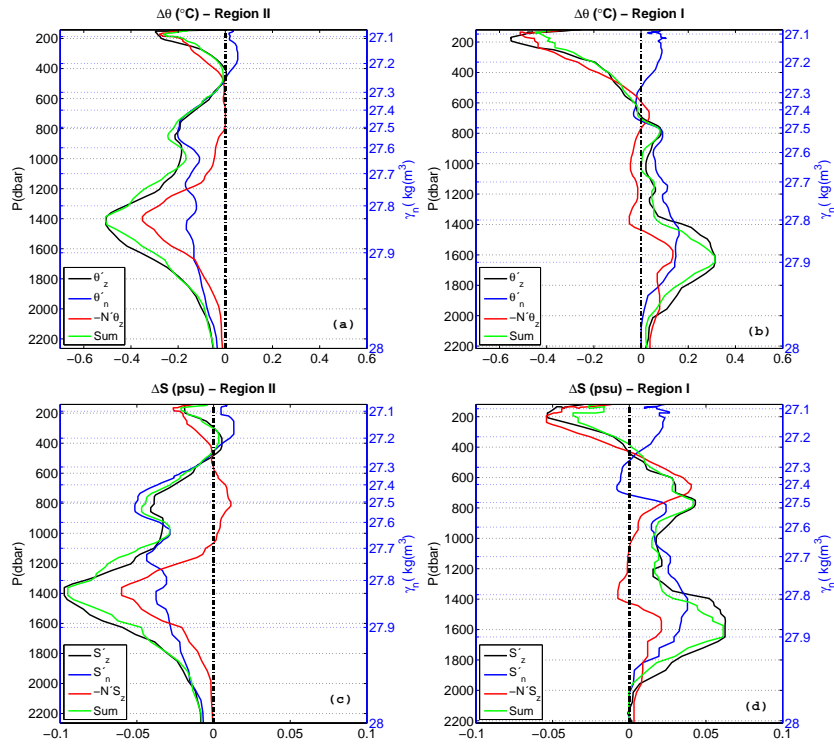


Figure 4.3: Decomposition of θ (top) and S (bottom) isobaric seasonal cycle following Eqs. (3.1) and (3.2). **(a)** and **(c)** are for the open ocean and **(b)** and **(d)** for the slope. Note that the decomposition is only approximate; the difference between the isobaric change (black line) and the sum (green line) of the isopycnal (blue line) and heave (red line) terms indicates the inaccuracy of the decomposition.

Table 4.1: Decomposition of θ and S seasonal changes along isoneutral surfaces (γ^n) as isopycnal plus heave terms in the open ocean and shelf-slope region. Positive (negative) values indicate warmer/saltier (cooler/fresher) conditions in summertime. P_n is the mean pressure of isoneutral γ^n (see Fig. 4.3).

γ^n	P_n (dbar)	θ'_z	θ'_n	$N'\theta_z$	$\theta'_n - N'\theta_z$	S'_z	S'_n	$N'S_z$	$S'_n - N'S_z$
Open ocean									
27	142	-0.09	0.02	-0.19	-0.16	-0.017	0.007	-0.013	-0.004
27.1	187	-0.27	0.01	-0.15	-0.16	-0.018	0.001	-0.016	-0.016
27.2	367	-0.01	0.04	-0.06	-0.01	0.005	0.010	-0.006	0.004
27.3	558	-0.05	-0.04	-0.01	-0.05	-0.011	-0.012	-0.000	-0.011
27.4	677	-0.13	-0.14	-0.00	-0.14	-0.025	-0.036	0.007	-0.028
27.5	791	-0.19	-0.19	0.00	-0.22	-0.038	-0.050	0.011	-0.044
27.6	927	-0.18	-0.13	-0.04	-0.18	-0.032	-0.035	0.004	-0.031
27.7	1100	-0.21	-0.14	-0.08	-0.25	-0.038	-0.038	-0.005	-0.048
27.8	1314	-0.43	-0.11	-0.29	-0.42	-0.084	-0.029	-0.048	-0.080
27.9	1627	-0.32	-0.12	-0.11	-0.24	-0.059	-0.026	-0.018	-0.045
28	2264	-0.05	-0.03	-0.01	-0.04	-0.008	-0.006	0.000	-0.007
Shelf-slope									
27	119	-0.27	0.09	-0.33	-0.37	-0.029	0.024	-0.011	-0.021
27.1	147	-0.54	0.09	-0.50	-0.39	-0.048	0.026	-0.044	-0.016
27.2	332	-0.17	0.06	-0.24	-0.18	-0.010	0.016	-0.025	-0.009
27.3	533	-0.07	-0.01	-0.03	-0.06	0.011	-0.004	0.021	0.013
27.4	645	-0.01	-0.02	0.03	-0.01	0.029	-0.007	0.040	0.029
27.5	764	0.08	0.08	0.00	0.07	0.043	0.021	0.022	0.044
27.6	926	0.03	0.04	-0.03	0.01	0.023	0.013	0.005	0.017
27.7	1120	0.04	0.09	-0.01	0.06	0.020	0.026	-0.001	0.021
27.8	1371	0.14	0.13	-0.05	0.07	0.039	0.036	-0.009	0.024
27.9	1649	0.31	0.15	0.10	0.31	0.062	0.033	0.017	0.062
28	2215	0.03	0.00	0.03	0.03	0.002	0.000	0.001	0.000

The main results of the analysis can be summarized as follows:

- Waters just below the seasonal thermocline cool and freshen in summertime all along the section, with maximum changes at ~ 200 dbar ($\gamma^n \sim 27.1$). At the open ocean, maximum cooling/freshening values exceed ~ 0.3 °C and ~ 0.02 , and this seasonality decays progressively down to ~ 500 dbar (the salinity minimum at the base of ENACW) where there is no signature at all. The process is driven mainly by heave. On the slope the behaviour is similar but the amplitudes are larger and seasonality extends deeper down the water column. In both areas there is a weaker (than heave) isopycnal warming/salting signature that cannot counteract the much greater cool-

ing/freshening by heave.

- In the density range $\gamma^{\theta} \sim 27.3\text{--}27.6$ (500–900 dbar; upper MW) we observe cooling/freshening at the outer ocean (0.2°C and 0.03) and salinification on the slope (0.05). Isopycnal change drives the cooling/freshening at the outer ocean and there is a slight salt increase centred at 27.45 (note that as this water layer is characterized by a weak or null thermal gradient – see Fig. 2.3c, d – it is not possible to achieve warming/cooling by heave). The salinity increase on the slope is caused by the strong uprising of isopycnals in this region (i.e. heave).
- The lower layer of MW down to the LSW (range $\gamma^{\theta} \sim 27.7\text{--}27.95$, 1100–1800 dbar) is the portion of the water column where the seasonal cycle is greatest. At the outer ocean there is strong cooling/freshening up to $0.5^{\circ}\text{C}/0.1$, caused by the combination of isopycnal change and heave, the latter being dominant at 1400 dbar where the amplitude of seasonality peaks. On the contrary the warming and salting on the slope (up to 0.31°C and 0.062 near 27.9) is strongly dominated by isopycnal change. Maximum values for the outer region may seem higher in Fig. 4.3 than those shown in Table 4.1; the reason is that Fig. 4.3 provides a continuous profile of seasonal amplitudes, while Table 4.1 extracts the values at the pressure levels of key isopycnals.

In summary, there exists a noticeable seasonality signature in the Finisterre section for the whole permanent thermocline, down to c.a. 2000 m. There is a notable asymmetry between the slope region and the outer ocean. ENACW properties oscillates by heave with stronger amplitude on the slope. Upper MW on the slope appears to be upraised in summertime, while at the outer ocean its signature weakens (cools and freshens without significant heave signature). The water layer corresponding to the lower MW exhibits the largest changes, with strong cooling/freshening at the outer ocean and warming/salting on the slope. Overall, it appears that MW during summertime gets tightly attached against the slope and uplifted, reinforcing there its thermohaline signature and losing its presence at the outer ocean. In wintertime the situation reverses, MW seems to detach from the slope and spreads out to the open ocean, even developing a secondary branch at, and west of, the Galician Bank.

4.1.2 Statistical significance

In this section we will check the robustness of the existence of seasonality from a statistical point of view. Testing for seasonality (or whatever periodic variation) in a time series relies in principle on spectral analysis, however, having a very limited number of observations and a signal with strong interannual variability (nonlinear trends) such seasonality testing is not straightforward. As previously noted, a convenient approach is to apply a simplified harmonic analysis determining the best fit of time series from hydrographical properties to the functional form:

$$\psi(t) = a \cos(\omega t + \phi) = a_1 \cos(\omega t) + a_2 \sin(\omega t), \quad (4.1)$$

where $\omega = 2\pi f$ is fixed to one year ($f = 1$, which is indeed the shortest solvable scale given the Nyquist frequency of our semiannual series). Then, it can be checked whether the amplitude coefficient is significant by means of a t test. This approach was followed by [Bray \(1980\)](#) to determine seasonality of hydrographical properties at intermediate levels in the Bay of Biscay, which is a very similar problem to ours, though her dataset consisted of 11 points along three years. Prior to attempt any harmonic analysis it is needed to remove the record mean and trends (linear or not) if relevant (e.g. [Emery & Thomson, 2001](#)). This is indeed our case given that, as will be shown next, interannual variability is much larger than the apparent seasonality signal. Therefore, we treat our 8 yr time series of hydrographical properties subtracting firstly a two-degree polynomial (hence removing slowly varying patterns) to compute later the bestfit of Eq. (4.1) of the residual.

Figure 4.4 shows the bestfit of properties (salinity at isobars, pressure of isopycnals and salinity at isopycnals) at levels of selected neutral surfaces $\gamma^{\eta} = 27.4$ and $\gamma^{\eta} = 27.8$ for the slope region, where we have determined seasonality driven by heave and by isopycnal change respectively (Fig. 4.3). In accord with what is observed in the sections (Fig. 4.1), a see-saw pattern is intermittently present at portions of the time series. After the removal of the trend, the fitting results are significant for heave at 27.4 and for isopycnal change at 27.8. Anomalous years with opposed patterns also arise (e.g. summer 2009 again in agreement with Fig. 4.1). Figure 4.4 provides us an insight of the relative strength of the seasonality with respect to the interannual variability. Values of the estimated amplitude of seasonality represent roughly a 20 % of maximum variations or excursions along the whole time series of properties, interannual variability included (Fig. 4.4 and others not shown). In summary, this analysis reinforces the case for the existence of seasonality superimposed on the interannual variability of hydrographic properties at depth.

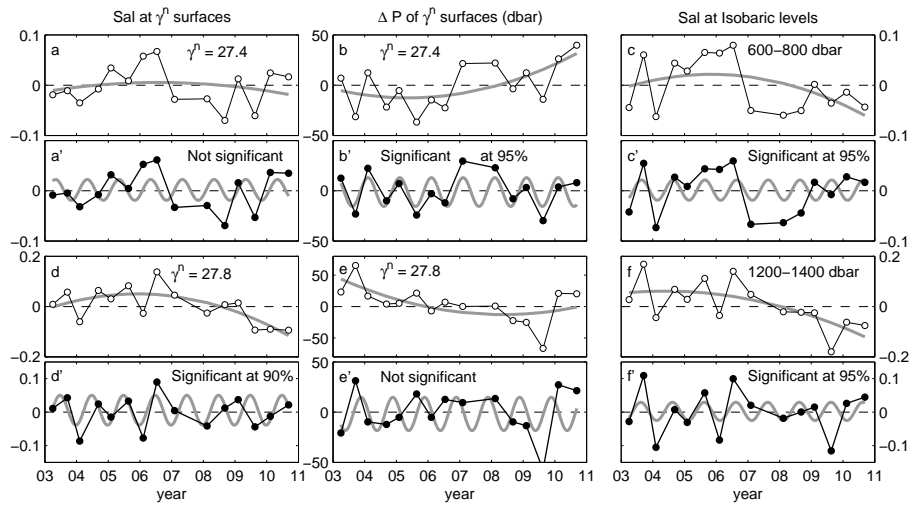


Figure 4.4: Time series of (a), (d) salinity anomaly at isopycnal levels $\gamma^{\rho} = 27.4$ and 27.8, (b), (e) pressure (anomaly) at which these isopycnal levels lie, (c), (f) salinity anomaly at isobaric levels 600–800 and 1200–1400 dbar, roughly corresponding to these isopycnals. A two-degree polynomial trend is computed for each series (thick gray). Prime versions of figures correspond to the simple harmonic analysis applied to the residual (time series minus trend, black dots) following Eq. (4.1).

4.2 Variations of thermohaline properties at inter-annual scales

The next step is to deal with the deseasonalized signals representing the interannual changes of water mass properties. Results from the present Section will be discussed later in Sect. 5.2.

4.2.1 Raw timeseries at isobaric levels. Overall trends

A first insight of the evolution of thermohaline properties of water masses with time is provided by the θ/S diagrams in Fig. 2.1c. Note the isopycnal shift of central waters towards a warmer/more saline type at all three sections, seen here as the displacement of the straight line between 27.1 and 27.3 in the θ/S diagram. At deeper water below about $\sigma_\theta \sim 27.4$ the diagrams show no clear changes in properties. Figure 4.5 shows average temperature and salinity timeseries at isobaric levels in order to emphasise interannual changes at the three sections. There are some coherent pattern in changes between the sections and through the water column, as well as notable year to year changes. Linear trends with confidence intervals derived from a T-test at all layers are shown in Fig. 4.6.

Changes in the upper layer (surface to the maximum development of winter mixed layers, 200 dbar) are not well sampled by semiannual sections but the imprint of the cold winters 2005, 2006 and 2009 clearly emerges. The surface salinity is characterized by two periods of smooth increase either side of the cold, fresh winter of 2009. The upper ENACW (200-400 dbar) shows a salinity rise of about 0.5 during 2005 and 2006 and remained pretty stable since then, yielding an overall salt-increase trend of 0.006yr^{-1} . A $0.30\text{ }^\circ\text{C}$ decrease in potential temperature in the 200-400 dbar layer at Finisterre indicates that the strong cooling in winter 2005 was able to penetrate deep in the water column. The effect was transitory since this layer and 400-800 dbar (the Salinity Minimum at the base of ENACW to the upper MW) show a significant warming trend of $0.018 - 0.016\text{ }^\circ\text{C yr}^{-1}$ over the time series. Lower thermocline waters (27.5-27.9, 800-2000 dbar including MW and LSW) are characterized by a drop in temperature and salinity in late 2008 and a recovery in 2011. The 1200-1600 dbar layer is the only part of the water column with overall significant negative trends in thermohaline properties. The deep water (28-28.1, ~ 2000 dbar to > 4000 dbar) showed much weaker interannual variations. A freshening of >3600 dbar water (including bottom water) of about 0.004 was observed along the three transects from late 2009 to late 2012. There

were coherent variation across most of the water column in 2008 and 2013 (cooling and decreasing salinity) and 2011 (warming and increasing salinity). Next we examine the robustness of the signals described above by considering the effect of the sampling frequency on the results.

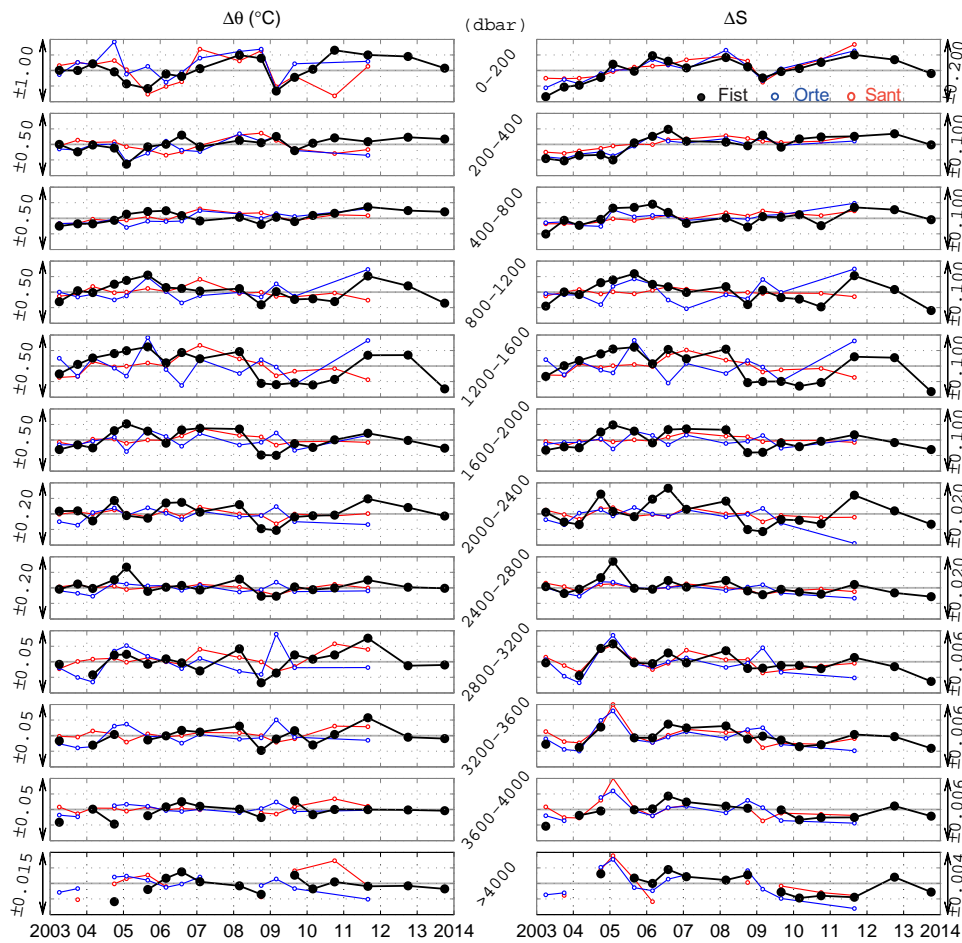


Figure 4.5: Potential temperature (left) and salinity (right) interannual variability at Finisterre (black), Ortegal (blue) and Santander (red) during 2003-2013. Changes were averaged over pressure intervals from surface to 200 dbar and then every 400 dbar down to 4000 dbar. Abyssal water (>4000 dbar) changes were averaged from 4000 dbar to the bottom.

