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AN OVERVIEW OF THE LITERATURE CONCERNING THE OCEANOGRAPHY OF THE EASTERN NORTH ATLANTIC REGION

Evan Mason, Steve Coombs and Paulo B. Oliveira

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AN OVERVIEW OF THE LITERATURE CONCERNING THE OCEANOGRAPHY OF THE EASTERN NORTH ATLANTIC REGION

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ABSTRACT

An overview of the main oceanographic features of the eastern North Atlantic boundary, with emphasis toward the upper layers, is presented. The principal features discussed are: water mass boundaries; forcing by wind, density and tides; topographic features and effects; fronts; upwelling and downwelling; poleward flows; coastal currents; eddies. The occurrence and spatial and seasonal variability of these features is described in five regional sections: Celtic Sea and western English Channel; Bay of Biscay; western Iberia; Gulf of Cadiz; northwest Africa. This paper is intended to provide a base of physical oceanographic knowledge in support of research in fisheries, biological and chemical oceanography, and marine biology.

Key words: Northeast Atlantic; oceanography; physical processes; seasonal variability.

RESUMO

Título: Revisão da literatura sobre a oceanografia do Atlântico Nordeste

Apresenta-se uma revisão das principais características oceanográficas da região oriental do Atlântico Norte, com destaque para as camadas superiores. Os principais aspectos discutidos são: as massas de água e suas fronteiras; forçamento do vento, gradientes de densidade, marés; topografia e seus efeitos; zonas frontais, afloramento e convergência costeira; escoamento para o pólo, correntes costeiras e vórtices de mesoscala. A ocorrência, a variabilidade sazonal e espacial destas estruturas e processos são descritas em cinco sub-regiões: Mar Céltico e parte ocidental do Canal da Mancha, Golfo da Biscaia, costa ocidental da Peninsula Ibérica, Golfo de Cádiz e Noroeste de África. Este trabalho visa fornecer uma compilação dos conhecimentos sobre a oceanografia física da região para apoio a estudos de biologia marinha e oceanografia das pescas, biológica e química.

Palavras chave: Atlântico Nordeste; oceanografia; processos físicos; variabilidade sazonal.

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1 INTRODUCTION

The purpose of this review is to identify and describe the different oceanographic regimes that characterise the near-shore, shelf, slope and offshore regions of the northeastern Atlantic. Many of the features and processes described are potential barriers or concentration loci, influencing the dispersal and survival of fish egg and larval stages and the migration of adults, and also promoting mixing and exchange of characteristic water properties and constituents, ultimately resulting in the regional distributions and differentiation of fish populations that is observed.

1.1 Topography

The section of the eastern boundary of the North Atlantic discussed in this review extends from the Senegalese coast at about 15°N to the northern Celtic Sea at about 53°N (Figure 1). The area of interest therefore encompasses a range of climatic conditions, from equatorial to the temperate mid-latitudes. The extent of the continental shelf is the largest for any of the major eastern boundary upwelling systems. Shelf widths are typically narrow (10-100 km wide), with the important exception of the tidally-energetic northwest European shelf, which is semi-enclosed and has dimensions of the order of 1000 km. The coastal orientation is predominantly meridional or has a strong meridional orientation, barring the northern and southern Iberian coasts which are zonal, and the complicated geometry of the Celtic Sea. A number of persistent hydrographical features are associated with the shelf topography, such as upwelling filaments that extend offshore from capes and headlands. Several submarine canyons, which are associated with eddy generation and are also sites of coastal sediment deposition, are found along the Iberian shelf (e.g. Fiúza *et al.*, 1998; Peliz *et al.*, 2002) and the northwest African shelf (e.g. Barton *et al.*, 2004). A further feature of major topographical importance is the Canaries archipelago at 28°N, which lies at the transition zone between coastal and oceanic waters.

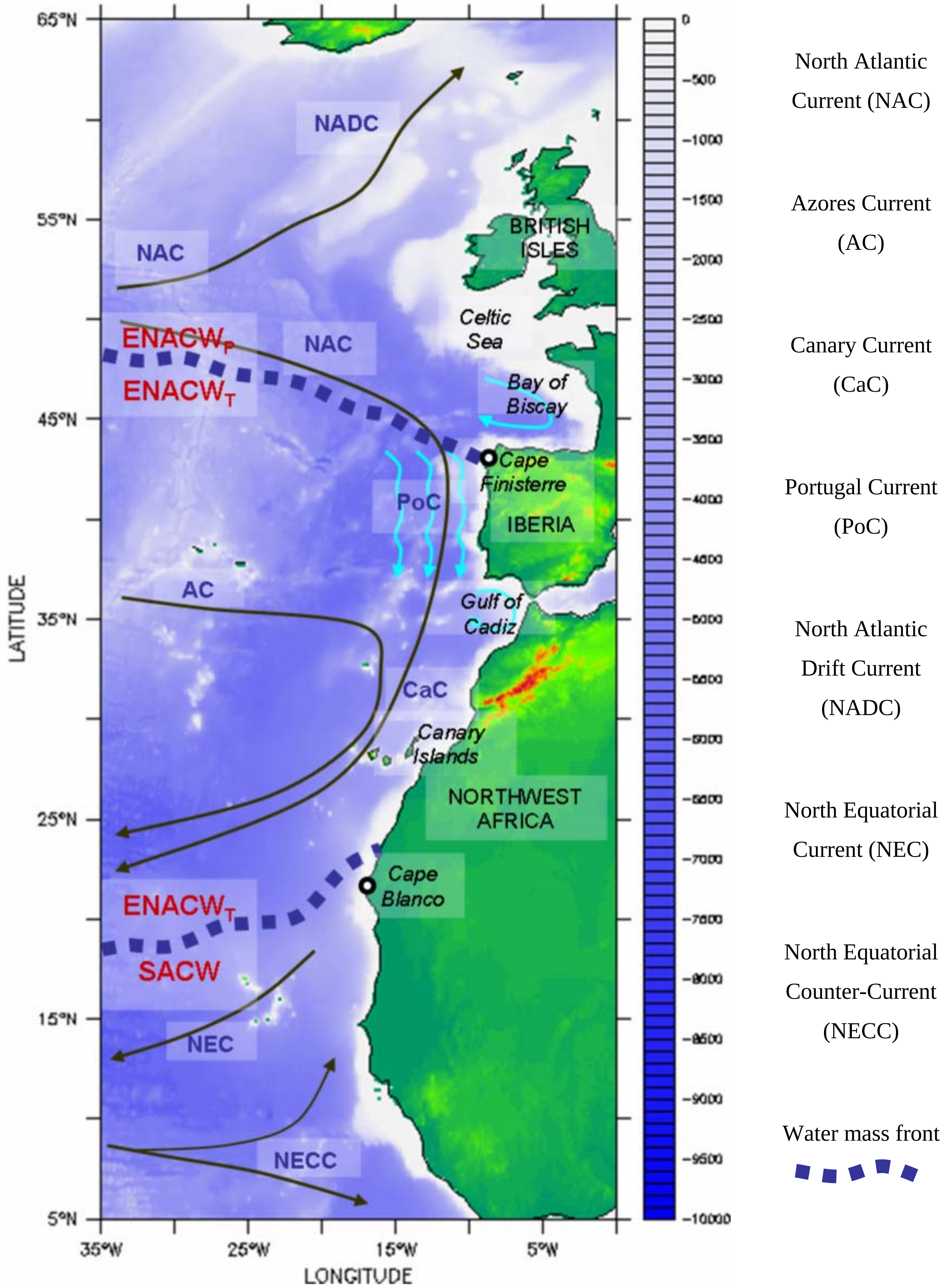

Figure 1. **The eastern North Atlantic region. The principal water masses are eastern North Atlantic Central Water of sub-polar ($ENACW_P$) and sub-tropical ($ENACW_T$) origins, and South Atlantic Central Water (SACW). The main large-scale surface currents are the North Atlantic Current (NAC), the Azores Current (AC), the Canary Current (CaC) and the Portugal Current (PoC). Also shown are the North Atlantic Drift Current (NADC), the North Equatorial Current (NEC) and the North Equatorial Counter-Current (NECC). The general circulation of the Bay of Biscay and the Gulf of Cadiz are indicated.**