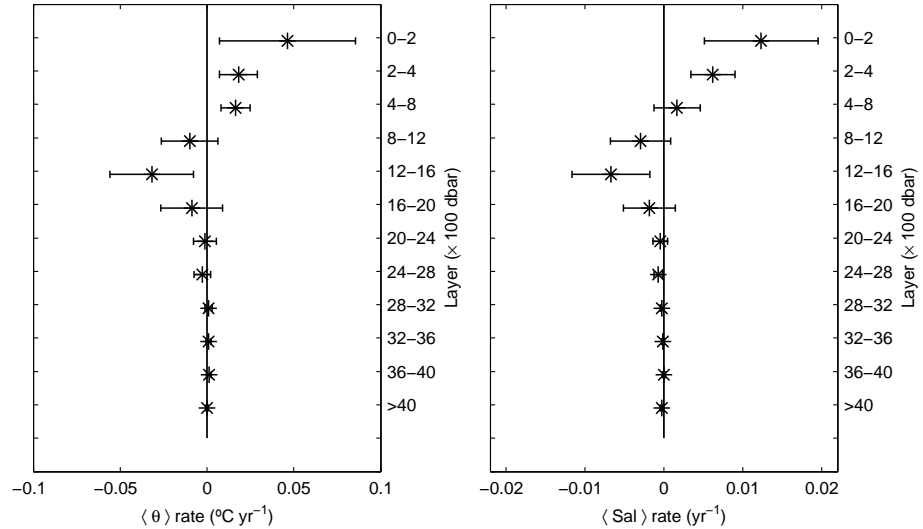


Figure 4.6: Potential temperature (left) and salinity (right) trends for series at Finisterre during 2003-2013 as shown in Fig. 4.5

4.2.2 Consistency of the record. Correlation, autocorrelation and mesoscale field

A well acknowledged problem when describing the state and the evolution of ocean hydrography from repeated hydrographic sections is to identify the uncertainty induced by the mesoscale field. Mesoscale variability in the region exhibit length scales of about 50-100 km diameter and decay scales of $\sim 250 \pm 100$ days (Pingree, 1994) that may be a source of noise for our semiannual time series specially at the slope (Pingree & Le Cann, 1992a,b) and MW levels (Richardson et al., 2000). Eddy-like structures are evident in the sections (not shown) and, accordingly, an estimate of correlation scale by a bestfit of salinity fields through an isotropic Gaussian function of the distance between the observations (e.g. Bretherton et al., 1976) yielded a decay scale of $\sim 10 - 30$ km. Average station spacing is about 25 km (closer in the shelf-break and increasing to 37 km at the outer ocean) so the mesoscale field should be captured by the sampling scheme. In this section we use a statistical approach to infer parts of the water column that may be dominated by mesoscale variability.

If the short-term variability induced by the mesoscale field were greater than or comparable to the large scale interannual variability, we would expect the se-

ries to be no different to noise and have no autocorrelation. To detect the presence of autocorrelation in the record (that is, to test the existence of statistically significant temporal coherence of variability), we applied a Durbin-Watson test (e.g. *von Storch & Zwiers, 2001*) to the series on isobars (Fig. 4.5). Table 4.2 show the p-values; most surface to 2000 dbar Finisterre and Santander series indicate autocorrelation (above 90% and 95% for most series) which suggests that series are dominated by the large scale background signal. The Ortegual series fail the autocorrelation tests indicating that these series are more strongly affected by local short-term circulation and the mesoscale field. Deeper layers do neither show autocorrelation but not relevant trends or shifts have occurred at great depths.

Table 4.2: p-values from Durbin-Watson test. Test of autocorrelation was applied to series at isobaric levels shown on Fig. 4.5. Zero values indicate $p \leq 10^{-3}$

P(dbar)	Finisterre		Santander		Ortegual	
	$\Delta\theta(^{\circ})C$	ΔS	$\Delta\theta$	ΔS	$\Delta\theta$	ΔS
0-200	0.04	0.00	0.17	0.00	0.80	0.07
200-400	0.95	0.00	0.00	0.00	0.23	0.01
400-800	0.00	0.02	0.01	0.80	0.93	0.24
800-1200	0.08	0.12	0.09	0.02	0.62	0.12
1200-1600	0.01	0.02	0.00	0.00	0.49	0.83
1600-2000	0.01	0.00	0.01	0.00	0.08	0.26

If the mesoscale field were dominant in a section we would expect that properties would only correlate vertically at the scale of eddies (within a single 400 dbar layer or two contiguous layers at most). In contrast, if changes were caused by shifts in large-scale advective patterns there may be correlation across a greater pressure range. Correlation coefficients (R) with associated p-values (p) were calculated to test the hypothesis that there is no correlation between the isobaric levels shown in Fig 4.5. Table 4.3 shows correlation coefficients and p-values for potential temperature and salinity anomalies along the Finisterre, Ortegual and Santander sections. The main outcome is that potential temperature and salinity anomalies are highly correlated (from 70% to 90%) across the lower thermocline (from 800 to 2000 dbar) in the three sections (except salinity anomalies at the Ortegual section with only $\sim 60\%$), suggesting that changes may be more affected by the large-scale. At deep levels below 2400 dbar, correlation of changes is also statistically significant at the 70-90%.

Table 4.3: Correlation coefficients (R) and p-values for potential temperature and salinity anomalies ($\Delta\theta$, ΔS). Anomalies were averaged every 400 dbar along the Finisterre (subscript *Fist.*), Ortegal (*Orte.*) and Santander (*Sant.*) sections.

Pressure layers (dbar)	$\Delta\theta$						ΔS					
	$R_{Fist.}$	$p_{Fist.}$	$R_{Orte.}$	$p_{Orte.}$	$R_{Sant.}$	$p_{Sant.}$	$R_{Fist.}$	$p_{Fist.}$	$R_{Orte.}$	$p_{Orte.}$	$R_{Sant.}$	$p_{Sant.}$
0–200→200–400	0.36	0.13	0.07	0.80	0.66	0.00	0.69	0.00	0.66	0.01	0.79	0.00
200–400→400–800	0.24	0.33	0.26	0.36	0.12	0.67	0.42	0.08	0.42	0.12	0.84	0.00
400–800→800–1200	0.50	0.03	0.44	0.11	0.48	0.06	0.75	0.00	0.72	0.00	-0.15	0.57
800–1200→1200–1600	0.84	0.00	0.77	0.00	0.90	0.00	0.83	0.00	0.56	0.03	0.75	0.00
1200–1600→1600–2000	0.79	0.00	0.71	0.00	0.89	0.00	0.80	0.00	0.61	0.02	0.86	0.00
1600–2000→2000–2400	0.60	0.00	0.59	0.02	0.45	0.08	0.68	0.00	0.35	0.21	0.33	0.22
2000–2400→2400–2800	0.54	0.02	0.67	0.00	0.73	0.00	0.53	0.02	0.69	0.00	0.89	0.00
2400–2800→2800–3200	0.69	0.00	0.92	0.00	0.61	0.01	0.87	0.00	0.92	0.00	0.83	0.00
2800–3200→3200–3600	0.81	0.00	0.93	0.00	0.84	0.00	0.83	0.00	0.95	0.00	0.90	0.00
3200–3600→3600–4000	0.44	0.10	0.91	0.00	0.71	0.00	0.75	0.00	0.96	0.00	0.98	0.00
3600–4000→>4000	0.96	0.00	0.67	0.01	0.87	0.00	0.85	0.00	0.96	0.00	0.94	0.00

Finally we focus on the correlation of the same water masses at the three sections. No significant correlation may indicate the dominance of the eddy-field, but could also indicate different regional signals of water masses variability. There are significant correlations between all three sections in the surface waters and the upper thermocline (200-400 dbar, especially in salinity, Fig. 4.5). R is as high as 80% between Finisterre and Santander, and more than 90% between Finisterre and Ortegal, significant at the 95% confidence level at 0-200 dbar. This correlation stands also for central waters salinity anomalies, 77% for Finisterre-Santander and 93% Finisterre-Ortegal. For these specific layers, stats suggest largescale influence; note that this does not contradict the dominance of mesoscale at Ortegal since the autocorrelation test shows large p -values except for salinity anomalies at 0-200 dbar and 200-400 dbar (Table 4.2)

If the timescale of the mesoscale variability is 250 ± 100 days (Pingree, 1994) we might expect that 1 or 2 sections per year will undersample that temporal variability, leading to biased results. We can make use of an independent, float-based product with monthly resolution to estimate the uncertainty due to the low frequency sampling. The UK Met Office EN4 product (Good *et al.*, 2013) is constructed as objective analysed fields based on observed subsurface ocean temperature and salinity (mainly from float profilers) on monthly basis. We recovered the grid points along 43°N , $9^\circ\text{-}15^\circ\text{W}$ (Finisterre section) for the 2002-2012 period and computed deseasonalized time series of temperature and salinity at the same isobaric levels as our record. The approach is to compare a low-pass filtered version of the time series (representing variability at the time scale we are interested in), to a time series of the same product sub-sampled six-monthly (the rate of our in-situ ocean time series). The standard deviation of the residuals between these two series is taken as representative of the error associated with the semiannual sampling.

The analysis was restricted to the upper 2000 dbar of the water column since there are limited data points below that level (Good *et al.*, 2013). Figure 4.7 shows monthly and six-month averaged time series of salinity variations from the EN4 product at two representative levels (a) 200-400 dbar and (b) 1600-2000 dbar.

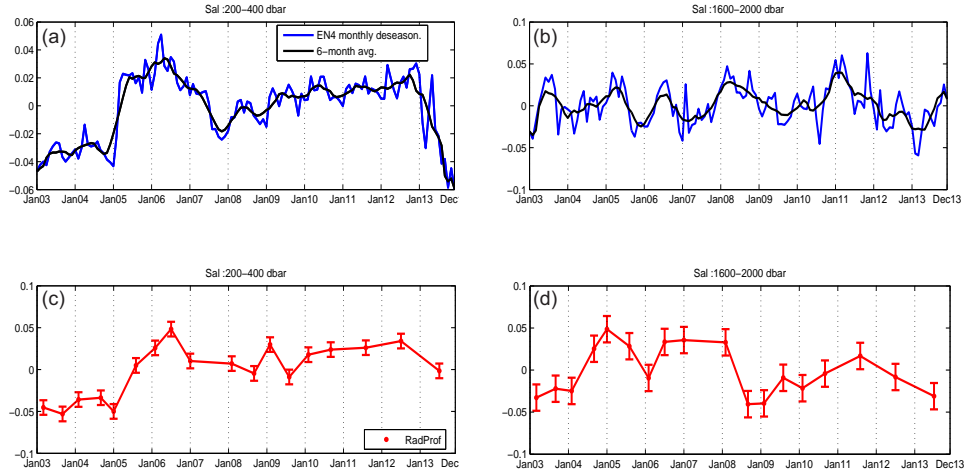


Figure 4.7: Monthly deseasonalized salinity time series from objective analysis (blue line) and 6-month moving average of time series (black line), retrieved from the EN4 UK Met Office product ([Good et al., 2013](#)) for the [43°N, 9-15°W] region at a) 200-400 dbar and b) 1600-2000 dbar. Interannual variability observed along the Finisterre section at the same pressure intervals (c) and d)) with estimated error bars calculated from residuals of the blue and black lines.

Table 4.4: Estimate of short-term variability based on the EN4 product. Estimated short-term variability at the Finisterre section ($\Delta\theta_\varepsilon$, ΔS_ε) compared to the interannual variability exhibited by the section ($\Delta\theta$, ΔS).

ΔP (dbar)	$\Delta\theta_\varepsilon$ (°C)	$\Delta\theta$ (°C)	ΔS_ε	ΔS
0-200	0.32	1	0.015	0.24
200-400	0.05	0.4	0.008	1.2
400-800	0.04	0.2	0.011	0.1
800-1200	0.1	0.4	0.026	0.1
1200-1600	0.15	0.4	0.032	0.1
1600-2000	0.08	0.4	0.016	0.1

Salinity variations observed from the sections with estimated error bars are shown for the same pressure intervals in (c) and (d) respectively. Table 4.4 shows a comparison between errors calculated as deviations of the residuals of EN4 time series ($\Delta\theta_\varepsilon$, ΔS_ε) and the size of maximum anomalies observed in the in-situ time series ($\Delta\theta$, ΔS). Estimated errors are an order of magnitude smaller than the

background signal supporting also the idea that the large scale variability has been identified above mesoscale noise.

4.2.3 Decomposition of interannual changes: isopycnal vs. heave

Having established that large-scale variations can be detected above mesoscale variability in the time series at Finisterre below the surface layer, we next examine the relative influence of source area fluxes vs circulation changes at that section through the decomposition into heave and isopycnal variations. The θ - S relationship and the approximate depths of the key isoneutrals is given in Fig. 4.8a, and decomposition of the property changes on those isoneutrals is shown on Fig. 4.9. Changes due to heave are the product of vertical gradients of properties (Fig. 4.8a') and the temporal variation of isoneutrals pressure (N' , see Fig. 4.8b). Note that the decomposition is only approximate so the difference between the isobaric change (black line) and the sum (green line) of the isopycnal (blue line) and heave (red line) terms accounts for the inaccuracy of the decomposition (Fig. 4.9). Isopycnal changes for salinity and temperature must yield the same functional form (as constrained by density). Changes due to heave differ since they are modulated by their respective reference gradient profiles (Fig. 4.8a') with the main difference of the large salinity inversion in the upper MW levels.

A dominance of isopycnal origin in the variability of central waters (27.1 and 27.2) clearly emerges from Fig. 4.9, particularly in the salinity record. The decade is mostly characterized by the rapid warming and salinification after the winter of 2005. Isonneutral 27.2 shoaled by about 100 m in 2005 causing immediate cooling and freshening, but after that there was a process of isopycnal warming/salinification that ended by late 2006, and the water mass has remained stable in this saltier/warmer state since then. The event in 2005 also affected the isoneutral associated with the Salinity Minimum (27.3, \sim 500-600 dbar). This level had a notable shift in properties after 2011 when the Salinity Minimum was up to 0.15°C warmer and 0.025 saltier than average.

Changes along the isoneutrals of the upper MW to the MW core (27.5-27.7) have an almost exclusively isopycnal origin consisting on warming and salinification since 2003 with maximum in summer 2005 followed by fresher and cooler conditions interrupted at both levels by the abrupt strong warming (0.3-0.4°C) and salinification (0.08-0.1) of isopycnal origin in 2011. Changes on isoneutral 27.9 (LSW) show isopycnal change and heave during the sampling period. Warming/salinification by isopycnal change was observed again from 2003 to 2005, after which warming/salinification by heave was caused by the deepening of isoneu-

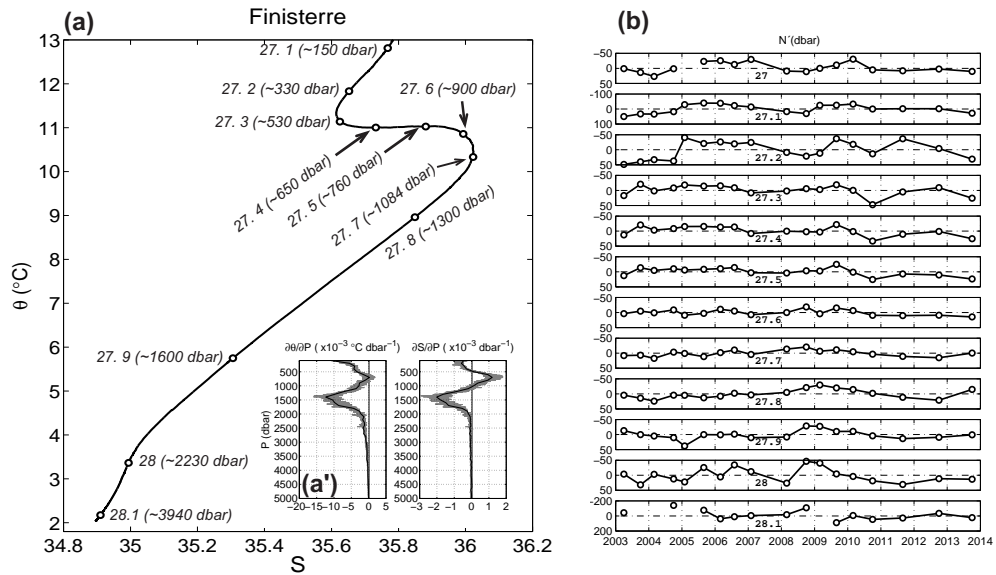


Figure 4.8: **(a)** Location of isoneutrals used for the study (lying at mean pressure in brackets) in the Finisterre along-section/ whole time period θS average. **(a')** Potential temperature and salinity vertical gradients. Black lines represent gradient of properties averaged over time; grey lines show gradients calculated for all profiles sampled during all cruises. **(b)** Interannual variations of isoneutrals pressure in the 2003-2013 period.

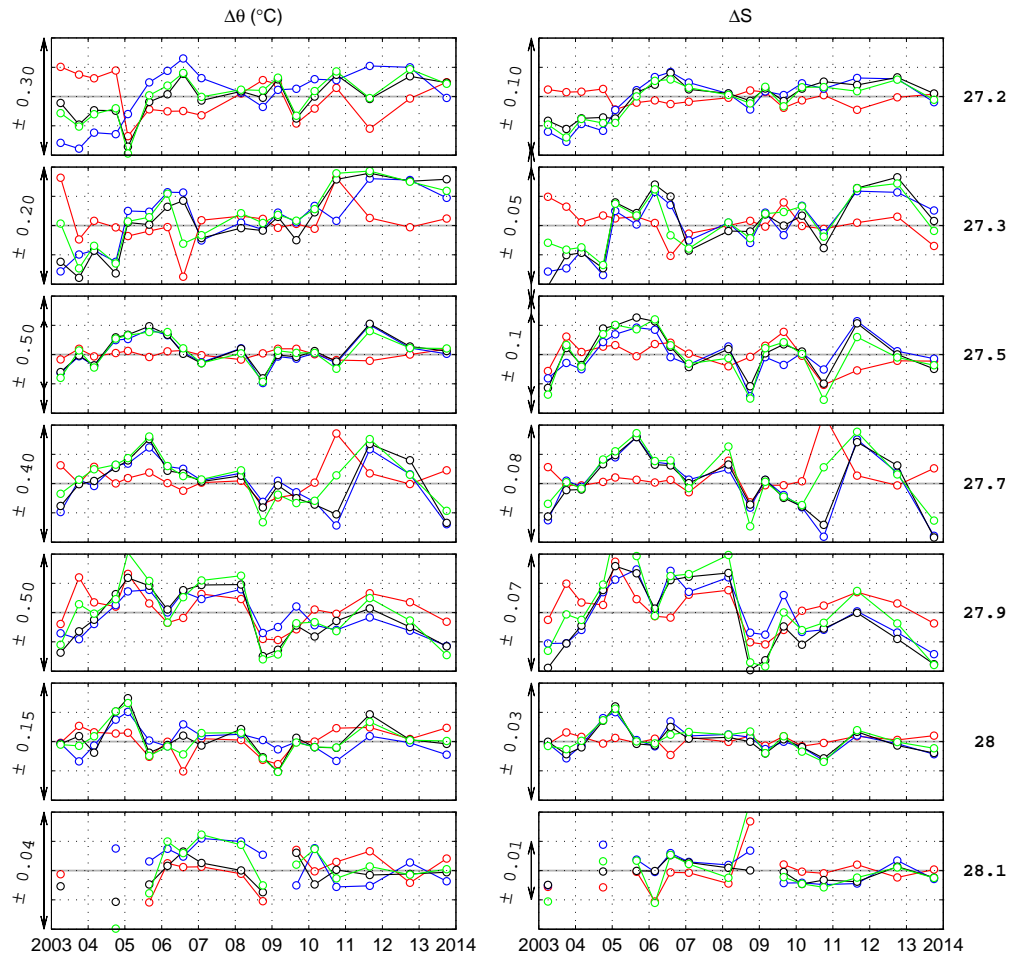


Figure 4.9: Time-series of potential temperature (left) and salinity (right) decomposition of isobaric changes (black, data from Sect. 4.2.1) on isopycnal change (blue) and heave (red) at selected isoneutrals: thermocline waters (ENACW, 27.2), the Salinity Minimum (27.3), Upper MW (27.5), MW core (27.7), Deep Mediterranean - LSW interface (27.9) and deep waters (ENADW, 28; LDW, 28.1). Differences between the black and green (sum of the blue and red) lines represent the inaccuracy of decomposition.

tral 27.9 by about 50 dbar (see Fig. 4.8b).

The fresher state around 2009-2011 in the deepest layer is consistent with isopycnal change, however the weak density gradients below the permanent thermocline mean that this conclusion has low certainty. Overall the variations in the deepest water of the Finisterre section are weak and show no clear trends.

During 2008, 2011 and 2013 there were coordinated responses across large parts of the water column so we present the vertical decomposition of changes on Fig. 4.10. In 2008 there was cooling and freshening down to about 2500 dbar (with maximum values at ~ 1500 dbar) as a combination of isopycnal change and heave. However between the Salinity Minimum and upper MW (~ 500 -900 dbar, 27.3-27.6) changes had a mainly isopycnal origin. In contrast Figs. 4.10b,e show warming and salinification across the permanent thermocline in 2011, with maximum values at ~ 1300 dbar (27.8) and with an almost exclusively isopycnal origin from 500 to 2000 dbar (27.3 to ~ 28). Note that all isoneutrals between 27 and 28.1 deepened in 2011 (Fig. 4.8b). Changes in 2013 showed cooling and freshening of the whole water column (Fig. 4.10c,f), with a maximum at 1300 dbar (27.8, deep MW) through a combination of isopycnal change and heave, the latter associated with the shoaling of isoneutral 27.8 (Fig. 4.8b). The water column down to the Salinity Minimum (~ 100 -500 dbar, 27-27.3) cooled by isopycnal change and warmed by heave due to the deepening of associated isoneutrals. In the next section we will discuss changes of water mass properties in terms of the NAO-induced forcing and main North Atlantic circulation patterns affecting the Finisterre section.

4.3 Currents and water mass transport

Potential temperature (a), salinity (b) and their respective vertical gradients (d,e), neutral density (c) and zonal density gradient (f) are shown in Figure 4.11 show for both cruises chosen for the study of transports of the different water masses in the region (Results from this section will be discussed later in Sect.5.3). Slightly cooler and fresher thermocline and intermediate waters (from below the mixed layer ~ 200 dbar down to 2000 dbar) is observed according to the event that affected the whole water column during 2008 (Prieto et.al 2014, under review).

Salinity vertical gradient (Figure 4.11e) is characterized by three marked extremes: two minimum at ~ 200 dbar and 1300 dbar, associated with the interface between upper and thermocline waters and the MW-LSW interface, and a maximum at ~ 700 dbar defining the ENACW-MW interface. Three crossings

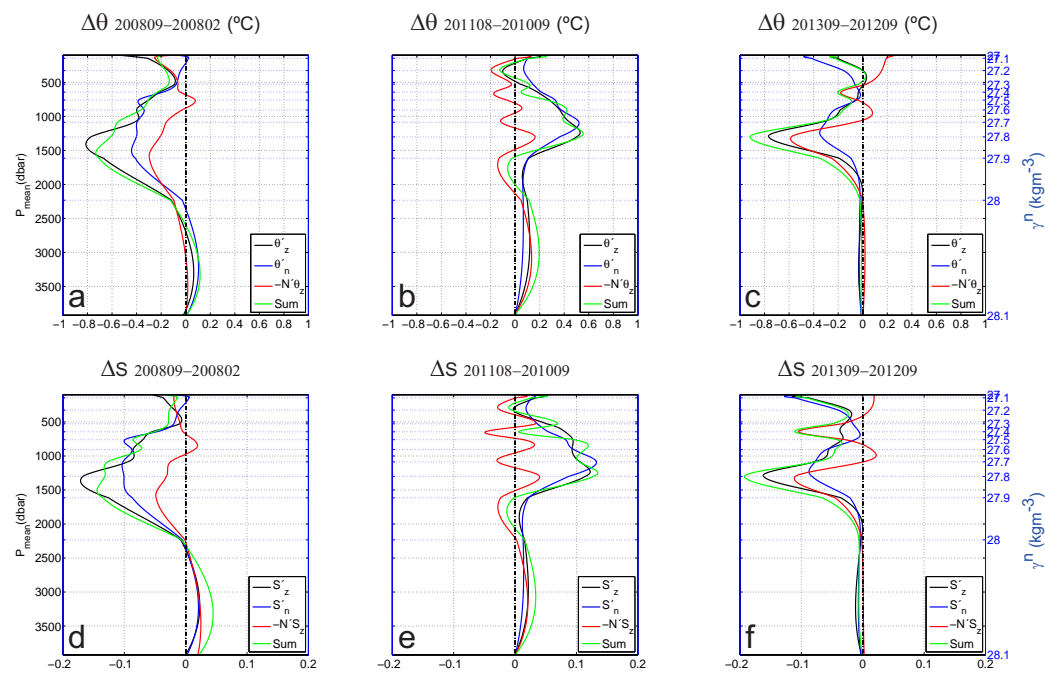


Figure 4.10: Decomposition of the temperature (a,b,c) and salinity (d,e,f) isobaric changes (black) on isopycnal changes (blue) plus changes by heave (red) following *Bindoff & McDougall (1994)*. The sum of both (green) accounts for the inaccuracy of the decomposition. Method was applied to changes observed in the water column between 27–28.1 kgm^{-3} , from February 2008 to September 2008 (left), from September 2010 to August 2011 (middle) and from September 2012 to September 2013 (right)

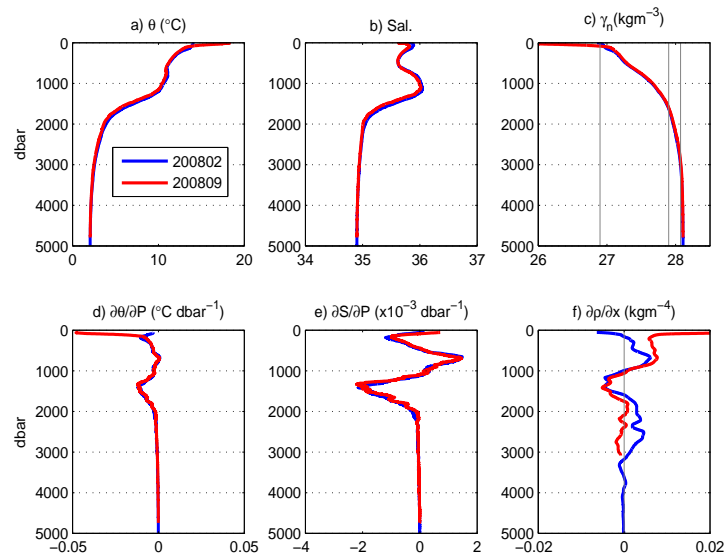


Figure 4.11: Potential temperature (a), salinity (b), neutral density (c), vertical gradient of temperature (d) and salinity (e), and horizontal gradient of density (f) during RadProf200802 (blue) and RadProf200809 (red). Vertical gray lines in c) show pressure of 26.9 (~ 100 dbar), 27.9 (~ 1650 dbar) and 28.072 (~ 3000 dbar) isoneutrals.

by zero define the mixed layer (~ 100 dbar), the Salinity Minimum (~ 500 dbar) and the salinity maximum associated to MW (~ 1100 dbar). Zonal density gradients (Fig. 4.11f) along the zonal section (computed as the overall zonal average) are characterized in general by large values above 3000 dbar, showing two relative maxima at 700 dbar and 1300 dbar, coinciding with those of salinity vertical gradient, a zero crossing at 1000 dbar during both cruises and small or near zero values below 3000 dbar, suggesting this depth as a good reference level for geostrophy calculations.

4.3.1 Directly measured meridional velocities vs geostrophic fields.

Figure 4.12 shows the current velocities across the Finisterre section as measured by LADCP during a) RadProf200802 and b) RadProf200809. Detided fields are shown by a', b', respectively. Red (blue) shades indicate northward (southward) flow. Great differences exist especially during winter above the Galicia Bank ($\sim 12^\circ W$) and near the slope ($\sim 10^\circ W$) associated to positive tidal velocities of ~ 0.05 m/s in winter (Fig.3.3). During summer, negative (southward) tidal velocities eastwards from $10^\circ W$ enhance the real northward flow measured by the LADCP (Fig.4.12b').

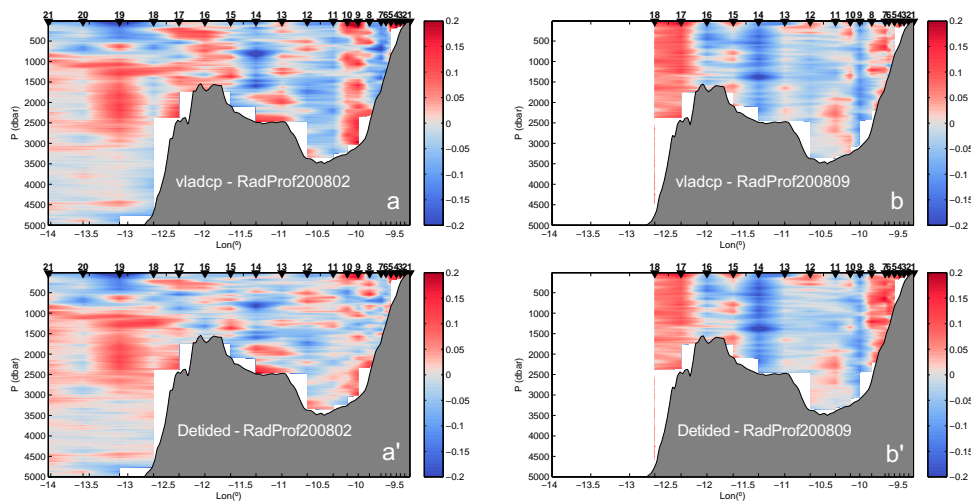


Figure 4.12: Current velocities from LADCP across the Finisterre section during a) RadProf200802 and b) RadProf200809. a'), b') denote velocity fields after tidal barotropic component removal.

Vertical sections of geostrophic velocity obtained from CTD profiles are shown in Figure 4.13 before (a,b) and after bottom-track correction (a',b'). As in Fig. 4.12, red (blue) shades indicate poleward (southward) flow. According to applied offsets, large differences are evident at the slope region during both surveys and near the zero reference velocity level as expected. Notable barotropic character of the currents is clearly seen along the section, showing alternating northward and southward flows. Poleward flow along the upper-slope is evident down to ~ 1300 dbar in winter, in both the initial and corrected fields, extending also towards the continental shelf in the corrected field. In summer, this northward slope current extends deeper (reaching MW depths ~ 1000 dbar) and offshore ($\sim 9.8^\circ\text{W}$) in the corrected field, extending also towards the continental shelf. Mesoscale structures are evident along the section, for example in summer it is observed northward geostrophic current around 10.5°W both in the initial and corrected fields reaching depths near 1500 dbar (Fig. 4.13b,b').

An anticyclonic recirculation around the Galicia Bank was observed both in winter and summer from corrected geostrophic fields, characterized by a southward flow in the upper 2000 dbar just over the Galicia Bank (from 11.5° to 12°W in winter, widening in summer from 11° to 12°W) and a northward flow in the west part of the seamount from 12° to 12.5°W .

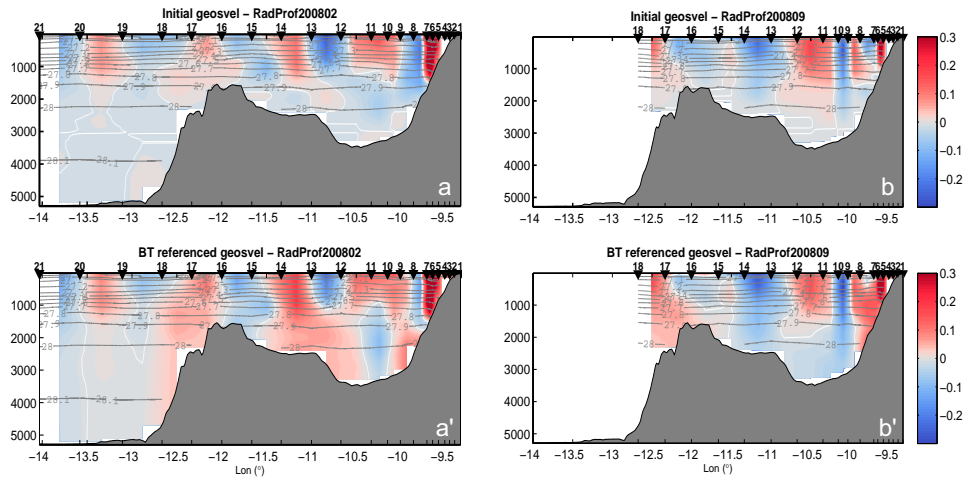


Figure 4.13: Initial geostrophic fields for a) RadProf200802 and b) RadProf200809. Red (blue) shades indicate northward (southward) flow overlaid by neutral density contours. Primes a', b' denote referenced geostrophic fields using LADCP bottom-track (BT).

4.3.2 Transport estimates

Mass transports calculated from direct LADCP velocity measurements along the Finisterre section during February and September 2008 are shown in Fig. 4.14, for the whole zonal section and separately for the slope and open ocean (see Fig. 2.3a,b) to study the contribution of both regions to the net transports.