1.2 Wind forcing

The near-surface circulation of both the North and South Atlantic Oceans is primarily driven by the wind. The preferential north-south orientation of the continents that bound the Atlantic Ocean lead to meridional eastern and (intensified) western boundary currents which, together with the wind-induced zonal currents – westward flow under the trade winds, and eastward flow under the mid-latitude westerly winds, centred about a large sub-tropical atmospheric high-pressure cell – form the closed oceanic gyres. Over the sub-tropical African coasts, the wind stress displays a predominantly equatorward (upwelling-favourable) component. However, off the Iberian Peninsula (IP) which, at 36-44°N, lies at the northern limit of the trade wind belt, meridional shifts in the atmospheric highs mean that the equatorward wind forcing reverses seasonally to become poleward in autumn and winter. Further to the north, over the northwest European shelf, the wind stress has a more westerly component; the winds here are energetic and prevail for much of the year.

1.3 Principal currents

The main large-scale currents associated with the eastern part of the anticyclonic North Atlantic sub-tropical gyre are the North Atlantic Current (NAC), the Azores Current (AC) and the Canary Current (CaC). These currents have transports of 35, 12 and 5 Sv (1 Sverdrup = 10^6 m^3 s^{-1}), respectively (Stramma, 2001). The NAC, to the north of Iberia between 48-53°N, and the AC, south of Iberia between 34-35°N, are basin-scale currents; exchange between the two is enabled by the broad, slow, generally southward-flowing Portugal Current (PoC), which transports about 3 Sv (Pérez *et al.*, 2001). A major portion of the NAC heads northeastward, becoming the North Atlantic Drift Current (NADC) located between Iceland and the British Isles; the remainder continues eastward toward northern Europe. The CaC is supplied by an eastward branch of the AC, which passes north of and around Madeira; however, the CaC also receives a small contribution from the PoC (Barton, 2001). The total transport of the CaC is seasonally constant, although there is geostrophic circulation variability; the CaC is stronger in summer near the African coast, while in winter it is stronger west of the Canary Islands. A further feature is a subsurface, density-driven poleward current found all along the northwest European and northwest African shelf edges; in IP waters this current is named the Iberian Poleward Current (IPC) and has a seasonal surface expression.

Finally, dense Mediterranean Water (MW) leaves the Strait of Gibraltar and descends to below 1000 m in the Gulf of Cadiz. With a characteristically-high salinity and temperature signature, MW spreads as a tongue far into the North Atlantic, with a portion also forming a poleward undercurrent that flows along the slope of the IP.

1.4 Water masses and the major frontal boundaries

At the scale of the area here reviewed, two major zonally-orientated, oceanic water mass fronts are found. Much of the region (above 600 m) is dominated by high-salinity North Atlantic Central Water (NACW), although there is considerable latitudinal variability: between the Canaries and Iberia, sub-tropical Eastern North Atlantic Central Water ($ENACW_T$) has been identified (e.g. Fiúza and Halpern, 1982; Fiúza, 1984; Ríos *et al*., 1992) that is thought to originate at a frontal region near the Azores, spreading with the general eastward flow towards Iberia; at the northern $ENACW_T$ limit, off Cape Finisterre (43°N), a subsurface front exists between the $ENACW_T$ and cooler, fresher $ENACW_P$ of sub-polar origin (e.g. Fraga *et al*., 1982). To the south of Cape Blanco (21°N) there is a major frontal region between less saline, higher nutrient South Central Atlantic Water (SACW), and the slightly warmer $ENACW_T$ transported southward by the CaC. In the region of each of these water mass boundaries, i.e. Cape Finisterre and Cape Blanco, intense upwelling is experienced (Fraga, 1981).

2 PROCESSES AND PHENOMENA

In the following section, a summary of the main physical processes and phenomena found over the study area is presented. The descriptions are necessarily brief, with the focus directed towards highlighting the relevance of each particular process in terms of its potential to affect biological or chemical distributions or processes. The features described are common to all eastern boundary systems.

2.1 The coastal boundary layer

Near-shore flow speeds are typically slow along coastal boundaries: alongshore flows are retarded by the drag of a shallow bottom and coastline irregularities, whilst cross-shore flows are inhibited by the proximity of the solid coastal wall. Persistent cross-shore surface flows near to the coast occur only when there is vertical circulation, such as during wind-driven upwelling circulation (see 2.4 below) or in the case of estuarine circulation, although even then it remains weak near-shore. The implication, therefore, is that, in the absence of additional forcing (such as offshore Ekman transport; see 2.4 below), the time taken for propagules (such as fish eggs and larvae) to leave the coastal boundary layer and enter stronger alongshore flow regions over the shelf is significant, and is inversely proportional to their original distance from the coastline. Wolanski (1994) used the term "sticky water" to describe the tendency for near-shore retention of larvae as a consequence of these weak currents and associated weak dispersion. Scales (and strengths) at which the coastal boundary layer may be observed are variable, but it is consistently evident as a reduced mean and reduced standard deviation in both alongshore and cross-shore flows. See Largier (2003) for more on the coastal boundary layer.

2.2 Fronts

Fronts are ubiquitous features of eastern ocean boundaries, of which there exist several different types; their presence and formation is often accompanied by current jets, convergences, eddies, subsurface intrusions and lenses, and they are associated with strong mixing and stirring. Fronts are narrow zones of enhanced gradients of water properties, horizontal or vertical, that may be observed over a large range of scales (100 m to 10000 km along-front; 10 m to 100 km across-front; 1 m to 1 km down-front), and they persist from hours to many years. Fronts separate water masses that might change little over hundreds of kilometres, and yet, frontal gradients might be as sharp as 10 °C and 1 unit of salinity over distances of the order of 100 m (Belkin, 2002). Frontal boundaries are unstable features, often taking on a meandering convoluted character. In general, frontal regions are more biologically productive relative to surrounding waters, and adult fish, larvae and eggs may benefit from the associated increases in nutrient and plankton concentrations. Shore-parallel fronts are recognised as barriers, reducing the export of particles from coastal waters overlying the continental shelf (e.g. Bakun, 1996). Thus, although fronts may have enhanced biological productivity and locally energetic mixing, they nevertheless represent water mass boundaries and are potential barriers to planktonic dispersal and adult migration.

2.2.1 Tidal mixing fronts

A tidal mixing front is a surface-water temperature discontinuity that marks a transition from a tidally mixed to a vertically stratified water column. In continental shelf seas, the energy of shoaling tidal waves is dissipated by seabed friction, and this energy is sometimes sufficient to break down vertical stratification. Whether the breakdown is complete, i.e. whether it extends all the way to the surface, depends on the energy of the wave, the water depth and the strength of the stratification. Over the extensive northwest European shelf, the strength of stratification is a response to the input of surface buoyancy due to (seasonal, solar) heating. A widely used criterion to determine the location of a tidal front is the Simpson-Hunter (SH) parameter:

$$SH = \log_{10}\left(\frac{H}{\overline{U}^{3}}\right),$$

where U is the mean flow and H is the water depth (Simpson and Hunter, 1974). A tidal front is usually indicated by SH = 1.8 to 2.0. Pingree *et al*. (1982) present a map of the SH parameter distribution (see their Figure 5) showing the mixed and frontal regions off Ushant. Tidal mixing fronts are, therefore, relatively stationary and seasonally-enhanced features of shelf seas, found in the Celtic Sea and the English Channel (see 3.1 below), which, in addition, introduce significant baroclinicity into the circulation, constituting an important seasonal aspect. Since tidal mixing fronts occur seasonally, they might be expected to have relatively less influence on fish stock structures than more permanent features.