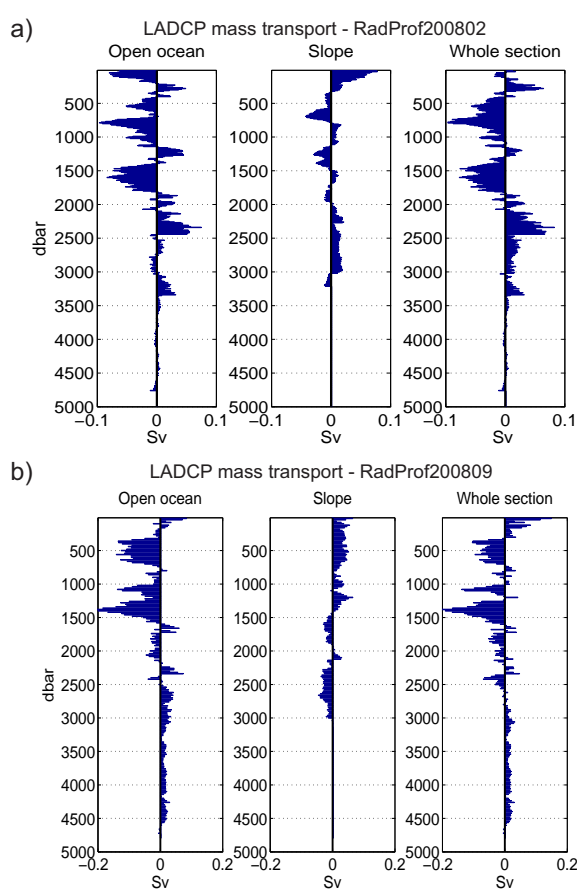


Figure 4.14: Mass transport (Sv) from LADCP measurements across the open ocean region, slope region and the whole Finisterre section during a) RadProf200802 and b) RadProf200809.

Interesting features deduced from Fig. 4.14 are:

1. the slope northward mass transport down to 400 dbar in winter, extending down to 1500 dbar in summer,

2. the two maxima southward mass transport at the open ocean region ~ 800 and ~ 1600 dbar during winter, appear shallower in the water column ~ 600 and 1400 dbar during summer,
3. the seasonal change in mass transport in the upper 2000 dbar of the water column with 4.2 Sv flowing southward during winter, and 3.2 Sv flowing southward during summer; below 2000 dbar, 2.9 Sv appear flowing northward during winter and 1 Sv during summer. This causes a net southward flow of 1.3 Sv in winter and 2.2 Sv in summer across the Finisterre section, that is, a larger southward flow is observed during summer time, associated to a less amount of deeper than 2000 dbar waters flowing northwards across the section (\sim the third part than in winter).

Cumulative mass transports were calculated for pressure intervals 0 - 500 dbar, 500 - 1000 dbar, 1000 - 1500 dbar, 1500 - 2000 dbar, 2000 - 3000 dbar, 3000 dbar to the bottom and integrated from top to bottom to obtain the total mass transport across the section (see Table 4.5).

As stated above, mass transport across the Finisterre section from LADCP measurements showed a larger southward transport in summer (-2.2 Sv) than in winter (-1.3 Sv) with a larger southward mass transport at the open ocean (3 Sv) due to the southward transport of waters down to 2000 dbar and the northward transport of waters deeper than 2000 dbar during summer time. During winter, a more intense northward flow is observed across the slope region (1.3 Sv) in response to the northward transport of upper layers between 0 - 500 dbar (1 Sv) and deep waters between 2000 - 3000 dbar both in the slope (1.26 Sv). Southward transport (-0.8 Sv) at the slope region between 500 - 1500 dbar in winter (RadProf200802) and northward transport (1 Sv) in summer (RadProf200809) may indicate the detachment of the MW vein from the slope in winter.

Geostrophic mass, heat and salt transports at neutral density layers are shown in Table 4.6 for the open ocean, slope region and whole Finisterre section during RadProf200802 and RadProf200809. An interesting behaviour arise from these geostrophic transports: poleward flow is observed deeper in the slope region transporting heat and salt in the water column down to MW in summer, whereas southward transport reach deeper at the open ocean region down to MW levels, transporting heat and salt southwards.

A decoupling between upper layer - central waters and intermediate waters, is observed at the slope region in winter and at the open ocean region in summer:

LADCP Mass transport (Sv)						
P (dbar)	RadProf200802			RadProf200809		
	Open ocean	Slope	Whole section	Open ocean	Slope	Whole section
0-500	-0.7790	1.0130	0.2340	-0.8998	0.9527	0.0529
500-1000	-1.8846	-0.3942	-2.2788	-1.2411	0.6791	-0.5620
1000-1500	-0.4082	-0.4410	-0.8493	-2.6504	0.4826	-2.1678
1500-2000	-1.3199	-0.0121	-1.3320	-0.2060	-0.3603	-0.5663
2000-3000	1.1666	1.2617	2.4283	0.7259	-0.9332	-0.2072
3000-bottom	0.5963	-0.0983	0.4980	1.2032	0.0005	1.2038
Total	-2.6289	1.3291	-1.2998	-3.0682	0.8215	-2.2467

Table 4.5: Mass transports calculated from detided LADCP velocities

in the first case, poleward flow is evident down to UMW and southward below; in the second case, southward flow is evident down to ENACW and northward below. Contrary, a coupling appears at the open ocean in winter (with southward flow down to MW and northward below, and at the slope region in summer, with northward flow down to MW levels and southward flow below.

A comparison can be made between transports calculated from geostrophy and detided LADCP measurements (see Fig. 4.15). A northward flow of 1.3 Sv was observed from LADCP at the slope region during a typical winter (Fig. 4.15a), whereas 3 Sv were observed flowing southward at the open ocean region in summer (Fig. 4.15b). Geostrophy seems to subestimate southward mass transport at the open ocean region, resulting in net southward transport across the Finisterre section of 0.1 Sv in winter and a net northward mass transport of 1.1 Sv in summer, contrasting with the net southward transports of -1.3 Sv in winter and -2.2 Sv in summer obtained from LADCP. A striking feature is the poleward flow structure of about 0.2 m/s in the first 1000 dbar that appears well defined between st.12-11 both in the summer initial and corrected fields (see Fig. 4.13b,b'), being displaced ~ 36 km westward in summer relative to its winter location. This structure, however, was not observed in LADCP measurements, and it may be a source of noise when computing transports.

4.3.3 The 43°N, 11°W mooring line

Time series of the Finisterre currentmeter mooring at 43°N, 11°W provide continuous measurements of current velocities at the ENACW (350 dbar), MW (1000 dbar) and LSW (1800 dbar) levels. Meridional volume transport can be computed across a slice bounded by stations 12 and 14 ~ 75 Km away. Time series of meridional current meter velocity showed a mainly barotropic circulation at this location during September (Fig. 4.16b,d), when the three water masses flow southward. In winter, volume transport is dominated by that of LSW at least until mid-February (Fig. 4.16a,c), whereas ENACW transport appear to dominate the southward transport in late September. Averaged transports showed a net southward volume transport of 0.5 Sv during February (Fig. 4.16c) and 5.4 Sv during September (Fig. 4.16d), that is a larger southward transport in summer, according to LADCP.

Table 4.6: Mass, heat and salt geostrophic transports across the Finisterre section, open ocean region and slope region during a typical winter (RadProf200802) and summer (RadProf200809).

		RadProf200802			RadProf200809		
Water mass	$\gamma_t \text{ kg m}^{-3}$	Open ocean	Slope	Section	Open ocean	Slope	Section
Mass transport (Sv)							
Upper layer	surface – 27.162	-0.64	1.70	1.06	-0.15	0.00	-0.15
ENACW	27.162 – 27.38	-0.58	0.97	0.38	-0.36	0.54	0.18
UMW	27.38 – 27.62	-0.24	0.75	0.50	0.06	0.50	0.57
MW	27.62 – 27.82	-0.04	-0.12	-0.16	0.27	0.48	0.76
DMW	27.82 – 27.922	0.05	-0.85	-0.80	0.12	-0.15	-0.03
LSW	27.922 – 27.975	0.01	-0.62	-0.60	0.09	-0.21	-0.11
ENADW	27.975 – 28.0986	-0.05	-0.42	-0.47	0.09	-0.21	-0.12
Total		-1.5	1.4	-0.1	0.10	1.00	1.10
Heat transport (PW)							
Upper layer	surface – 27.162	-0.033	0.091	0.057	-0.005	0.000	-0.005
ENACW	27.162 – 27.38	-0.026	0.044	0.017	-0.017	0.024	0.007
UMW	27.38 – 27.62	-0.010	0.032	0.021	0.002	0.023	0.025
MW	27.62 – 27.82	-0.002	-0.003	-0.006	0.010	0.020	0.031
DMW	27.82 – 27.922	0.002	-0.021	-0.019	0.002	-0.003	-0.001
LSW	27.922 – 27.975	0.001	-0.010	-0.009	0.001	-0.003	-0.002
ENADW	27.975 – 28.0986	-0.001	-0.005	-0.006	0.001	-0.002	-0.001
Total		-0.069	0.128	0.059	-0.006	0.059	0.054
Salt transport ($\times 10^9$ Kg/s)							
Upper layer	surface – 27.162	-0.023	0.061	0.038	-0.004	0.000	-0.004
ENACW	27.162 – 27.38	-0.020	0.034	0.013	-0.013	0.019	0.006
UMW	27.38 – 27.62	-0.009	0.027	0.018	0.002	0.018	0.020
MW	27.62 – 27.82	-0.002	-0.004	-0.006	0.009	0.017	0.027
DMW	27.82 – 27.922	0.002	-0.030	-0.028	0.004	-0.005	-0.001
LSW	27.922 – 27.975	0.001	-0.022	-0.021	0.003	-0.007	-0.004
ENADW	27.975 – 28.0986	-0.002	-0.015	-0.016	0.002	-0.005	-0.003
Total		-0.053	0.051	-0.002	0.003	0.037	0.040

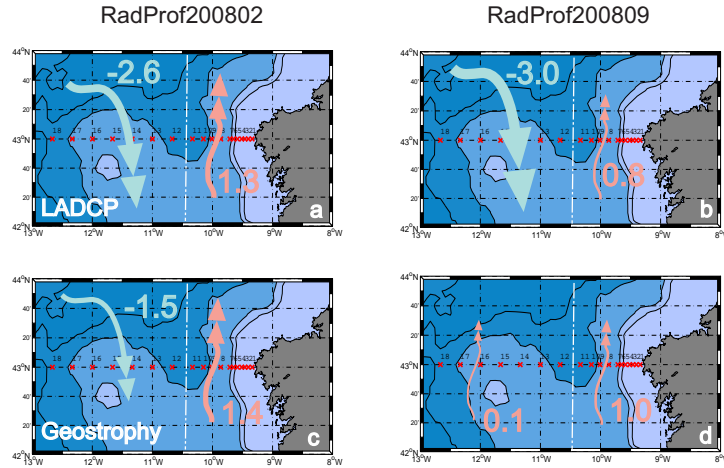


Figure 4.15: Total mass transport across the slope and open ocean regions from detided LADCP and geostrophy during RadProf200802 (a,c) and RadProf200809 (b,d). Red (blue) arrows indicate northward (southward) transport.

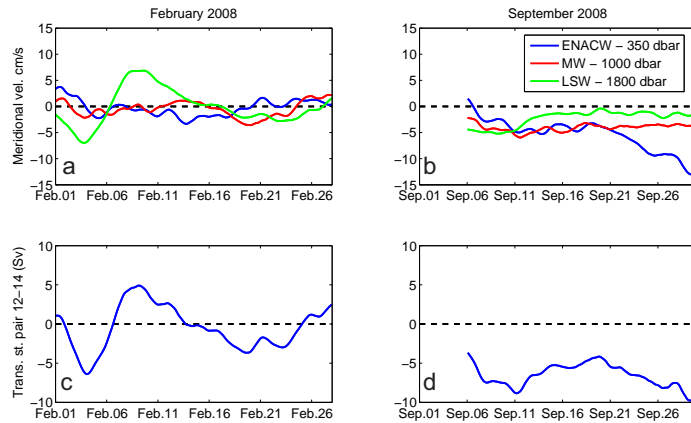


Figure 4.16: Time series of meridional velocity in cm/s from the Finisterre mooring at 43°N , 11°W during a) February 2008 and b) September 2008 with current-meters at 350 dbar (ENACW, blue), 1000 dbar (MW, red) and 1800 dbar (LSW, green). Depth-integrated meridional transports between stations 12 and 14 (~ 75 km spaced) during (c) February 2008 and (d) September 2008.

CHAPTER 5

Discussion

5.1 The seasonal signature at depth

5.1.1 Origin and processes involved

As introduced in Sect. 2.2, the 43° N section (Finisterre section, Fig. 2.1a,b) is located in the north-east limb of the North Atlantic subtropical gyre, an area of very weak circulation. Covering ~ 200 nm from the continental shelf to the open ocean and sampling the whole water column, its semiannual monitoring has provided the most detailed view of seasonality at all levels in the area to date.

Changes in the thermohaline properties at the sampling site may be caused by changes in the intrinsic properties of water masses and/or by changes in the circulation patterns (altering the water masses geographical distribution). It is however difficult to disentangle among them as the processes involved in the formation/transformation of water masses are intertwined with those altering the overall dynamics. As the water masses spread further from their sources towards the sampling site, it is more unlikely to discern a signature of shifts remotely imprinted on water properties, and even less any kind of seasonal swing. In the present Sect. we will discuss the observed seasonality signatures in our dataset in the context of the known oceanographic processes that cause seasonality in the North Atlantic eastern boundary.

5.1.1.1 Large scale circulation: the subtropical gyre

Our first-glance overall view indicates a more intense presence of northern origin waters in summertime all along the open ocean region (colder and less saline

waters at all depths). Such enhanced advection of subpolar waters during summer suggests large-scale fluctuations of the subtropical gyre (in position and/or strength). *Chidichimo et al.* (2010) and *Kanzow et al.* (2010) studied the seasonality of transports at the subtropics, downstream of our sampling site, by means of time series measured by the RAPID-MOCHA array (26.5° N). They concluded that the major contribution of seasonality to the meridional overturning circulation is precisely the seasonality at the eastern boundary, characterized by a maximum southward transport in spring (between April–May). Other works studying the seasonal variability of the Canary Current indicate in agreement and enhancement of southwards flow in early spring and weakening and even reversal (poleward flows) in late summer–autumn at some levels (*Machin et al.*, 2010; *Mason et al.*, 2011). These changes in the subtropics are connected with seasonal variations in the shape of the subtropical gyre, which has a larger zonal (east–west) and smaller meridional (north–south) extents in summer than in winter (*Stramma & Siedler*, 1988) (see their Fig. 3). Upstream of our sampling site, the North Atlantic Current also appears to suffer seasonality in its transport. Specifically, *Yaremchuk et al.* (2001) studied the circulation of the upper 1000 m in the 40–55° N and 20–40° W region finding among other results that the outflow leaving the eastern side of their study box is maximum around April. Therefore, known seasonality of the subtropical gyre, both upstream (NAC eastwards leakage) and downstream (Canary Current system), points to an enhanced flow in spring and southward displacement of the overall subtropical gyre system in summer. These findings are consistent with the enhanced presence of northern waters in summertime at our sampling site.

The main forcing mechanism capable of altering the circulation of an oceanic basin is the seasonality in the wind stress and wind stress curl, which causes variability in Ekman transports (*Stramma & Isemer*, 1988) and rearrangements of the density field which trigger the generation of Rossby waves that later propagate westwards into the basin (*Krauss & Wuebbler*, 1982). Actually, the subtropical gyre of the North Atlantic, particularly its eastern boundary, has been documented to be subject to large-scale seasonal variability in the wind stress curl (*Bakun & Nelson*, 1991, and Fig. 5.1).

Literature provides some clues in order to assess the role of propagating Rossby waves vs. Ekman transport in the observed seasonality at depth in our section, but a firm conclusion is difficult to achieve. *Hirschi et al.* (2007) studied the effect of seasonality in wind forcing on the meridional overturning circulation, stating that southwards of about 36° N the thermal wind contribution (variations of the density field) dominates the seasonal cycle while northwards of this latitude the

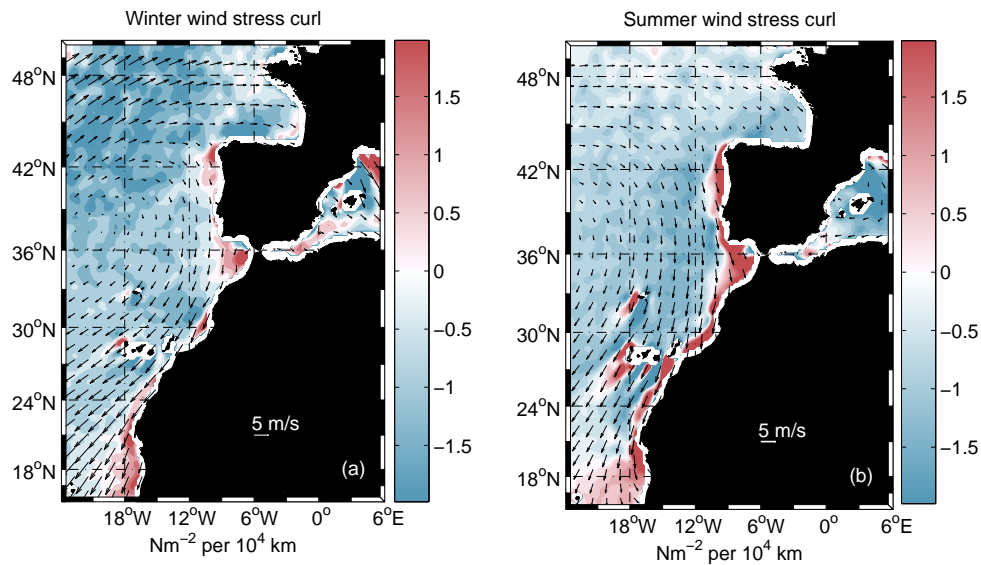


Figure 5.1: (a) Winter and (b) summer wind stress curl in the North Atlantic. Arrows show the wind velocity field in both seasons

major contribution is due to Ekman transport. In agreement, *Osychny & Cornillon (2004)* indicate from sea surface height (SSH) signatures that northwards of 41°N there is almost negligible energy at annual or shorter scales, being concentrated at periods longer than 1.5 yr. From these works, propagating Rossby waves at our 43°N section should not be expected to be the main contributor to the seasonality signatures at depth. On the other hand, *Bray (1982)* did attribute observational evidence of seasonal variability below the thermocline in the Bay of Biscay to the westward reflection of a seasonally wind-forced large-scale Rossby wave, and *Friocourt et al. (2008a)* linked seasonal flow reversals of modelled slope currents to annual Rossby waves. As a matter of comparison, Fig. 5.2 provides wind stress curl annual anomalies obtained from the SCOW (scatterometer climatology of ocean winds) climatology (*Risien & Chelton, 2008*) at (i) the RAPID (Rapid Climate Change) /MOCHA (Meridional Overturning Circulation and Heat Transport Array) region at 26.5°N (where *Kanzow et al. (2010)* show that a wind stress curl forced Rossby wave model successfully accounts for the observed seasonality) and (ii) off Finisterre at 43°N . Wind stress curl anomaly that would trigger Rossby waves is 2–3 times weaker in our region.