2.2.2 Shelf-break fronts

In contrast to tidal mixing fronts, shelf-break fronts generally persist throughout the year and their locations are less rigidly controlled by tides and topography. Shelf-break fronts separate (usually fresher) shelf waters from slope and oceanic waters. They are primarily haline structures, but may also have a seasonal pattern of thermal structure; in upwelling systems for instance, especially where the upwelling front is locked to the shelf edge. In the Celtic Sea in summer, a band of cold water observed along the shelf break is thought to result from enhanced shelf-break mixing due to dissipation of internal tidal energy. Shelf-break fronts are likely to have a relatively low impact on (shelf-based) fish stock structures (which usually concern discrimination in the along shelf direction) due to their orientation (representing potential boundaries in the cross-shelf direction) and position over the shelf-break.

2.3 Internal waves

Waves are clearly features of the ocean surface, but they can also occur as internal waves on density interfaces (such as at the thermocline). In the deep ocean, internal waves do not generally affect the sea surface where they are not usually observable although, at the shelf-edge for example, "breaking" internal waves can have a surface expression. Internal waves typically have much larger amplitudes and longer periods than surface gravity waves (from 10-20 minutes to several hours, compared to several seconds or minutes for surface gravity waves). Internal waves transport energy and momentum 'elsewhere', and this energy may become remotely available for mixing. Internal tides are internal waves with the same period as the tide. When internal waves enter regions of shallower water (i.e. passing over a shallow bank or reaching the continental shelf) the denser, colder (and nutrient rich) water within the wave rises and, in some cases, the wave may break. Internal waves may often be detected by the use of synthetic aperture radar (SAR) imagery (e.g. da Silva *et al.*, 1998). Ryan *et al.* (2005) report on the biological impact of internal tides in Monterey Canyon on the U.S. West Coast: internal tides cause layers of cold water to rise and fall by as much as 100 meters over a few hours, and this dense water sometimes "sloshes" out over the edge of the canyon. The dense water was found to contain sea-bottom sediments, which are often rich in nutrients such as nitrate. In the present study region, internal waves may be important in the Celtic Sea (see 3.1 below), and also off western Iberia (3.3.1).

2.4 Upwelling and downwelling

Wind-driven coastal upwelling (downwelling), the basic mechanisms of which are well understood, occurs when equatorward (poleward) winds induce net offshore (onshore) surface Ekman transport, resulting in transport divergence (convergence) near the coast. A particular interest of upwelling and downwelling circulations concerns the role of their secondary, cross-shelf circulation, which redistributes not only heat and salt, but also nutrients and biological fields. Upwelling systems affect higher trophic levels through food chain-related processes (enhanced primary production), and also through the opportunities provided by the related current system for transport of eggs, larvae and prey. It should be noted that rather less is known about upwelling- and downwelling-related dynamics at scales smaller than the mesoscale, and these may significantly impact biological responses (for example, see Graham and Largier, 1997). A comprehensive review of upwelling dynamics is provided by Brink (1998).

2.4.1 Upwelling

The key features of a fully-developed/steady-state upwelling situation are (1) offshore surface layer transport away from the coast, (2) zero cross-shore transport at the coast, (3) upwelling of isopycnals close to the coast, and (4) a geostrophic equatorward coastal current. The first of these features is often cited as a source of (offshore) dispersion of passive particles (e.g. eggs/larvae); hence, in general, a region of persistent upwelling and consequent offshore advection is deemed likely to pose a significant boundary for fish populations. However, a number of finer details are relevant. Modelling studies have shown that, over stratified shelves (summer), the upwelled isopycnals form a frontal region that moves offshore at a constant rate as the upwelling develops (Austin and Lentz, 2002). This creates an “inner shelf” region, inshore of the upwelling front, which typically has low cross-shelf circulation. This frontal region may act as a barrier, retaining propagules near to the coast (i.e. a local biological retention zone), when the observed environmental conditions (the upwelling) would suggest dispersal offshore. Furthermore, a study by Lentz (2001) has shown that the upwelling responses may break down altogether in the absence of stratification (such as in winter or during strong mixing).

Upwelling systems display considerable alongshore (3-dimensional) variability. Alongshore gradients in the effective alongshore wind stress, in bathymetry or in stratification, may all lead to differences in alongshore transport, which in turn create pressure gradients counter to the mean wind stress. Such instabilities in the equatorward coastal current may result in the formation of upwelling filaments, which are large-scale, cold water structures that extend hundreds of kilometres offshore. They penetrate to depths of 200 m and support strong offshore velocities (squirts). Filaments can be expected to contribute significantly to exchange between the shelf and ocean (Huthnance, 1995), and hence to loss (or gain) of particles from the shelf, which, if persistent, represent a potentially strong boundary for fish populations. A further pertinent feature is upwelling “shadow zones” (Graham and Largier, 1997). Shadow zones are typically found in the lee of capes or promontories, and feature anomalously warm water at the surface combined with high stratification. A cyclonic flow pattern is observed, and a frontal region between warmer shadow-zone water (inshore) and colder upwelled water (offshore) develops. Upwelling shadow zones are characteristic small scale features of upwelling regions, which may play a disproportionately significant role in biological productivity: their well-defined circulation and stratification promote retention of propagules that otherwise may be advected offshore or downstream (e.g. Wing *et al*., 1998; see also 3.3.1.1 below).

2.4.2 Downwelling

Initial responses to downwelling-favourable winds are essentially the same as for the upwelling case, except that the cross-shelf and alongshelf flows are reversed. An “inner shelf” region inshore of the downwelling front is established, which is characterised by a lack of vertical stratification and strong mixing. Cross-shelf circulation in this region is weak; the momentum balance is in the alongshore direction, poleward at the surface. Offshore of the downwelling front, surface flow is onshore. The result of these observations is that, under downwelling-favourable wind conditions, propagules at the surface tend to move onshore until they enter the inner shelf region, where they are advected polewards. Thus, downwelling favours alongshore transport of propagules, although conditions for biological production tend to be unfavourable.

2.5 Poleward flow

Poleward currents are typical of eastern boundaries and upwelling systems, and may be defined as persistent poleward flows of restricted width and thickness, which are bound to the continental slope, run counter to the dominant regional circulation, and may at certain times and locations extend upwards to the sea surface. Because they often flow against prevailing equatorward winds, poleward currents may be influenced by a larger-scale, non-local forcing mechanism; candidates include an along-shore surface slope (sea level declining poleward) or an offshore pressure gradient (sea level rising toward the coast). However, some of these flows are a result of region-specific local forcing, such as in the case of buoyancy-driven coastal currents (see 2.6 below). Eastern boundary poleward flows transport warm water poleward, consequently maintaining lateral density gradients across them. Poleward flows are generally located over the shelf-break or continental slope, they are 20-100 km wide, and have speeds of about 0.1 m s^{-1}. Most eastern boundary poleward flows are subsurface flows (undercurrents), with cores typically at about 300 m depth; in upwelling regions they may be capped by equatorward-flowing surface currents. In upwelling systems, it is important to note that the upwelled water is often drawn from the poleward undercurrent. Typically, poleward flows are below the depths (<50 m) at which the early life stages of pelagic fish generally occur and will, therefore, have little direct effect on structuring the population; any influence will tend to reduce population heterogeneity due to promoting along-shelf transport. See Neshyba *et al*. (1989) for a more complete review of poleward flows.