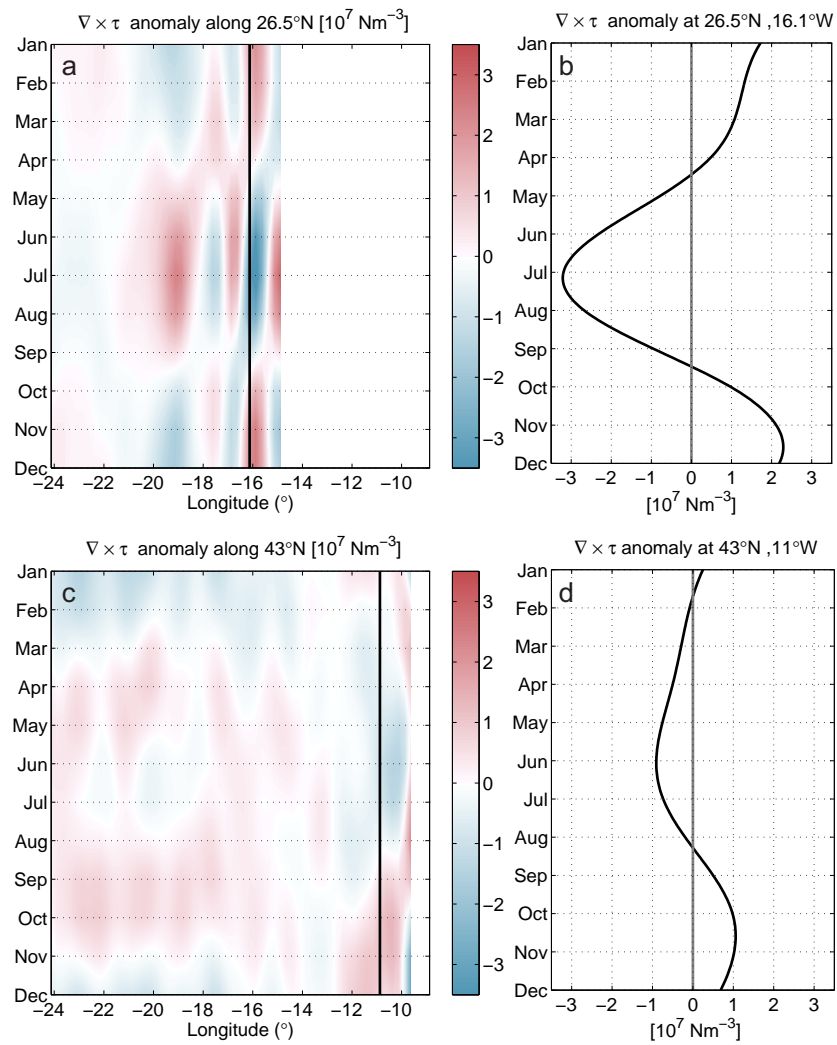


Figure 5.2: Wind stress curl anomaly from SCOW climatology at (a) 26.5° N, (c) 43° N, and extracted for (b) 26.5° N, 16.125° W and (d) 43° N, 10.875° W

5.1.1.2 Continental slope dynamics: upwelling, the Iberian Poleward Current and MW spreading

Shelf-slope dynamics at the western Iberian margin are subject to specific processes, some exhibiting strong seasonal character, that interact with the large-scale subtropical gyre seasonality. Northerly winds during the summer months cause widespread upwelling along the shelf (*Wooster et al., 1976*), while south-westerly winds in wintertime induce onshore Ekman transport. This upwelling/downwelling regime interacts with the slope dynamics characterized by the presence of the Iberian Poleward Current (IPC), a poleward flow driven by the meridional density gradient that carries warmer and saltier waters of southern origin (*Haynes & Barton, 1990; Frouin et al., 1990; Pingree & Le Cann, 1990*). The IPC is thought to be a permanent feature although it reaches a maximum development in wintertime after the strengthening of the meridional density gradients in late fall and winter (*Peliz et al., 2005*), aided by downwelling favourable winds, which adds up to a fifth of the total transport (*Frouin et al., 1990*). On the contrary, during the summer the IPC is not observed at the sea surface and it is thought to be displaced offshore and to persist below the surface (*Peliz et al., 2005; Ruiz-Villarreal et al., 2006*).

The effect of the upwelling/downwelling cycle on the central waters is reflected in our dataset as enhanced cooling/freshening on the slope in summertime, down to 600 dbar, caused by heave (Fig. 4.3b, d). The minor warming and salt increase at isopycnal levels on the slope in summertime is a striking signature (i.e. the 27.15 isoneutral is saltier and warmer in summertime but it lies much deeper). A rather speculative explanation for the presence of such central waters of southern origin on the slope in summertime may be linked to the presence of the deeper and offshore-displaced IPC pointed out by *Peliz et al. (2005)*.

Below central waters, the seasonality at the level of MW appears a robust feature with contrasting character on the slope (more presence in summertime) and at the open ocean (more presence in wintertime). The MW spreading from the Gulf of Cádiz formation area has been widely studied (e.g. *Iorga & Lozier, 1999a,b; Bower et al., 2002*), but information about its seasonality is still scarce and relies on reduced datasets. *Ambar et al. (1999)* reported insights of seasonality of the MW spreading around the Portimao Canyon (off the southern coast of Portugal, 37° N) and off the north-west coast of Portugal at about 41° N, respectively from XBT lines and year-long current meter moorings between July 1993 and May 1994. Their main outcomes were a broader signature of MW in wintertime (i.e. more irregular and extending further offshore) and a cycle in tem-

perature with notable offset: warming phases respectively at Portimao and 41° N from September–January and June–November (in agreement with our findings). *Varela et al. (2005)* also found a more intense MW signature in spring–summer on the slope at 42° N from a year-long weekly hydrostation between May 2000 and April 2001.

The IPC and the upwelling/downwelling cycle interacts with the dynamics of the Mediterranean water spreading at depth, and it has been proposed that poleward flows connect with the levels of MW (e.g. *Torres & Barton, 2006*). *Lafuente et al. (2008)* provided indications of the coupling of MW and the upper slope dynamics while giving further insights of seasonality of the upper branch of the MW off north-western Iberia. From a 4-month record in the upwelling season, of two mooring lines located on the slope slightly northwards of 43° N, they concluded that during the settlement of a wind-induced upwelling the MW behaviour was consistent with a combination of a shoreward and upwards displacement of its core. Such result is in agreement with our observations of salt increase by heave down to 800 dbar on the slope. Moreover, we see that the along-slope vein gets strongly reinforced in summer down to its lower core with the presence of purer Mediterranean water (i.e. warmer and saltier on isopycnals) while the density structure remains. From a historical dataset of current meters, *Huthnance et al. (2002)* inferred flow reversals on the slope at ENACW and MW levels around February–March, an observation that also matches with the loss of presence of MW in winter on the slope.

Daniault et al. (1994) and *Mazé et al. (1997)* provided a larger-scale view of MW spreading along the Iberian margin and outer ocean, finding a bifurcation of the coastal trapped MW tongue south of the Galicia Bank at around 42° N. They proposed an anticyclonic circulation of MW around the Bank to later re-entry in the Iberia Basin, and suggested that the bifurcation may exhibit strong variability, though not necessarily seasonal. Their research was mostly supported by Bord-EST 3 cruise, taking place around May but “at a time when the near-surface winter regime was still dominant”. Our results indicate that this pattern indeed corresponds to the typical hydrographical structure during wintertime. The diminished presence of MW in summertime at the outer region (colder and fresher at isopycnal levels) also implies a change in density structure (shoaling of isopycnals around $\sigma^{\theta} \sim 27.8$ making waters corresponding to the lower core of MW thinner). The pattern is consistent with the combination of the large-scale summer strengthening of the subtropical gyre and the reinforcement of MW along-slope vein.

5.1.1.3 Water masses formation at sources

5.1.1.4 ENACW mode water

Circulation and thermohaline properties of the central water east of the mid-Atlantic Ridge between 39° N and 54° N were analysed by *Pollard et al. (1996)* who observed that most of the transport associated with the North Atlantic Current (NAC) west of 20° W flows northwards (20 Sv) but that 10 Sv recirculate anticyclonically towards the west of the Iberian Peninsula, within the region south of the North Atlantic Current. *Mazé et al. (1997)* determined that the eastward flow towards the Iberian Margin at 12.5° W was of the order of 2 Sv from 38° N to 43° N.

The majority of ENACW water is formed by deep winter convection in a region north of 45° N from where it is subducted and advected southwards, being the main ventilation region of mode waters in the North Atlantic. The annual renewal of these mode waters was studied in detail from hydrography data and air–sea flux budgets (e.g. *Paillet & Arhan, 1996a,b*), the region at the Finisterre section being a frontal zone marked by an abrupt meridional shoaling of winter mixed layers (between 40° N and 45° N). *Gaillard et al. (2005)* conducted specific studies west of the Finisterre section (POMME–Multidisciplinary Meso Scale Ocean Program–area) finding a relevant contribution of the mesoscale activity in the net subduction rate of mode waters.

Pérez et al. (2000) described that the thermohaline properties of regional ENACW exhibit strong decadal variability related to large-scale climatic patterns affecting air–sea fluxes variability at the areas of formation. On the contrary, at seasonal timescales *Pérez et al. (1995)* found “a gradual increase of temperature and salinity from spring to late autumn” both at the open ocean and on the shelf without any insights of seasonality of salinity at ENACW isopycnal level. In agreement with their results, we observe colder and fresher waters in summertime due to the uplifting of ENACW isopycnal levels. Such a cycle can be associated, besides the dynamic response to the upwelling cycle, with a greater thickness of the well-mixed new mode waters in spring, just after the end of entrainment phase and during the subduction period (*Marshall et al., 1993*).

5.1.1.5 MW

There is even evidence of seasonality of MW right at its origin. Recent observations of the Mediterranean outflow water in the Strait of Gibraltar (*Lafuente et al., 2007*) showed the existence of a seasonal cycle with warmer and lighter

water leaving the Mediterranean Sea in winter, and cooler and denser waters in spring and summer (exhibiting also a maximum in transport) with amplitudes $\Delta T \sim 0.05^\circ\text{C}$ and $\Delta\sigma \sim 0.015\text{ kg m}^{-3}$. *Fusco et al. (2008)* reported seasonality in the Gulf of Cádiz with a salinity maximum around May–June, apparently in agreement with the maximum outflow around spring. Following the work of *Daniault et al. (1994)*, the amount of pure Mediterranean outflow waters present in the MW vein around Finisterre is about a fifth ($\sim 20\%$, see their Fig. 14). Therefore seasonality of MW at source cannot be the cause of the actual observed changes but only of a minor fraction.

5.1.2 Results from circulation models

Some modelling exercises within the area have focused on seasonality either at the boundary or offshore, and from surface to mid-depths, providing a non-homogeneous set of outcomes. For example, a large-scale model of MW spreading used by *Filyushkin et al. (2008)* indicated pronounced seasonality in the transport through the Strait of Gibraltar but did not find a noticeable seasonal signature in salinity distributions of MW across the North Atlantic. A more regional model focused on the western Iberian margin provided by *Coelho et al. (2002)* succeeded in reproducing the general pattern and seasonal variations of upper circulation, including upwelling during the summer, a winter surface poleward current over the shelf and a permanent year-round undercurrent transporting MW along the Portuguese and Spanish slopes. The model was also able to account for notable offshore export of MW at $41\text{--}42^\circ\text{N}$, in agreement with observations, but it did not find a clear difference between winter and summer at MW levels.

Among modelling efforts that suggest notable seasonality at depth, stands the work of *Friocourt et al. (2007)*. This study, supported by a regional model including western Iberia and the Bay of Biscay, simulates a strong baroclinic slope current at 42°N with a complex structure having cores at different levels and seasonal flow reversals at all depths. Salinity distribution in the core of MW indicates the presence of two flows: a narrow jet trapped on the slope, and a wider tongue flowing intermittently north-westwards toward the Galicia Bank. The position of the offshore MW tongue varies throughout the year, remaining in the vicinity of the slope from July to October and starting to move offshore around November, while the slope jet weakens and even reverses (i.e. flows equatorward) from November to January. *Friocourt et al. (2008a)* elaborated further on the seasonality of the slope current showing that reversals of the deeper slope currents are at least partly forced by seasonal changes in the flow upstream of the slope-current

system. *Slater* (2003) provided, from the large-scale OCCAM (Ocean Circulation and Climate Advanced Modelling) model, a pretty similar outcome showing boundary flow at MW depths directed northward year-round but displacing offshore in winter when a southward flow on the slope develops below 300 m. Both descriptions agree with our results, although we can not infer from our dataset whether the along slope MW vein comes to reverse in winter or simply weakens.

5.1.3 Quantitative comparison with other regions of the north-east Atlantic Ocean

As pointed out in the preceding discussion sections, there are many different dynamical processes that may imprint seasonality on to the deep ocean. However, many of the reviewed works only provide qualitative insights of seasonality or quantitative fluctuations of properties in a limited time frame, and rarely estimates of seasonal amplitude. All the available data about seasonal estimates collected in the reviewed literature have been joined with the seasonal isobaric changes found in our section, both in the slope and outer region (black lines in Fig. 4.3) in order to make a quantitative comparison between them (see Fig. 5.3).

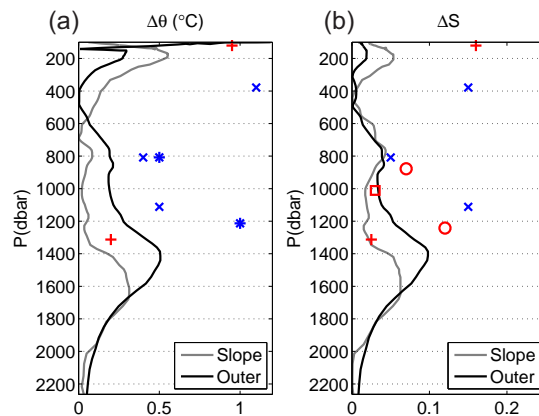


Figure 5.3: **(a)** Potential temperature and **(b)** salinity isobaric changes in the slope (grey line) and outer ocean (black line) superimposed by seasonal anomalies found by *Chidichimo et al.* (2010) (+), *Machin et al.* (2010) (o), *Ambar et al.* (1999) (*), *Varela et al.* (2005) (x) and *Bray* (1982) (□).

Chidichimo et al. (2010) estimated the seasonal amplitude in potential temperature and salinity from the EBH(eastern boundary)–RAPID mooring array

(4 yr continuous record) at the subtropical North Atlantic eastern boundary along 26.5° N, finding anomalies of 0.95°C , 0.16 at 120 m and 0.2°C , 0.025 at 1300 dbar. These values are in accord with our temperature seasonal anomalies at the outer upper ocean (see Fig. 5.3a) and temperature and salinity anomalies of intermediate layers in the shelf-slope region (0.14°C and 0.03 along 27.8). A little to the north, seasonality estimated by *Machin et al. (2010)* from the 9 yr continuous record of EBC4 mooring at $28^{\circ}46'$ N, $13^{\circ}28'$ W in the Lanzarote Passage (Canary Basin), was ~ 0.07 at 870 m and ~ 0.12 at 1230 m in salinity from July–October and from November–December, when maximum salinity was found in both levels. Off the northern coast of Portugal, from a specific mooring at 41° N, $9^{\circ}44'$ W, *Ambar et al. (1999)* also observed a shift from summer to winter of 0.5°C at 800 m and 1°C at 1000 m, between July 1993 and May 1994. In the Galician shelf/slope, a shift was also observed from summer to winter in a year-long series, from May 2000 to April 2001, of weekly hydrographical profiles at fixed location 42.13° N, 9.5° W by *Varela et al. (2005)*: they found seasonal shifts in temperature and salinity of about 1.1°C and 0.15 at ENACW_{sp} levels (300 – 450 m), 0.4°C and 0.05 at the upper MW (~ 800 m) and 0.5°C and 0.15 in the lower MW (~ 1100 m). In these cases, seasonal estimates are, in general, higher than those encountered in our section. *Bray (1982)* also estimated a seasonal amplitude of 0.03 in salinity at ~ 1000 m, from a series of 11 cruises over 3 yr in the Bay of Biscay area (2 – 20° W, 42 – 52° N), in concordance with our results (~ 0.02 – 0.03 along 27.6 – 27.7 between 900 – 1100 dbar). Note that red points in Fig. 5.3 represent observations spanning several years (from 3 to 9 yr), while those in blue are based on 1 yr sampling. We would like to highlight here the importance of establishing continuous and long time series of ocean properties in order to get a statistically significant seasonal cycle.