2.6 Buoyancy-driven coastal currents

Buoyancy-driven coastal currents have a typically wedge-shaped density structure, where the sea surface slopes downwards offshore. Buoyancy is a function of temperature and/or salinity, with freshwater outflows from rivers being a common source of buoyancy in these coastal currents. Provided that the coastal current has a sufficiently large volume, geostrophy forces it to remain trapped to the coast, flowing poleward along eastern boundaries. The width of these currents is usually of the order of the baroclinic Rossby radius, thus they are narrower and more intense in the higher latitudes. Baroclinic instabilities at the frontal boundaries between buoyancy-driven coastal currents and oceanic waters further offshore, sometimes result in the spawning of Rossby-radius scale eddies. Such eddies, typically found in cyclone-anticyclone pairs, are responsible for significant shelf-ocean exchange of water, and also therefore, nutrients and plankton. Considerable variability in alongshore transport of water and propagules also occurs, as a result of the seasonal nature of buoyancy–

driven coastal currents (highest rainfalls generally occur during the winter for the study area). Wind and other forcing mechanisms (density) may also affect buoyancy-driven coastal currents: upwelling-favourable winds tend to spread buoyant plumes offshore; whilst downwelling-favourable winds do the opposite (see 2.4 above). Where buoyant plumes are large and persistent, such as for major river outflows, there is the potential both for transport within the plume and for the plume to act as a barrier to biological transfer across it.

2.7 Topographic features

Topographic features at the seabed, or the coastline itself, often play important roles in the formation of a number of the physical phenomena described in section 2, such as eddies and upwelling-related filaments and shadow zones. In these cases, the occurrence of these phenomena is often predictable under certain conditions, and therefore the phenomena may constitute a significant link in the life cycles of various fish species, where retention and/or entrainment of eggs and larvae are important. Topographic features clearly contribute to the high spatial variability, in terms of convergence and divergence, found in the coastal ocean. Localised and/or intensified coastal upwellings are often associated with the presence of capes or promontories, as is the formation of upwelling filaments (see 2.4.1 above). Coastal promontories play an important role in triggering instabilities in upwelling-related coastal jets, leading to downstream shadow zones (see 2.4.1 above). Coastal embayments may be linked to the retention of eggs and larvae (see 2.8 below for more), because they can be effective at retaining water and reducing alongshore flow (e.g. Monteiro and Largier, 1999). Offshore banks, such as the Porcupine Bank in the Celtic Sea, are responsible for quasipermanent anticyclonic circulations; these are sometimes referred to as Taylor Column circulations (e.g. Heath, 1992). Closed around-bank flows may also be induced by tidal rectification (e.g. Huthnance, 1973). These processes promote the retention of biogenic material and hence enhanced productivity in these regions (e.g. White *et al.*, 1998). Tidal current interactions with offshore banks also promote the breakdown of stratification; these local mixing effects may be important in determining chlorophyll levels in the vicinity of the banks. Offshore canyons that intersect the continental slope may significantly modify the local circulation: these permit increased cross-shelf motion by allowing the geostrophic constraint on steep slopes to be broken. They also produce enhanced upwelling, typically at their equatorward sides. Offshore islands, such as the Canary Island archipelago (see 3.3.3.1 below), are associated with several effects, such as eddy generation and pronounced, localised Ekman pumping. A final topographic feature is the Joint Effect of Baroclinicity and Relief (JEBAR) (Huthnance, 1995; Hill *et al*., 1998). JEBAR is thought to play a role in the generation of slope currents: the basic idea is that an offshore pressure gradient develops as a result of differences between shelf and oceanic meridional gradients in steric height (sea level contractions/expansions related to temperature), which drives poleward along-slope flow. See Trowbridge *et al*. (1998) for a review of topographic effects in the coastal ocean.

2.8 Eddies

Where water flows along a curved or uneven coastline, it may separate from the coast and generate an eddy. There are many examples of these including, within the present study area, eddies in the Bay of Biscay (see 3.2 below), off western Iberia (see 3.3.1), off northwest Africa (see 3.3.3) and at the Canary Islands (see 3.3.3.1). The scales of these eddies are variable, generally from 10 to 100 km. Water recirculation within eddies means that propagules can be retained over significant time scales and thus, at specific locations, local recruitment may be enhanced; eddies generated by offshore islands have so far provided the best examples. In contrast to island eddies, the retentive ability of coastal eddies is reduced as a consequence of greater current variability and bottom friction. Nevertheless, retention within headland eddies, and also along semi-enclosed bays, remains significant (e.g. Graham and Largier, 1997; Largier, 2003). Eddy retention zones, as well as retaining local production, can also entrain propagules, in this way enhancing recruitment of non-local eggs and larvae. Eddies also form along the boundaries between coastal currents and offshore waters, causing complex variations in the dispersal of planktonic eggs and larvae. Indeed, it should be recognised that eddies are as much agents of dispersal as retention: eddies retain water with particular properties (and propagules) and also transport it to new locations.

The eastern North Atlantic features a number of particular types of eddies, including “MEDDIES” and “SWODDIES”. Mediterranean Water (MW) outflow follows the contours of the southwestern Iberian shelf (Bower *et al*., 2002); as it progresses around Cape St. Vincent, the undercurrent is destabilised (usually when passing over and through canyons such the Portimão Canyon in the Gulf of Cadiz), generating deep, mesoscale anticyclones (Serra and Ambar, 2002) called MEDDIES (McDowell and Rossby, 1978; Armi and Stommel, 1983). MEDDIES generally propagate southwestward from their site of generation, and may persist for several years. They dissipate slowly by mixing, or more rapidly disappear by colliding with seamounts such as those of the Horseshoe Seamount chain southwest of Cape St. Vincent (e.g. Richardson and Tychensky, 1998; Richardson *et al*., 2000). Slope Water Oceanic Eddies (SWODDIES) is the term used to describe anticyclones, sometimes with one or two cyclonic companions, which form during the separation of a poleward slope current at a number of the more prominent capes along northern Iberia and southern France (Pingree and Le Cann, 1992). SWODDIES are relatively long-lived features, lasting up to 1 year, with depths of over 1000 m, which migrate offslope into the deep ocean. The area south of the Canary Islands is implicated in the generation of eddies that travel westwards along 26°N (Pingree and Garcia-Soto, 2004); on average these structures persist for 2.8 years, travelling ~3300 km to ~50°W, i.e. more than half way across the North Atlantic Ocean.

2.9 Summary

A wide range of physical oceanographic features have been described in this section. At particular times and locations, one or more of these features may operate at the same time, leading to complex circulations at a variety of spatial and temporal scales. In the next section, the occurrence and relevance of these phenomena are explored by geographical region.

3 REGIONAL DESCRIPTIONS

In this section, the study area is divided into 5 regions for a focussed description and discussion. The sections deal with each region from the north to the south: Celtic Sea and English Channel (3.1); Bay of Biscay (3.2); western Iberian Peninsula (3.3.1); Gulf of Cadiz (3.3.2); and northwest Africa (3.3.3). The northern regions experience greater seasonal variability than the southern regions. Where appropriate, descriptions are subdivided into seasonal descriptions to further explain the seasonality. In this way, the review proceeds from a consideration of the large-scale picture down to smaller spatial and temporal scales.

3.1 The Celtic Sea and English Channel

The Celtic Sea and the English Channel (Figure 2) are near the northern limit of significant spawning of certain fish species, such as sardine. These two adjacent areas are located between 48-52°N, and share a considerable area of the topographically-complex northwest European shelf (<200 m). This shelf widens dramatically north of the 180-km wide Armorican Shelf (~48°N) of the northern Bay of Biscay. The Celtic Sea is bounded to the south and west by an extensive, 1000-km shelf edge, which descends rapidly to depths greater than 1000 m. In the west, the shelf edge is punctuated by the Porcupine Seabight, north of which lie Porcupine Bank. The shelf region, where the residual circulation is not particularly strong, receives large tidal energy input from the Atlantic Ocean, generating strong tidal currents that provide a dominant source of mechanical energy. Their location in the mid-latitudes means that the Celtic Sea and the English Channel are subject to a strong seasonal surface heating and cooling, and to energetic (westerly) winds that prevail for much of the year. Freshwater buoyancy input is generally low, with some river input from the coasts of Ireland and southwestern Great Britain, and also the continental coast of the English Channel, most notably the Seine in the eastern English Channel. To the north, the Celtic Sea is bounded by the Irish Sea, which feeds the Scottish Coastal Current through the North Passage; and to the east the English Channel provides a connection to the North Sea. A recent review of the oceanography of the Celtic Sea and English Channel is provided by Simpson (1998).