5.1.4 Capability of Argo floats to resolve seasonality in Finis-terre

It is interesting to determine whether the spatio-temporal coverage of the Argo fleet is enough to resolve the seasonal signature as our in situ dataset does. We used the data viewer interface provided by the Argo website¹ to retrieve the winter (January/February/March) and summer (June/July/August) salinity maps averaged for the Mediterranean water influence levels (1000 – 1250 dbar) in the region 30 – 50° N, 0 – 20° W (Fig 5.4).

¹<http://www.argo.ucsd.edu/index.html>

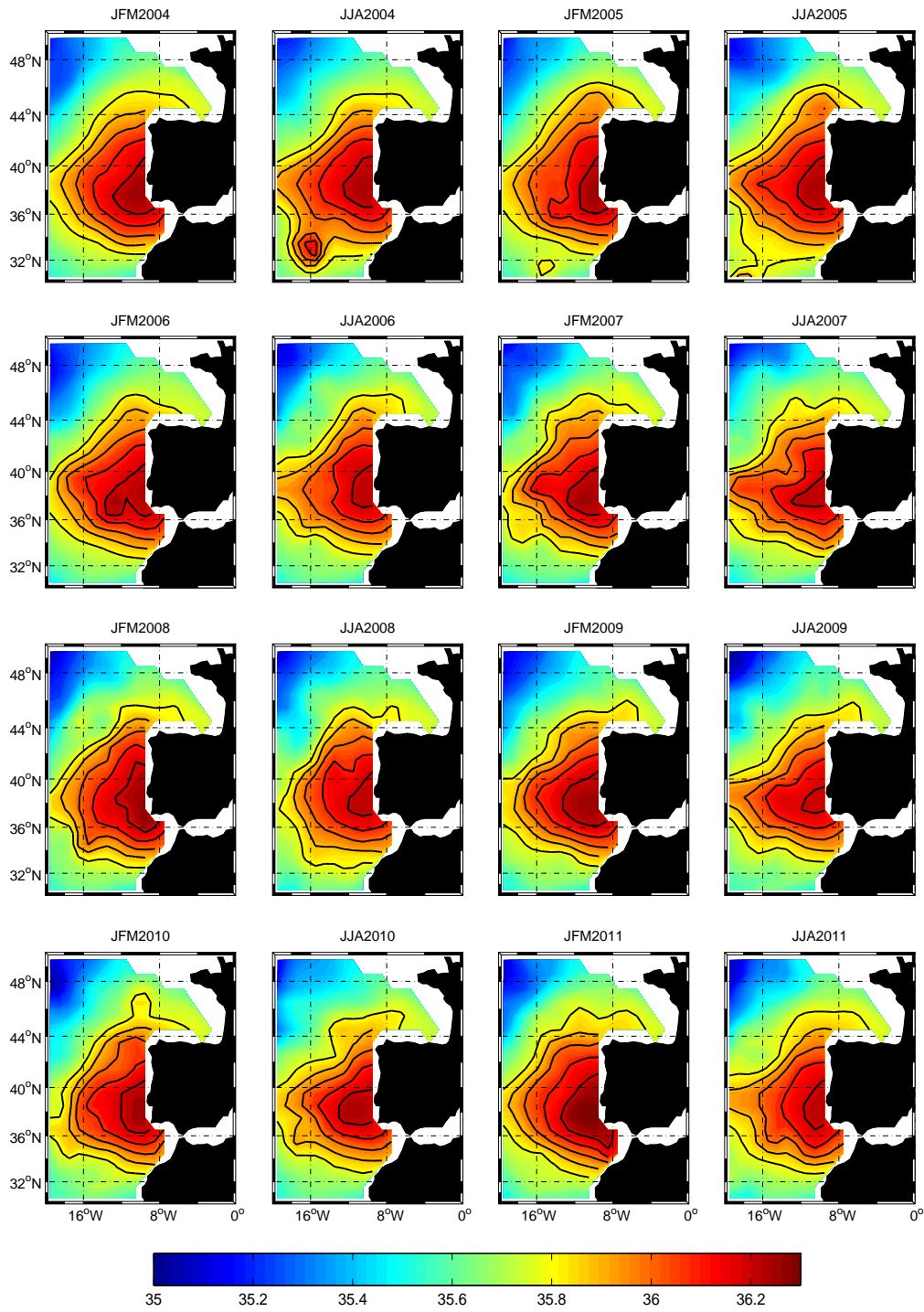


Figure 5.4: Winter (JFM) and summer (JJA) salinity fields at 1000–1250 dbar level from Argo floats for the period 2004–2011. Black contours are plotted from 35.8 to 36.2 every 0.1.

It appears difficult to evidence a recurrent seasonal pattern from such horizontal sections. The inspection of the available profiles (figures not shown) that support each of these interpolated field evidences that there may be large areas without profiles, specially the slope region, which is strongly undersampled (currents are stronger at these areas so floats are quickly advected). Nevertheless, we have computed the summer minus winter anomaly map (Fig. 5.5). The overall general view indicates less MW content in the open ocean of western Iberia in summertime (out to 16° W), which is consistent with our results from the section, and a higher MW content patch at the northwestern corner, which could be interpreted in a very speculative manner as a signature of enhanced northwards MW slope flow in summertime. However, the patchy character of the fields raises suspicions on its representativeness and the increase of MW in summertime west of 16° W seems quite odd.

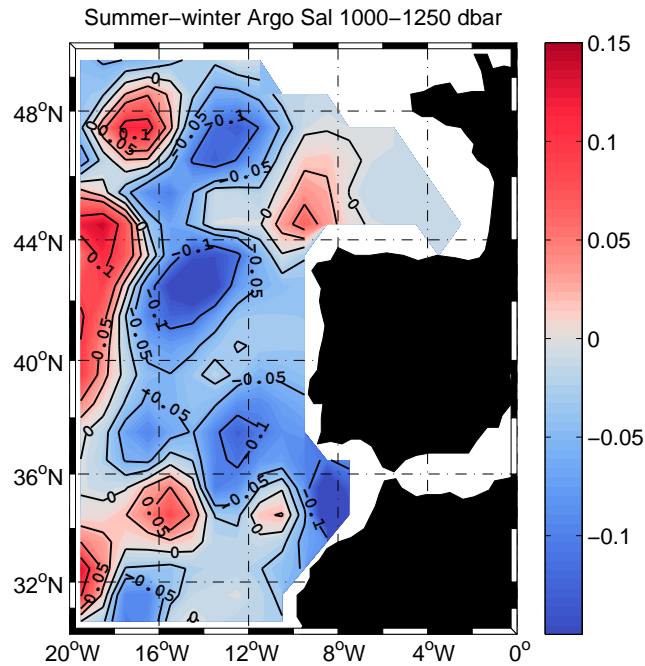


Figure 5.5: Summer (JJA)–winter (JFM) average salinity field at 1000–1250 dbar level from Argo floats for the period 2004–2011

Fig. 5.6 provides the time series from interpolated fields at some locations in

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the eastern boundary along 42.5° N. Though there are periods when there appears to be a seasonally oscillating behavior, the time series yields much less clear signs of seasonality than the time series derived from the hydrographical section do. In summary, we do not feel that we can draw firm conclusions about seasonality of the western Iberian margin hydrography from the Argo buoys, especially at the slope. Their coverage is not enough to capture the seasonal changes in the whole water column; the continuous repetition of the hydrographical section is becoming worthwhile for the study of the regional deep oceanography in the western Iberia at seasonal timescales.

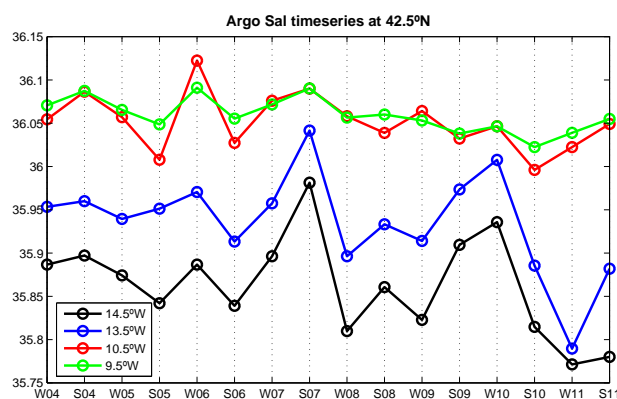


Figure 5.6: Time series of the salinity field along the 42.5° N section from the interpolated field at 1000–1250 dbar level from Argo floats for the period 2004–2011, at different locations in the slope (9.5° W and 10.5° W) and the outer ocean region west of the Galicia Bank (13.5° W and 14.5° W)

5.2 North Atlantic forcing driving interannual changes

The monitoring programme along the Finisterre section has provided a detailed view of interannual variability during the 2003–2013 period. The variability is the result of a complex combination of changes in large scale advective patterns, the intrinsic properties of water masses at their formation areas and local processes. The coherent response during specific years across large parts of the water column suggests an influence of the large-scale circulation pattern. In this section we discuss the causes of the observed coherent variability in relationship with the present understanding of North Atlantic circulation.

5.2.1 Climatic indexes and large-scale circulation of the North Atlantic

The North Atlantic Oscillation (NAO) is a leading pattern of weather and climate variability over the Northern Hemisphere strongly affecting the ocean through changes in heat content, gyre dynamics and deep water formation ([Hurrell & Deser, 2010](#)). During a negative phase of the NAO, the Labrador Current is enhanced bringing more fresh water into the subtropics, and the northward penetration of Mediterranean Water in the eastern North Atlantic also increases, there is southward displacement of the Subpolar Front in the western North Atlantic basin and the advection of more warm/salty subtropical water towards the northeastern North Atlantic ([Eden & Willebrand, 2001](#); [Núñez Riboni et al., 2012](#)). A high positive NAO phase is associated with the strengthening and eastward expansion of the Subpolar Gyre (SPG) bringing cooler, fresher conditions to the eastern region ([Sarafanov, 2009](#)). Figure 5.7 shows an scheme of North Atlantic circulation changes during positive and negative NAO phases.

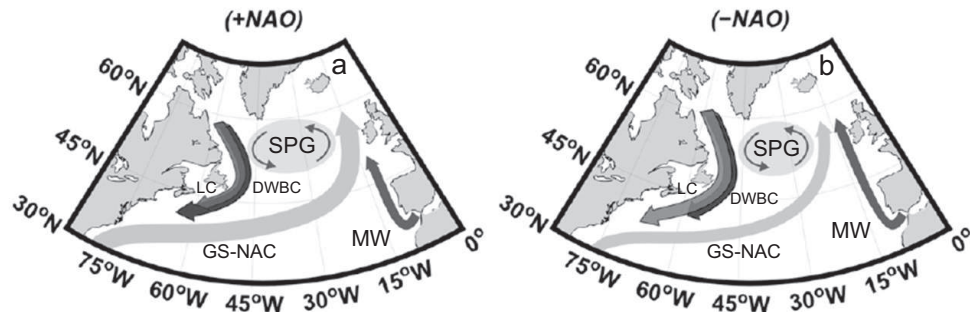


Figure 5.7: Schematic changes of the North Atlantic basin circulation during (a) positive and (b) negative NAO phases modified from [Chaudhuri et al. \(2011\)](#). LC: Labrador Current (upper 1000 m); DWBC: Deep Western Boundary Current (> 1000 m); GS-NAC: Gulf Stream-North Atlantic Current; SPG: Subpolar Gyre; MW: Mediterranean Overflow Water (500-1500 m). Thick (thin) lines mean enhanced (diminished) associated transport.

Strong and rapid shifts in the winter NAO index may cause similar effects in the North Atlantic Ocean circulation as a long period of sustained NAO index ([Chaudhuri et al., 2011](#)). For example, the circulation of the North Atlantic subpolar gyre appears to have weakened in response to a large shift in the NAO from

a period of very high to a very low index in winter 1996 ([Häkkinen & Rhines, 2004](#)).

The data presented here indicate that the temperature and salinity in the North Atlantic eastern boundary is responsive to extreme states of the NAO. In 2010, a low winter NAO index was followed by a large increase of temperature and salinity of almost the whole water column in 2011. Years with a high positive winter NAO index, such as 2007-2008 (high positive) and 2012 (very high positive) were followed by water mass property changes of the opposite sense; cooling/freshening across much of the water column. Note that changes associated with high positive NAO events were due to heave and isopycnal change (sinking of isopycnals) while changes after the high negative NAO only involve isopycnal change (Fig. 4.10). We do not know at this stage whether this may be a typical behaviour of large negative/positive NAO years.

It has been shown that the strength and extent of the subpolar gyre can have a profound influence on the water mass properties of the subpolar gyre ([Hátún et al., 2005](#); [Häkkinen & Rhines, 2009](#)). An index of the strength of the gyre circulation has been developed and shown to be only indirectly related to the NAO index ([Häkkinen et al., 2011a,b, 2013](#)). The subpolar gyre index is reproduced with the NAO winter index in Fig. 5.8 for the period 1993-2012. There is no clear relationship between the SPG index and variations in water mass properties at interannual scales in our mid-latitude section.

In summary, the time series presented here reflect a local oceanic response induced by the NAO. The influence is characterized by the presence of more cool/fresh water from the northwestern North Atlantic one year after a high positive NAO index year, and more intrusion of subtropical waters from the south one year after a strong negative NAO index year. The results provide more evidence of the role of the NAO in driving rapid changes in the deep ocean as well as in surface waters.

5.2.2 Changes in the main water masses

Besides causing changes in advective patterns (and not unconnected with these), large scale atmospheric variability conditions the formation of different water masses from modal to deep waters at their source areas. Next we discuss the variability observed in each water mass.

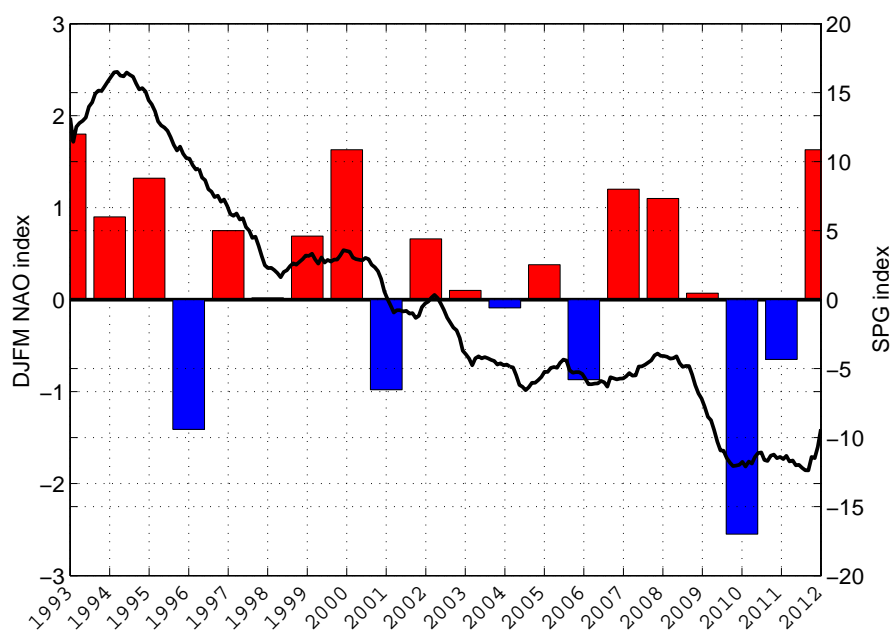


Figure 5.8: Winter(December-March) Hurrell NAO index with bars (red:positive NAO years; blue: negative NAO years), obtained as the first principal component of the winter SLP from <https://climatedataguide.ucar.edu/climate-data/hurrell-north-atlantic-oscillation-nao-index-pc-based> , *Hurrell & NCAR Staff (2014)*. SPG index obtained as the first principal component of SSH (*Häkkinen et al., 2013*)

5.2.2.1 ENACW

During the winter of 2005, the ENACW layers of the Finisterre section cooled by about 0.5°C in the upper 200 dbar and about 0.3°C in the layer 200-400 dbar. A more saline mode water was the outcome of this process, seen as isopycnal warming in the neutral levels 27.2 and 27.3, and a shoaling of isoneutral 27.2. This same pattern of change has been documented in the Bay of Biscay where it was attributed to the heat loss and the low net precipitation-minus-evaporation of the cold, dry winter affecting a wide region including the Western Mediterranean Sea (*Somavilla et al., 2009*). The implication is that the winter 2005 atmospheric conditions profoundly affected a wide area of upper ocean from the western Bay of Biscay to the eastern margin. The results are consistent with previous work that shows that the main driver for ENACW salinity variations is the local precipitation minus evaporation (P-E) balance (directly affected by the NAO), and that properties acquired by the ENACW every winter remains at least one year after the water mass formation (*Pérez et al., 2000*). For the deeper layers of the ENACW (the Salinity Minimum) the influences on the water mass properties are more complex because these layers can mix with warm saline Mediterranean Water and the slightly lighter Subarctic Intermediate Water (SAIW) (*Pollard et al., 1996*). The results from the Finisterre section suggest that the deep ENACW was affected by the cold winter of 2005 (no change in layer thickness but significant isopycnal change), but it is possible that the conditions were also influenced by the incursion of Subarctic Intermediate Water carried by a stronger North Atlantic Current flowing northeastward at lower depths (*Häkkinen & Rhines, 2009*).