Tidal waves proceed from the deep ocean onto the shelf, whereupon they are reflected and amplified to varying degrees by the coastal and seabed topography. The constrained flow in the English Channel results in particularly strong tidal responses, especially at the French coast (due to the action of the Coriolis force) where tidal elevations in some locations exceed 8 m (e.g. St. Malo). Tidal currents here are rectilinear and frequently greater than 1 m s^{-1}, inducing strong frictional stresses that extract energy from the tidal wave. In the less restricted Celtic Sea, the tidal currents are less intense and exhibit anticyclonic (clockwise) rotation. Further north towards the Bristol Channel, the tidal flows become progressively stronger, exhibiting strong cyclonic rotation.

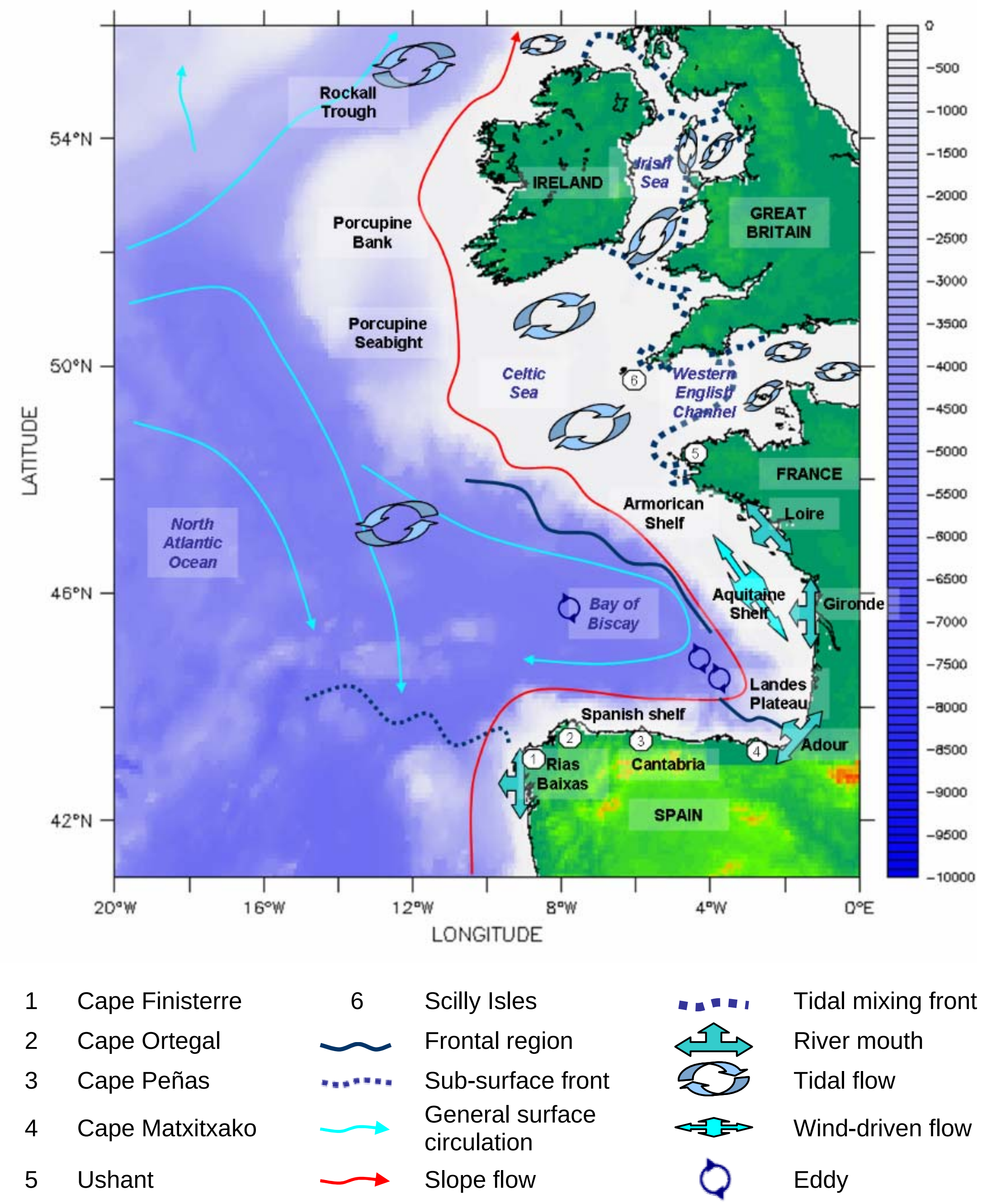

Figure 2. General diagram showing the principal physical features of the Celtic Sea and Bay of Biscay.

Seasonal atmospheric trends over the Celtic Sea follow closely the patterns induced by the large scale climatology of the eastern North Atlantic (see 1.1 above), resulting in the predominance of relatively strong southwesterly winds for much of the year. Wind direction, however, is much more variable than in areas to the south, even in summer, because of the frequent passage, usually southwest to northeast, of low-pressure systems. In summer, winds are significantly weaker and generally more westerly in direction. The mean annual wind stress for the years 1978-1982 was 0.06 Pa (equivalent to a wind speed of ~6.4 m s^{-1}) in a direction towards 081°T (Pingree and Le Cann, 1989). Wind stress and pressure gradients due to atmospheric pressure changes and induced surface slopes, were shown by Thomson and Pugh (1986) to contribute only about half of the observed regional subtidal current variance; with the remainder, the authors suggest, resulting not from oceanic influence, but from forcing via the Irish and North Sea connections. Pingree and Le Cann (1989) also note that the observed variability between winds and surface currents is a result of the contribution of non-local winds to the surface current field. In areas of strong tidal influence, the residual currents are damped by (bottom) frictional effects; in this way, transient responses to wind forcing are inhibited. The frequent passing of low-pressure systems through the Celtic Sea area sometimes induces storm surges; the currents associated with these events are smaller than tidal currents by a factor of about 10. Transient coastal upwelling has been reported for the southwest coast of Ireland (Edwards *et al*., 1996).

The inputs of mechanical energy by tides and wind in the Celtic Sea and western English Channel area result in a highly mixed homogeneous water column for much of the year, with summer stratification due to solar heating occurring only in the less energetic areas. Typically, the large tidal flows in the Channel result in the water column remaining mixed throughout the year, whilst in the Celtic Sea (and part of the western Channel) there is a varying amount of summer stratification. Further north, particularly strong tidal mixing is experienced in the North Channel region (northern boundary of the Irish Sea) where, at spring tides, tidal currents exceeding 1.5 m s^{-1} ensure almost complete vertical homogeneity over the full water depth of up to 200 m. The distribution of mixed and stratified regions can be shown by mapping of the stratification parameter (see 2.2.1 above). The transition zones between stratified and vertically-mixed waters in the Celtic Sea and western English Channel (which generally correspond to stratification parameter of 1.5; Pingree and Griffiths, 1978) are marked by sharp fronts (Figure 2), whose locations are revealed by satellite infrared imagery, with the well-mixed regions featuring cooler waters. These tidal mixing fronts are remarkably constant in location, varying in extent and position according only to weather (freshwater input) and with some adjustment in response to the spring/neap cycle. Tidal mixing fronts are major structural features of the Celtic Sea, which also play an important role in the biological environment by supporting enhanced production and biomass. High chlorophyll concentrations have been reported from one of the most prominent tidal mixing fronts, the Ushant tidal front, located off northwestern Brittany (Pingree *et al.*, 1978). Here, the complex topography that constitutes the headland and the strong tidal currents contribute to the breakdown of stratification; the strong density gradients drive along-front flows, although in many cases the flow is subject to geostrophic shear. Pingree *et al.* (1978) describe how these instabilities promote the formation of large-scale, mainly cyclonic, eddies at the Ushant front. Enhanced nutrient concentrations and biomass distributions have also been reported (Simpson and Tett, 1986) from the vicinity of the Scilly Isles (off the southwest of the U.K.), where flow around the islands induces mixing. Tidal mixing fronts are also evident where the Celtic Sea gives way to the Irish Sea. In the eastern English Channel the water column is permanently mixed, but beyond this in the southern and central North Sea seasonal stratification and mixing processes are again active.

A pronounced feature of the Celtic Sea circulation is a persistent poleward-flowing slope current (see also 1.3, 2.5, 3.2 and 3.3.1), present along the entire length of the ocean-shelf boundary, which acts to impede shelf-oceanic exchange at subtidal frequencies. This is a density-driven, topographically-steered flow, which originates near the continental slope south of Portugal (Pingree, 2002). Observations of the poleward flow in the Celtic Sea have shown it to have a characteristic speed of a few cm s^{-1} and a total transport of ~1 Sv between the shelf break and 1000 m depth (Pingree and Le Cann, 1990). The contrast between the well-defined slope flow and the weaker and directionally more variable residual shelf currents of the Celtic Sea and northern Biscay of Biscay is well demonstrated in Figure 15 of Pingree and Le Cann (1990). Poleward transport is consistent, but there are seasonal trends and variations in the vertical structure over both the Celtic and Armorican slopes to the south (Pingree and Le Cann, 1989). The superficial layers of the poleward flow over shallower regions such as the Goban Spur at 49°N, are susceptible to wind forcing so that, for instance, a westerly wind may produce a southward flow (e.g. Pingree, 2002). At the surface, along the length of the shelf break, a further feature of the Celtic Sea is the occurrence of a region of cooler water, which is sometimes clearly visible in satellite images (Dickson *et al*., 1980). This is assumed to be a consequence of an internal tide, generated at the 200-m contour, which propagates both onto the shelf and into the ocean (Pingree *et al*., 1986). These progressive waves result in the mixing of cooler, nutrient-rich waters to the sea surface, with the potential to enhance phytoplankton production by supplying more nutrients to the euphotic zone.

3.1.1 Seasonal descriptions

In the higher latitudes, seasonal variability increases, and there is greater distinction between autumn and winter, and spring and summer, in comparison with that experienced further to the south. In the following sections (see also Figure 3), therefore, we treat each season separately. The principal reference used in this section is the study by Pingree and Le Cann (1989). It should be noted that their results were based on rather short time series, 1978-1982, and so caution should be exercised when drawing conclusions about the overall character of the described circulation. Substantial annual and decadal variability is typical for this region.

3.1.1.1 Spring (March to May)

In late spring, seasonal solar warming initiates the start of the summer seasonal stratification regime (May to October), over much of the Celtic Sea and western English Channel region. Similarly, stratification is initiated over the shelf areas to the west of Ireland, and much of the southern and central North Sea. In parallel with development of the seasonal thermocline, is the initial (relatively weak) appearance of the tidal fronts, most notably around Ushant and beginning to extend across the western end of the English Channel (Figure 2).

3.1.1.2 Summer (June to August)

Summer mean wind stresses of approximately 0.05 Pa (equivalent to a wind speed of ~5.8 m s^{-1}) are typically half those of winter values, and they take a generally westerly direction. In contrast to the wind stress, surface currents reach their maximum values of 5 cm s^{-1} in late summer, and are directed to the right of the wind stress. The reason for the maximum current speeds at this time is likely to be the seasonal stratification

(May to October), which promotes the formation of a (shallow) thermocline, decreasing the depth of frictional influence. Alternatively, Pingree and Le Cann (1989) suggest the possibility of a larger surface response to the more westerly winds that prevail at this time, in comparison with the rest of the year. Pingree and Le Cann (1989) show (using 1978-1982 data) that, in late summer, the transport over the upper slope (<2500 m) reaches its maximum values (~3.5 Sv). Also by late summer, the Ushant front extends across the mouth of the English Channel, with stratified waters to the west (Celtic shelf and western Channel) and mixed to the east (central and eastern English Channel).

3.1.1.3 Autumn (September to November)

In September there is a change in the wind stress direction (measured at Ushant; Pingree and Le Cann, 1989), from the summer west-to-northwesterly regime, to a more southwesterly direction, although with considerable variability from year to year. In early autumn there is still sufficient surface heating to maintain summer stratification and the thermal fronts over the shelf region. The Ushant front is the prime example, which as autumn progresses is observed to retreat westward, past the Scilly Isles, and then to dissipate.

3.1.1.4 Winter (December to February)

Based on the time period studied by Pingree and Le Cann (1989) the mean winter wind stress values are about 0.1 Pa (equivalent to a wind speed of ~8.3 m s^{-1}), and are directed towards the northeast. Towards the end of February a change in the mean wind direction commences (and continues until April), from southwesterly to northwesterly (measured at Ushant; Pingree and Le Cann, 1989). During this transition, the wind attains a slight northerly component. Weakening of the along-slope transport also occurs between February and April. These changes might be induced by weakening of the upper-slope flow in response to the changing large-scale wind stress.

3.2 The Bay of Biscay

The Bay of Biscay (Figure 2) is an open oceanic bay located at 43.5-48.5°N, which is just outside the northern limit of the trade wind belt. A major feature of the Bay of

Biscay is a sharp discontinuity in the coastline orientation from the southern, zonal coastline of Spain to the eastern, meridional coastline of France. The Spanish shelf is generally narrow and uniform (30-40 km), while the French shelf, which is divided into the Aquitaine shelf (50-150 km) and the Armorican shelf (180 km), broadens with latitude. At the junction between the two is the deep Cap Breton Canyon, which is bounded to the north by the Landes Plateau. The complex topography of the region has a strong effect on the general circulation, and hence on environmental and biological distributions (e.g. Bardey *et al*., 1999; Koutsikopoulos and Le Cann, 1996).

For all seasons, sea surface temperature (SST) distributions are seen to increase from north to south in the Bay of Biscay. In the coldest month (March) SST varies between 9.0-12.5 °C, and in the warmest month (September) between 16-20 °C (Bardey *et al*., 1999). The greatest SST variability (winter-summer) occurs in the southeastern part of the Bay. Water masses in the upper layers (100-600 m) of the Bay of Biscay correspond to eastern North Atlantic Central Water ($ENACW_P$) of sub-polar origin, with T=10.5-12.0 °C, S=35.45-35.75. Off the northwestern tip of Iberia (Cape Finisterre, 43.5°N) a sharp, deep (~200 m) water mass front extends westward from the coast, separating the $ENACW_P$ from warmer, saltier $ENACW_T$ (sub-tropical origin) to the south (e.g. Fraga *et al*., 1982; Ríos *et al*., 1992). This front is sometimes referred to as the Galicia Front (e.g. Fiúza, 1984; Peliz and Fiúza, 1999). A permanent front is also to be found lying along the axis of the Cap Breton Canyon, which represents the differences between Spanish shelf and slope waters, and those of the Landes Plateau and French continental shelf (Valencia *et al*., 2004). Bardey *et al*. (1999) also suggest the presence of a permanent but weak thermal front between 44-47°N.