5.2.2.2 MW-LSW levels

The Finisterre section is located on the northern edge of the subtropical gyre across the theoretical pathway of MW (Fig. 2.1a). The results at MW core (800-1200 dbar) show density compensated changes consisting of warming/salinification to 2005, cooling/freshening to late 2010, and sharp changes in 2011 and 2013. We have suggested that these changes in properties, particularly after strong NAO years, may be an advective response. Note that interpreting isopycnal changes at MW levels as having advective origin may appear counter-intuitive but reflects the fact that we are dealing with a water mass characterized by large lateral gradients. Also note that right at the core of MW there can not be property changes by heave since the vertical gradients of temperature and salinity vanish.

These changes are consistent with the proposed response of the SPG to the

NAO discussed in the literature. Periods of increasing and positive NAO index induces an expanding SPG and cooling and freshening of northern subpolar intermediate waters, but warmer and more saline MW in the eastern subtropical gyre since the spreading of the MW towards northern latitudes is blocked (*Lozier & Stewart, 2008; Lozier & Sindlinger, 2009; Chaudhuri et al., 2011*). This was the case for several decades up to the mid-1990s. After the mid-1990s the intermediate waters of northern North Atlantic warmed and became more saline as the SPG contracted and more subtropical water entered the region (*Sarafanov et al., 2009*). Our results are consistent with this view, with the warming/salinification observed to 2005 being likely a response to the contracting SPG (and more spreading of MW), and the subsequent cooling and freshening possibly a response to new positive phases of the NAO (and blocking of MW).

The most significant change in the properties of the LSW was seen in autumn 2008 as a drop in temperature and salinity that lasted to 2011. A similar change was seen in the Rockall Trough (which lies to the north) in 2006-2009 and is thought to indicate the arrival of a particular vintage of LSW formed in winter 2000 (*Nolan et al., 2012*). It is possible, but not certain, that the 2008 event at the Finisterre section indicates the arrival of the LSW₂₀₀₀ there. It has long been known that the largest changes to LSW properties are those induced by changes in air-sea fluxes during winter convection at source, sometimes as a direct response to the NAO (*Pickart et al., 2003; Flatau et al., 2003*). The range of property variations are so large that the variability of LSW at source may leave an imprint on the water mass far from the origin, and a range of travel times from the source to the eastern basins has been estimated based on matching Labrador Sea conditions with local changes (*Read & Gould, 1992; Cunningham & Haine, 1995; Pickart et al., 2003; Yashayaev et al., 2007*). Our results suggest that such remote imprints in LSW properties may reach even further than previously assumed.

5.2.2.3 Deep waters

The variations in the deepest water of the Finisterre section are weak and show no clear trends (see Fig. 4.5). Though there is evidence in the literature of changes in deep waters of the North Atlantic, during the time span of our timeseries the deep ocean in the Atlantic Iberian basin has remained stable. Examples of observed changes include decadal density-compensated temperature anomalies formed in the subpolar gyre and then propagated equatorwards (*Mauritzen et al., 2012*); and rapid freshening of the deep North Atlantic over past decades (*Dickson et al., 2002; Curry et al., 2003; Solomon et al., 2007; Atkinson et al., 2012*). *Atkinson*

et al. (2012) found a decreasing southward transport in the 3000-4700 m layer (LNADW) along 25°N, coincident with a density compensated cooling and freshening of the Denmark Strait Overflow Water in the Deep Western Boundary Current at that latitude.

5.3 On the north-west Iberia deep ocean circulation

5.3.1 Direct velocities vs geostrophy: Similarities and discrepancies

As part of an eastern boundary current system, the Coastal Transition Zone off the Iberian Peninsula is characterized by complex dynamics similar to that of the California Current system in the North Pacific (*Lynn & Simpson, 1987; Haynes et al., 1993*). The study region is characterized by a southward flowing Portugal Current (*Arhan et al., 1994*) favoured by northerly winds typical of upwelling systems, and a marked, narrow ($\sim 25\text{-}40$ km) and weak (~ 20 cm s⁻¹) poleward undercurrent flowing at $\sim 300\text{-}500$ m along the continental slope: the Iberian Poleward Current (IPC), which brings warm and salty subtropical waters northwards along the continental slope of the western Iberian Peninsula during winter (*Frouin et al., 1990; Peliz et al., 2003*). Looking at Fig. 4.13 large differences exist between the initial and bottom-track referenced geostrophic fields at the continental shelf region (st. 1-4) where there is an inversion in the corrected field towards a poleward flux (both in February and September) according to the winter development of the IPC along the continental shelf-upper slope (~ 400 dbar) between September - March (*Haynes & Barton, 1990*), suggesting that the LADCP bottom-tracking reference method improves the geostrophic circulation bringing it closer to the real circulation, since such a poleward flow over the continental shelf was also observed from detided LADCP (Fig. 4.12a',b'). This poleward flow jet extends down to MW levels (~ 1000 dbar even deeper in February (st. 5-7) than in September (st. 5-6), according to the shoreward and upwards wind-induced upwelling displacement of the MW core (*Lafuente et al., 2008*). Circulation in the western Iberia margin, although mainly seasonal, is superimposed by short-term fluctuations hosting a wide spectrum of mesoscale structures such as eddies, meanders, jets, counter-currents and upwelling filaments affecting the water mass interchange between the shelf-slope and open ocean in time scales less than one month (*Haynes et al., 1993; Huthnance, 1995; Huthnance et al., 2002; Peliz et al., 2002; Huthnance et al., 2009*).

The presence of an intense eddy field in the ocean makes circulation measurements very noisy and calculation of geostrophic transports variability very sensitive to the begin and end points of measurements (*Wunsch, 2008*). Since mesoscale structures can change during the time spent on sampling an hydrographic section, a synoptic view of the ocean may be lost as mesoscale structures evolve between the begin and end dates (*Vargas-Yáñez et al., 2005*). According to this mesoscale variability, the Finisterre section, which takes 3-4 days to be completed, evidenced a system of alternating southward/northward currents, jets and recirculations from geostrophic velocity fields at least down to 1000 dbar both during summer and winter 2008, making difficult the computation of reliable transports across the section.

There are several studies dealing with comparisons of geostrophy fields and directly measured current velocities, *Hinrichsen & Lehmann (1995)* studied the mesoscale mass distribution and flow associated to the MW from CTD and ADCP measurements in the southern Iberia Basin off Portugal coast and the Gulf of Cádiz, showing a good agreement between geostrophic and direct measured velocities, except near the coast where strong ageostrophic velocities are expected; *Lherminier et al. (2007)* calculated absolute transports across the Greenland-Portugal Ovide section during 2002 and 1997 using an inverse model constrained by direct ADCP measurements obtaining that direct current measurements differ significantly from geostrophy when the circulation is barotropic, a result that seems to match with our observations at Finisterre. In the Agulhas Current the velocity and vorticity study carried by *Beal & Bryden (1999)* showed an unchanged density structure of the current, that is, the baroclinic part of circulation provided by geostrophy appeared not to change, while the barotropic part of circulation seemed to be the main source of the intensity flow variability. Besides, LADCP and mooring measurements revealed an equatorward Agulhas Undercurrent not observed before. These authors attributed differences between LADCP and geostrophic velocities to the geostrophic method, and they used the LADCP to improve reference levels and overall geostrophic transports estimates, including improvement of deep water transport estimates. *Quaresma & Pichon (2013)* suggested that internal tides, which cause geopotential anomaly variations up to tens of dbar, may be also a source of discrepancy between geostrophy and ADCP velocities. They worked on a barotropic tidal model along the western Iberia margin, showing M2 barotropic current velocities up to 10 cm/s near the slope and Galicia Bank region that cause associated sea surface amplitudes up to 1 m in the region (see their Fig. 8 and 10). Moreover, they found internal tide generation "hotspots" around Cape Finisterre (at the slope region near the edge of the sub-

marine canyon of Mugia) and the southeastern side of the Galicia Bank (see their Fig. 17).

We hypothesize here that strong differences between our direct velocity measurements and geostrophic fields are related to: 1) the systematic error associated to LADCP measurements, 2) the error associated to the geostrophic method, since it is an approximation not applicable at ocean boundaries, and 3) the high influence of tides in the region, whose effects may be amplified by a complex topography characterized by the presence of seamounts and canyons. The most important issue that can be concluded from this study is that the mesoscale structures are being advected across the Finisterre section by the mostly barotropic overall flow (undetected by geostrophic computations). This emphasizes the importance that the continuous and systematic sampling of currents have for the knowledge of real ocean circulation patterns variations at all scales.

5.3.2 Transport estimates. Consistency with previous knowledge.

The large-scale circulation in the North Atlantic eastern basin is dominated by the NAC (48°-53°N) and the AC (34°-35°N), defining an inter-gyre region characterized by a weak and southward circulation (*Pollard & Pu, 1985; van Aken, 2001*). An inverse model of general circulation in the eastern North Atlantic was applied by *Paillet & Mercier (1997)* showing the decomposition of the North Atlantic Current (NAC) on several branches extending down to 2500 dbar. They found most of the NAC recirculating northward but some branches flowing southward later connecting with the Azores Current in the south. One of these branches flows southward along the western Iberia margin carrying 2 Sv in the 0-800 m layer in a typical late winter situation (March). This result agrees with the southward mass transport found down to 1000 m from our LADCP measurements in February (see Table 4.5). *Penduff et al. (2001)* applied a regional primitive equation model forced by seasonal and monthly surface fluxes finding that the southward recirculating branches of NAC bring 11.2 Sv toward the subtropical gyre across the intergyre region distributed as follows: 4.5 Sv from 0-800 m, 2.4 Sv from 800-1500 m and 4.3 Sv from 1500-2500 m. These climatological based-model results seem to overestimate our LADCP mass transports, but are reasonable values accounting for the extension of the survey region.

Many studies have focused on reproducing the seasonal slope poleward flow characteristic of the western Iberia margin, succeeding in reproducing the summer

upwelling, the winter surface poleward current and offshore exports both at IPC levels and the permanent MW undercurrent along the Portuguese and Spanish slopes (Coelho *et al.*, 2002; Friocourt *et al.*, 2007, 2008b,a). (Santos *et al.*, 2002) found from a numerical model study of the European margin that the slope current, intensified in autumn, is confined to the upper slope and outer shelf extending down to 1500-2000 m, according to our results. In the other hand, (Mazé *et al.*, 1997) pointed to the contrast between a variable meridional flow at upper levels of the eastern boundary in opposition to the more persistent southward regime in the open ocean, something that may also be concluded from our study.

Systematic measurements of velocity currents and the development of regional models of high resolution would be essential to solve circulation patterns in this part of the North Atlantic eastern boundary. Future work may also include inverse modelling based on boxes enclosed by the Finisterre (43°N), Ortegal (8°W) and Santander (3°, 47'W) deep sections, following the study carried by Fraile-Nuez *et al.* (2008) who used data from the cruise VACLAN0805 of the same oceanographic series than those analyzed in this paper (performed during 19 August- 11 September 2005) to obtain mass transports in the Bay of Biscay.

5.3.3 The shelf-upper slope. The Iberian Poleward Current.

It seems from Fig.4.13 that the poleward flow along the shelf - upper slope is present both during winter and summer carrying 4.9 Sv and 3.7 Sv northwards between stations 1-7, respectively. According to Coelho *et al.* (2002) poleward transport in the Iberian shelf-slope region is present all year round, decreasing from winter to summer. Peliz *et al.* (2002) found from a survey carried off the northern coast of Portugal, a circulation system with zonal variability characterized by a coastal northward counter current coexisting with upwelling southward jets, the slope poleward flow and the large-scale southward flow. These authors suggest that this coastal counter-flow is the response to the offshore drift of the upwelling current influenced by the Western Iberian Buoyant Plume. Lherminier *et al.* (2007) found a northward transport of 2.1 ± 0.4 Sv associated to the Eastern Boundary Current of the Iberian shelf and slope during summer 2002, not too different to our results.

In order to know what were the conditions during February and September 2008, the upwelling index (coastward transport per km parallel to coastline) was plotted from 2008 monthly time series of wind stress curl provided by the Villano-Sisargas and Silleiro oceanic-meteorological buoys of *Puertos del Estado*. Time series data were retrieved from the Spanish Institute of Oceanography visor tool

of upwelling index ², see Fig. 5.9. The index showed negative values during February and positive values during September, as expected during typical winter downwelling and summer upwelling regimes, suggesting that upwelling conditions could effectively coexist with a poleward slope current during summer time.

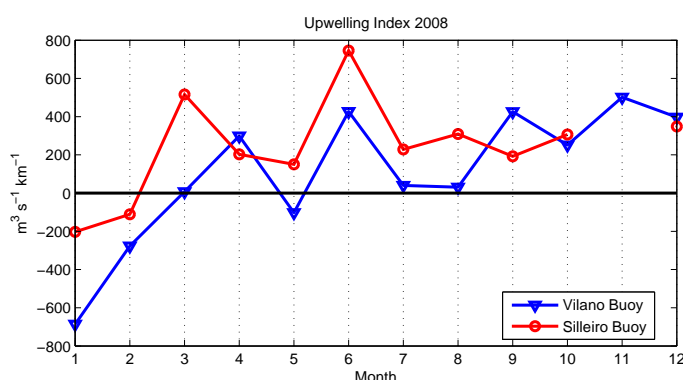


Figure 5.9: Monthly time series of upwelling index at Villano-Sisargas (Coruña, 43°29.4'N, 9°12.6' W) and Silleiro (Vigo, 42°7.3'N, 8°54.6' W) buoys during 2008.

During summer, this poleward flow current was observed both in the LADCP field (Fig. 4.12b') and in the LADCP referenced geostrophic velocity field (Fig. 4.13b') extending from the coast to 10°W, carrying ~ 7.5 Sv northwards, whereas 2.7 Sv appears flowing northwards west of 10°W, resulting on a total northward mass transport of 10.2 Sv across the Finisterre section during RadProf200809 (September). According to that, *Mazé et al. (1997)* found a northward geostrophic flow at all levels of 12.2 Sv across 43°N during May 1989, observing an intense higher than 1 Sv value at the easternmost part of the 43°N section that they proposed to be a remnant of the narrow surface winter slope current still present in May 1989.

The deeper-reaching slope poleward flow during summer and the deeper-reaching southward flow down to intermediate depths at the outer ocean in winter, are consistent with the attachment to the slope of the MW vein during summer and spreading over the Galicia Bank in winter (enhancement of the southward flow due to recirculation) suggested in Sect. 4.1.

²<http://www.indicedeafloramiento.ieo.es>

5.3.4 Recirculation around the Galicia Bank.

Mazé et al. (1997) also decomposed the flow associated to the MW into three longitudinal regions characterized by a northward transport of 7.6 Sv west of the Galicia Bank, part of which recirculates anticyclonically around the seamount, a southward transport of 3.8 Sv between the east part of the Galicia Bank and $10^{\circ}30'W$, and a northward flow of 2 Sv between this longitude and the continental slope. They also found that Central Water (ENACW) showed a zonal distribution similar to that of the underlying MW, suggesting that the southward flow east of the Galicia Bank is a recirculation of a part of the northward ENACW flow in the west. The coupling between ENACW and MW was also reported by these authors who confirmed the existence of a poleward slope flow at both water levels through $43^{\circ}N$ of 4.7 Sv, very similar to the 4.4 Sv obtained from our LADCP measurements. They also observed the warm and salty surface water displaced 200 km offshore with a corresponding high salinity anomaly at MW levels (900m) indicating that part of the MW core may be also displaced offshore.

A comparison can be made with those transports found at similar regions along the Finisterre section during summer: we have divided our section following the physical criterion based on the RadProf200809 LADCP velocity distribution, which showed poleward flows at the slope region and western Galicia Bank, and a southward flow between $\sim 10-12^{\circ}W$ (Fig. 4.12b'). Computed transports for these three regions are shown in Figure 5.10. Our results showed good agreement with those obtained by *Mazé et al. (1997)* in direction but not always in magnitude: we found a northward transport of 2.5 Sv at the slope, and a southward transport of 3.2 Sv east of the Galicia Bank (Fig. 5.10b), according to their northward flowing 2 Sv and southward flowing 3.8 Sv, respectively. However, the northward transport at the outer region west of the Galicia Bank is underestimated (1.8 Sv) in comparison with their 7.6 Sv, which in the other hand agree with the 7.8 Sv obtained from our bottom-track referenced geostrophic field (Fig. 5.10c). Contrary, mass transports computed for the referenced geostrophy both at the slope region (3.7 Sv northward) and the eastern Galicia Bank (12 Sv southward) would mean an improvement of geostrophic transports as being closer to those obtained from LADCP (4.4 Sv and 12.8 Sv respectively, see Fig. 5.10a).