The central, oceanic area of the Bay of Biscay is characterised by weak anticyclonic circulation, which may be linked to the negative wind stress curl found here (Le Cann and Pingree, 1995). Geostrophic currents are strongest near the surface, but at 1-2 cm s^{-1} they remain weak. Variability in these flows is introduced through the frequent presence of cyclonic and anticyclonic eddies, created by interactions between shelf and slope current instabilities and the bottom topography. Slope currents of 5-10 cm s^{-1} are consistently poleward, although there is some seasonal and local variability, particularly in the vertical structure (Pingree and Le Cann, 1990). The currents over the shelf are principally governed by the wind, by buoyancy and by tides. Tidal currents are aligned cross-shore and their intensities are proportional to shelf width, so that they are stronger over the northwestern Armorican shelf (e.g. Le Cann, 1990). Over the narrow southern shelf, wind and density-driven circulation is dominant; shelf waters respond rapidly to wind stress, but wind forcing variability leads to a complex circulation, with a further contribution from the presence of irregular topographic features. Further north, where wind strength increases progressively, the wind-induced circulation over the Armorican and northern Aquitaine shelves is more coherent, with currents of typically 10 cm s^{-1}. Mean annual wind stresses of 0.07 Pa (equivalent to a wind speed of ~7.5 m s^{-1}) at 083°T were reported by Hellerman (1967). Over the inshore regions off both coasts, the general wind circulation typically runs parallel to the coastline. Northerly and easterly winds produce upwelling on the French and Spanish coasts, respectively; with upwelling tending to be higher on the French coast (Borja *et al*., 1996). Freshwater outflows from the Loire and Gironde Rivers, which both peak at nearly 2000 m^3 s^{-1} in winter (Koutsikopoulos and Le Cann, 1996), induce significant surface currents; these plumes generally flow poleward (in response to the earth's rotation), but their distribution is frequently affected by the wind. In the southeast, the Adour River also makes a significant freshwater contribution. A frontal region around Cape Matxitxako is sustained by this outflow; saltier waters are always found to the west of this cape. Comparatively lower freshwater discharges and the narrow Iberian shelf, result in the persistence of buoyant plumes along the southern boundary of Biscay being much less than over the Armorican shelf along the eastern boundary.

3.2.1 Seasonal descriptions

In this section a more detailed description of the individual seasons in the Bay of Biscay is presented (see Figure 3). In contrast to the seasonal definitions given above for the Celtic Sea and English Channel (3.1.1), the seasons in the Bay of Biscay are defined as follows: spring, April to June; summer, July to September; autumn, October to December; and winter, January to March.

3.2.1.1 Spring (April-June)

Off the northern, Spanish coast of the Bay of Biscay, downwelling-favourable winds are prevalent during the spring, although weak upwelling events do occur [Borja *et al*.

(1996) have shown that recruitment of anchovy tends to be enhanced by these episodes of upwelling].

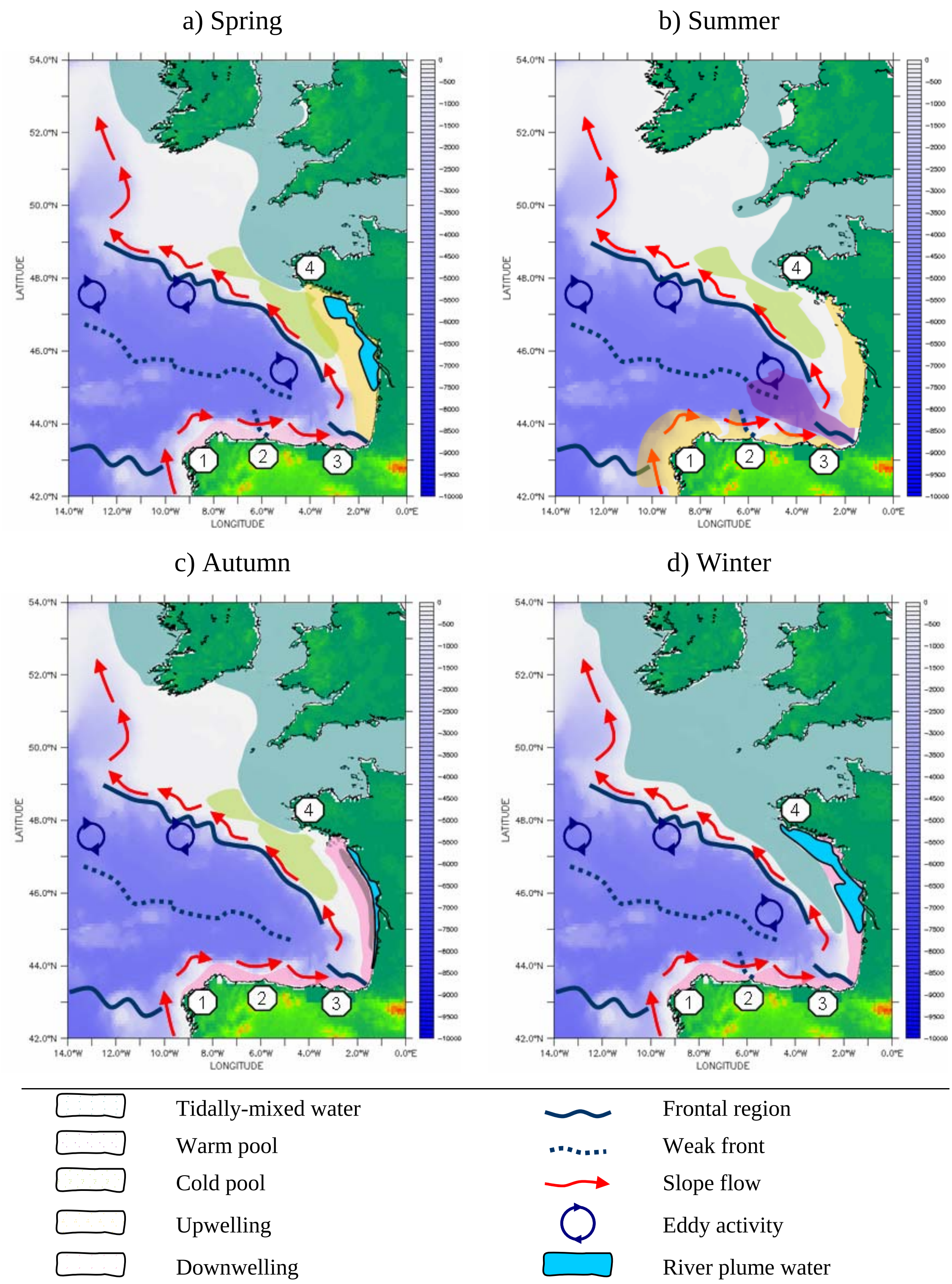

Figure 3. A series of four figures showing the seasonal evolution of the physical features that characterise the Celtic Sea and the Bay of Biscay. 1) Cape Finisterre; 2) Cape Peñas; 3) Cape Matxitxako; 4) Ushant. Adapted from Koutsikopoulos and Le Cann (1996).

Downwelling leads to a mixed water column with a depressed seasonal thermocline (which develops in spring), implying lower plankton concentrations.

Off the French coast, upwelling prevails, and here an important consequence is the offshore spreading of low-salinity freshwater from the river outflows. In this way, an extensive stable nutrient-rich environment may be created and sustained over the shelf area of the southeastern Bay of Biscay. Seasonal warming of surface waters begins in spring, and is particularly pronounced at the extreme southeastern corner of the Bay of Biscay, over the Landes Plateau. Further to the north, the development of the seasonal thermocline over the French shelf in April and May permits the identification of a persistent cold water (<12 °C) feature, which displays little temperature variation: the structure, centred over the ~100 m depth range and extending between the latitudes of southern Brittany and the Gironde estuary (46.0-48.5°N), has been named "bourrelet froid" (cold pool) by Vincent and Kurc (1969). Where river plume waters are present, the formation of this cold pool is inhibited.