Regarding to the seasonality of recirculation around the Galicia Bank, we cannot conclude that but only establish differences between the two surveys: in summer, a barotropic southward circulation in almost the whole water column at the east flank of Galicia Bank (between stations 13 and 16), both in the initial and corrected fields (and according to de-tided LADCP velocity field (Fig. 4.12b'))

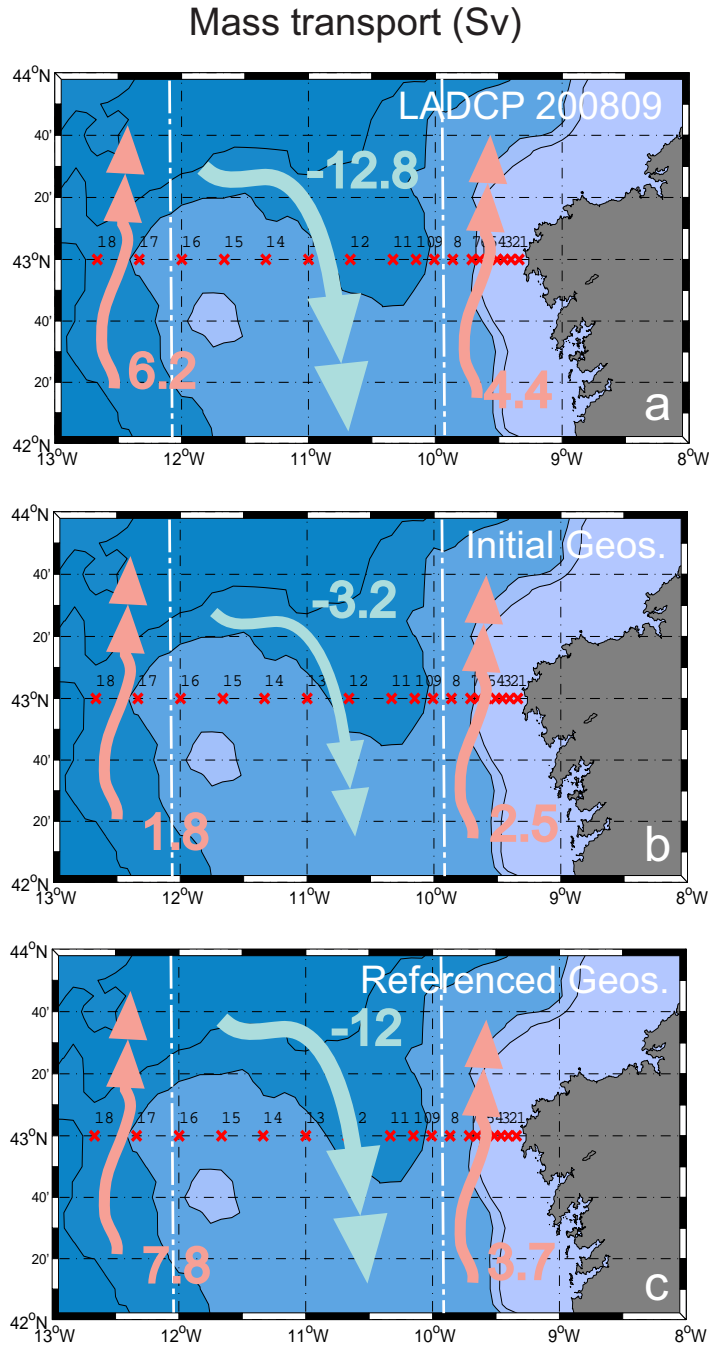


Figure 5.10: Mass transport (Sv) from (a) LADCP , (b) initial geostrophy and (c) BT referenced geostrophy during RadProf200809. Zonal section was divided in three regions: the slope region from the coast to $\sim 10^\circ\text{W}$ (st. 1-8), from $\sim 10^\circ\text{W}$ to $\sim 12^\circ\text{W}$ (st. 8-16) and westward from 12°W (st. 16-18). Northward (southward) transports are shown in red (blue).

was observed according to the larger presence of waters of northern origin during summer, as it was previously suggested in Sect. 4.1 in response to the seasonality of the North Atlantic subtropical gyre, suggesting the region between the Galicia Bank (eastward from $\sim 12^\circ\text{W}$) and the continental slope region (westward from $\sim 10^\circ\text{W}$) as a passage for northern origin waters during summer time. Computed mass transports show a southward transport of 8.0 Sv almost compensated by the 7.8 Sv flowing northward in the western part of the Galicia Bank (between stations 16 and 18). In winter, an enhancement of the northward flow at both sides of the Galicia Bank down to ~ 2500 dbar was observed in the referenced field between station pairs 18-17, 17-16 (1.5 Sv), 14-13 (4.1 Sv) and a southward flow above the seamount between station pairs 16-15 and 15-14 (2.5 Sv), suggesting a recirculation of the flux around both sides of the Galicia Bank (net northward transport of 3.1 Sv during winter), pointing to the permanent anticyclonic recirculation around the seamount. This northward transport associated to the Galicia Bank recirculation seem to agree again with estimations of *Mazé et al. (1997)*, who argued that *the supply of higher latitudes with MW by the western branch would not exceed 3.8 Sv*.

CHAPTER 6

Conclusions

Hydrographical properties of a zonal section ~ 200 nm off Cape Finisterre (43° N, $009\text{--}014^\circ$ W) have been sampled for the period 2003–2010 by means of two cruises per year (summer/winter). The sampling has revealed that there exists a noticeable seasonal signature in this part of the eastern boundary of the North Atlantic down to 2000 dbar. The magnitude of seasonality is roughly 20 % of the interannual variability observed for the overall period, reaching values up to 0.5°C , 0.04 in salinity just below the maximum extent of winter mixing (c.a. 200 dbar), and 0.4°C and 0.08 in salinity at around 1400 dbar, within the lower bounds of Mediterranean water.

The spatial structure of the seasonality shows a strong contrast between the near-slope area and the outer ocean. The most remarkable and novel finding is the behaviour of the Mediterranean water vein that gets attached and reinforced along the slope in summer and spreads offshore in wintertime. The observed pattern confirms, from a much larger dataset, some dynamical aspects regarding the Iberian margin seasonality and MW spreading that had been previously inferred from quite reduced datasets. Among them we can highlight the shoreward and upwards displacement of slope MW as a response to the seasonal upwelling and the appearance of a secondary branch of MW around the Galicia Bank, a feature that appears to be typical of the winter regime. The results have been discussed in terms of the seasonality of large-scale and continental slope processes and a brief review of modelling efforts regarding seasonality at the area has been made, showing that key features, reproduced by some of them as the winter flow reversal at MW levels on the slope, are consistent with our observations. The time series of in situ data provided here become a valuable tool for evaluating circulation models at regional or oceanic scales. The seasonality in the area must be taken into

account when analysing specific hydrographical observations or long-term series.

A monitoring program of the deep ocean in the northwestern Iberia and southern Bay of Biscay, running since 2003, has provided a detailed view of interannual changes in the last decade in the mid-latitude transitional intergyre region, complementing well the established monitoring programs in the subpolar and subtropical regions. The time series exhibited statistically significant autocorrelation, correlation across large portions of the water column, and geographical correlation mostly between Finisterre and Santander sections, suggesting that the sampling scheme is adequate to resolve background variability over mesoscale activity. The most comprehensive results are provided by the Finisterre section which is the deepest, longest and most often repeated. The Finisterre section has been shown to be representative of changes in the North Atlantic eastern boundary. The hydrographic time series are shown to have primarily isopycnal changes of termohaline properties. Statistically significant trends over the time series were found in the ENACW, dominated by a change to a saltier/warmer state after the extreme winter forcing of 2005. A significant long-term trend was also observed at the transition level between MW and LSW, dominated by the change to colder/fresher conditions towards the end of the series, from 2008 to 2010. Variability below the 2000 dbar is very small, oscillating around the limits of what is measurable.

The water mass properties of the section are strongly influenced by the state of the NAO, most obviously as a rapid and local response to the high positive NAO in winters 2007-2008 and 2012 (inducing cooling and freshening), and the extremely negative NAO index in winter 2010 (inducing warm and saline intermediate conditions). The results are consistent with present knowledge about North Atlantic circulation and its response to atmospheric circulation. The present work supports the importance of the maintenance of continuous monitoring programs of the deep ocean variability, through the periodic repetition of hydrographic sections, as a valuable tool for analyzing deep ocean changes on interannual to decadal scales, including unpredictable strong shifts essential to predict future local or global climate change. The results derived from this monitoring program are expected to contribute to future revisions and further developments in long-term Spanish climate monitoring strategy.

Geostrophic current velocities were computed from CTD measurements along the Finisterre deep section (> 5000 m, ~ 200 nm off Cape Finisterre), during winter and summer 2008, choosing a level of no motion at $\gamma^t = 28.072\text{kgm}^{-3}$ (~ 3000 m) or the deepest common level at shallower stations near the slope or the Galicia Bank seamount. Geostrophic fields showed an alternating system of northward and southward currents evidencing mesoscale structures not observable

by direct measurements from LADCP, which showed a more barotropic circulation according to a fixed currentmeter mooring at 43°N , 11°W . As a consequence, the geostrophic field seem to underestimate mass transport across the Finisterre section by one order of magnitude during winter, even reversing the flow during summer. LADCP measurements revealed a more intense southward mass transport in summer (2.2 Sv) than in winter 2008 (1.30 Sv) across the Finisterre section. Findings also include a poleward flow extending from the shelf and slope down to 1500 dbar during summer, in apparent opposition to the expected southward flow characteristic of a summer upwelling regime, and a permanent anticyclonic recirculation of the water column around the Galicia Bank. Results are consistent with the spreading of the MW vein over the Galicia Bank in winter (posterior recirculation) and the constraint of its core against the slope in summer. This study revealed the importance that systematic measurements of current velocities have on determining the real ocean circulation patterns in this region, both as a tool to obtain a direct measurement of currents, and as an useful method to improve regional circulation models in a region characterized by a wide range of dynamics and variability scales.

CHAPTER 7

Conclusiones

Las propiedades hidrográficas de una sección zonal ~ 200 nm al oeste de Cabo Finisterre (43° N, $009\text{--}014^\circ$ W) se han muestreado durante el periodo 2003-2010 a través de la realización de campañas semianuales de invierno/verano. El muestreo ha revelado la existencia de una señal estacional notable en esta parte del contorno oriental del Atlántico Norte hasta los 2000 dbar. La magnitud de la estacionalidad es aproximadamente un 20 % de la variabilidad interanual observada durante todo el periodo, alcanzando valores de 0.5°C y 0.04 en salinidad justo por debajo de la profundidad máxima de la capa de mezcla invernal (200 dbar) y 0.4°C y 0.08 en salinidad en torno a los 1400 dbar, en el límite inferior del agua mediterránea.

La estacionalidad muestra una estructura zonal con un fuerte contraste entre la región más cercana al talud continental y el océano abierto. El hallazgo más novedoso es el comportamiento de la vena de agua mediterránea, la cual aparece ceñida y reforzada a lo largo del talud en verano, propagándose hacia fuera de la costa en invierno. Este patrón observado confirma, a partir de un conjunto de datos mucho mayor, algunos aspectos dinámicos relacionados con la estacionalidad en el margen Ibérico y la propagación del agua mediterránea, inferidos previamente con bases de datos más reducidas. Entre ellos podemos destacar el desplazamiento hacia la costa y hacia arriba en la columna de agua de la vena de agua mediterránea en el talud en respuesta al afloramiento estacional, y la aparición de una rama secundaria alrededor del banco de Galicia, característica que parece ser típica del régimen de invierno. Los resultados se discutieron en términos de la estacionalidad de larga escala y los procesos de talud continental, y se ha realizado un repaso a los esfuerzos de modelización de la estacionalidad en el área de estudio, mostrando que las principales características reproducidas por algunos de ellos (como la inversión invernal a niveles del agua mediterránea) en el talud,

son consistentes con nuestras observaciones. Las series temporales de datos in-situ se convierten en una herramienta de valor para la evaluación de los modelos de circulación a escalas regionales y oceánicas. La estacionalidad en el área debe ser tomada en cuenta a la hora de analizar observaciones hidrográficas específicas o series temporales a largo plazo.

Este programa de monitorización del océano profundo en el noroeste de la Península Ibérica y sur del Golfo de Vizcaya iniciado en 2003, ha proporcionado también una visión detallada de los cambios interanuales durante la última década a latitudes medias en la región de transición entre los dos grandes giros, complementando los ya bien establecidos programas de monitorización de las regiones subpolares y subtropicales. Las series temporales exhiben autocorrelación, correlación a lo largo de grandes porciones de la columna de agua y correlación geográfica estadísticamente significativa principalmente entre la sección de Finisterre y Santander. Ello sugiere que el esquema de muestreo establecido resulta adecuado para resolver la variabilidad sobre la actividad de mesoscala.

Los resultados más completos los proporciona la sección de Finisterre, la más profunda, larga y repetida, siendo representativa de los cambios en el contorno oriental del Atlántico Norte. Las series temporales hidrográficas muestran principalmente cambios isopícnos de las propiedades termohalinas. Se encontraron tendencias estadísticamente significativas en las aguas centrales, dominadas por un cambio hacia un estado más salino y cálido después del forzamiento atmosférico extremo del invierno de 2005. El nivel de transición entre el agua mediterránea y el agua del mar de Labrador mostró una tendencia a largo plazo dominada por el cambio hacia condiciones más frías y menos salinas entre 2008 y 2010. La variabilidad observada por debajo de los 2000 dbar es muy pequeña, oscilando en los límites de lo que es medible. Las propiedades de las masas de agua en la sección están fuertemente influenciadas por el estado de la NAO, respondiendo de manera rápida y local al fuerte índice NAO positivo de los inviernos 2007-2008 y 2012 (induciendo enfriamiento y disminución de la salinidad), y el extremadamente negativo índice NAO en el invierno de 2010 (induciendo condiciones más cálidas y salinas de las aguas intermedias). Los resultados son consistentes con el conocimiento actual de la circulación del Atlántico Norte y su respuesta a la circulación atmosférica.

Este trabajo pone de manifiesto la importancia que tiene el mantenimiento de los programas de monitorización continua de la variabilidad del océano profundo a través de la repetición periódica de secciones hidrográficas. Se establecen como una herramienta esencial para el análisis de los cambios del océano profundo en escalas interanuales a decadales, incluyendo fuertes cambios impredecibles esen-

ciales para predecir futuros cambios en el clima local o global. Se espera que los resultados derivados de este programa de monitorización contribuyan a futuras revisiones y un mayor desarrollo de la estrategia de monitorización del clima a largo plazo en España.

Las velocidades geostroficadas de las corrientes se computaron a partir de las medidas de CTD realizadas a lo largo de la sección de Finisterre durante el invierno (febrero) y verano (septiembre) de 2008, eligiendo un nivel de referencia en $\gamma^n = 28.072 \text{kgm}^{-3}$ (~ 3000 m) o el nivel común más profundo en estaciones más someras cerca del talud o cerca del Banco de Galicia. Los campos geostroficados muestran un comportamiento alternante de corrientes hacia el norte y hacia el sur evidenciando estructuras de mesoscala no observadas por las medidas directas de LADCP, las cuales muestran una circulación más barotrópica de acuerdo al fondeo fijo de correntómetros situado en 43°N , 11°O . Como consecuencia, el campo geostrofico parece subestimar el transporte de masa a través de la sección de Finisterre en un orden de magnitud en invierno, incluso invirtiéndose el flujo durante el verano.

Las medidas de LADCP revelaron un transporte de masa hacia el sur más intenso en verano (2.2 Sv) que en invierno (1.30 Sv) a través de la sección de Finisterre. Los hallazgos incluyen un flujo hacia el norte extendiéndose en la plataforma y el talud hasta los 1500 dbar durante el verano, en aparente oposición con el flujo hacia el sur esperado característico del régimen de afloramiento en verano, y una recirculación anticiclónica de la columna de agua alrededor del banco de Galicia. Los resultados son consistentes con la propagación de la vena de agua mediterránea sobre el banco de Galicia en invierno (y posterior recirculación) y la restricción de su core contra el talud en verano.

Este estudio revela también la importancia que tiene la medida sistemática de la velocidad de las corrientes a la hora de determinar los patrones de circulación real del océano en la región, tanto para la obtención de medidas directas como para servir de herramienta en la mejora de los modelos de circulación regional, en una región caracterizada por un amplio rango de escalas de variabilidad y dinámica oceánica.

CHAPTER 8

Future work

- To maintain the annual sampling of the deep ocean hydrography and circulation along the Finisterre section through the series of oceanographic cruises *Radiales Profundas*, aiming for the establishment of a continuous and systematic temporal register of water mass property variations in this region of the North Atlantic Iberian margin.
- It is suggested the zonal extension of the Finisterre section towards the Central North Atlantic in order to further study the Mediterranean Water spreading in the North Atlantic.
- Moreover, given that the observed changes at the deepest levels are very small, it is also suggested to limit the regular sampling down to 2000 m depth (where main changes are observed), repeating the whole section down to the ocean bottom (more than 5000 m) every 2-3 years.
- To implement the repetition of the section with measurements from gliders, combining the oceanographic cruises on board research vessels with the deployment of such state-of-the-art autonomous vehicles.
- To get the time series of oceanic transports from direct (LADCP) and indirect (CTD, geostrophy) current velocity measurements during the whole time period 2003-2013, comparing the results obtained from the three methods as a function of conditions and available data during each cruise.
- To apply a higher-resolution ($0.125^\circ \times 0.125^\circ$) tidal barotropic model to interpolate the effects of tides at specific sampling locations. To consider the importance of the baroclinic component (internal tides) for the study

of ocean circulation at near shelf-slope regions characterized by a complex bathymetry.

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