3.2.1.2 Summer (July to September)

Strong vertical temperature gradients appear in summer in the Bay of Biscay, and SST variability tends to be greatest at this time. The warming, which begins in spring, is highest in the southeast and leads to the formation of a conspicuous warm pool in this area. Bardey *et al*. (1999) describe the mixing processes that take place between this mushroom-shaped, warm and chlorophyll-rich structure, and the surrounding colder waters; they speculate that this mixing could have a significant impact on the biology (see also Gil *et al*., 2002). Coastal divergence (upwelling) in the region of the Galicia Front off northwestern Iberia is particularly pronounced in summer, when the persistent northerly winds (see 3.3.1.1 below) are active off western Iberia. This upwelling can extend eastward over the northern, zonal Spanish coast past Cape Peñas at 6°W to as far 4°W, where a frontal zone normal to the coast is found (see also 3.2.1.4 below for the winter manifestation of this front). In general, the winds over the northern Spanish coast take a more easterly component during the summer, and here, this too favours upwelling. Notable upwelling centres are found at Cape Matxitxako (Bardey *et al*., 1999) and the central Cantabrian coast (Botas *et al*., 1990; Gil *et al*., 2002), which both interact strongly with the mushroom-shaped warm pool described above, and also the Landes coast (Jegou and Lazure, 1995) to the east. The upwelling on the Spanish coast is, however, markedly weaker than off western Iberia and off the French coast during the summer. Indeed, Borja *et al*. (1996) remark that upwelling in the Bay of Biscay is seldom strong enough to bring the thermocline to the surface; instead, the thermocline merely shallows (thus confining plankton within a favourable, light-rich environment). Koutsikopoulos and Le Cann (1996) describe the summertime presence of seasonal tidal thermal fronts over the northern part of the French shelf, off western Brittany, that develop as a result of interactions between bottom topography and tidal currents. Le Fèvre (1986) gives an account of the biological consequences of these structures.

3.2.1.3 Autumn (October to December)

Downwelling-favourable winds are prevalent on both the Spanish and French coasts in autumn, with higher values in the south. Mean wind speeds are considerably higher than for the spring and summer. Shelf currents are variable and complex at the French margin, where increasing river runoff at the surface responds rapidly to the prevailing wind field (Pingree and Le Cann, 1990). Decrease of surface temperatures and, in the north, increase of bottom temperatures occurs in autumn; the water column of the northern French shelf becomes nearly homogeneous as a result of vertical mixing processes. In early autumn, the tidal thermal fronts that develop in summer (see 3.2.1.2 above) are still evident. Autumn also sometimes sees the appearance of a narrow, warm water (14-16 °C) tongue close inshore along the French coast, which has been named “langue d’eau chaude” (warm water tongue) by Vincent and Kurc (1969). Le Cann (1982) suggested that this feature, which extends from the south to the Loire estuary, is a consequence of mixing between tidal and wind-induced currents, and ocean-atmosphere exchanges.

3.2.1.4 Winter (January to March)

The wintertime in the Bay of Biscay sees the largest mean wind speeds, and downwelling-favourable winds predominate on both the Spanish and French coasts. Over the French shelf regions, temperature inversions (colder water at the surface) occur, mainly as a result of the presence of cold, low-salinity river water; river discharges reach a maximum in winter, and the plumes tend to be trapped close to the coast as a result of the onshore downwelling circulation (Koutsikopoulos and Le Cann, 1996). At times, however, the surface plume waters may cover extensive areas of the shelf, particularly towards the end of winter and early spring. Over the southern, Spanish slopes, surface waters in the Bay of Biscay have their maximum poleward transport in both autumn and winter (October to February), although the flow has clear baroclinic aspects. Between December and March, when the incidence of northerly winds is infrequent off western Iberia, the relatively warm (and salty) Iberian Poleward Current (IPC, see 3.3.1 below), which transports $ENACW_T$, surfaces; it rounds Cape Finisterre, and proceeds along the shelf edge and upper slope off northern Iberia, where it is persistently observed in January and February satellite SST images (e.g. Frouin *et al*., 1990; Pingree, 1994). This tongue of anomalously warm water is referred to as the *Navidad* (Christmas), and it is implicated in the formation of SWODDIES (Slope Water Oceanic Eddies; see 2.8 above). SWODDIES are warm water, anti-cyclonic eddies, that slowly (<2 cm s^{-1}) transport slope water offshore and westward, persisting for up to a year (e.g. Pingree and Le Cann, 1992; Garcia-Soto *et al*. 2002). The frequency of SWODDIES appears to be related to winters with large negative NAO index (i.e. 1989/1990, 1995/1996, 1997/1998; Garcia-Soto *et al*. 2002). Bay of Biscay SWODDIES typically have low horizontal diffusion rates, so that material (biological) inside them tends to remain inside the eddy (e.g. Fernandez *et al*., 2004). At the easternmost extents of the Navidad/IPC influence, typically between 6°W (Cape Peñas) and 4°W, a frontal region is evident, which persists into spring (see also 3.2.1.2 above).

3.3 The Canary Upwelling Ecosystem

The extensive Canary Upwelling Ecosystem (CUE) stretches along the eastern boundary of the central North Atlantic from the northern tip of the IP at 43°N, to 10°N just south of Dakar in Senegal (see Figure 1). This range is equal to the annual meridional range of the trade wind belt. Upwelling is observed year round for the central region, whilst at the extremes the phenomenon is seasonal, following the meridional shift of the trade winds. In contrast to the previous regional descriptions, seasonal variability over the CUE is such that only spring/summer and autumn/winter descriptions are necessary. The CUE is treated below in 3 separate subsections: western Iberia; Gulf of Cadiz; and northwest Africa. The Iberian section of the upwelling system is effectively isolated from the much larger African section by the Gulf of Cadiz, across which there is apparently little direct flow (Barton, 1998;

Barton, 2001). However, it is well established that Mediterranean Water (MW) flows to the north along the Iberian continental slope as part of an undercurrent that probably extends from levels of 400-500 m layer to 1200-1300 m (Ambar and Howe, 1979; Ambar, 1983; Ambar *et al*., 1986; Iorga and Lozier, 1999; Bower *et al*., 2002). The main surface currents are the Portugal Current (PoC), the Azores Current (AC) and the Canary Current (CaC). The AC is the southern branch of the Gulf Stream, which splits into two south of the Azores: one part moves toward the central Canary basin in the southeast to feed the CaC, the other heads eastward for the GoC (e.g. Pingree, 1997; Johnson and Stevens, 2000; Barton, 2001). Barton (2001) notes that the position of the AC (the shoreward limit of the CaC) oscillates south (east) in summer and north (west) in winter.

3.3.1 Western Iberia

The western Iberian offshore region falls at the northern extreme of the CUE, at 37-43°N (Figure 4). The main upwelling season (northerly winds) occurs during the summer months between July and September (Wooster *et al*, 1976). In winter the winds relax, with intermittent periods of both upwelling- and downwelling-favourable winds. The thickness of the ocean surface layer (i.e. the upper ocean mixed layer and seasonal thermocline) varies widely according to the season: winter mixing reaches 200 m at ~40°N (Arhan *et al*., 1994). Water in the upper layers (<300 m) corresponds to Eastern North Atlantic Central Water of sub-tropical origin ($ENACW_T$), with T>12.5 °C and S>35.70 (see 1.4 above). The broad PoC system extends zonally from the Iberian coast to about 24°W, 300 km beyond the shelf. Observed at yearly timescales, the PoC defines the classic, slow, southerly flow regime of an eastern boundary current. At shorter timescales, the PoC reveals intricate interactions between coastal and offshore currents, bottom topography and water masses. There is also considerable seasonal variability: the mean flow of the uppermost water column (<600 m) fluctuates according to season and varies more with increasing proximity to the coast; northwards in autumn and winter, west and southwards in spring and summer (Huthnance *et al*., 2002). However, significant on- and off-shelf transport of about 2 Sv can be measured throughout the year (Pérez *et al*., 2001; Martins *et al*., 2002). Below the surface, a poleward flowing undercurrent is consistently present over the slope (see also 1.3, 2.5, 3.1 and 3.2; Huthnance *et al*., 2002); this relatively narrow and weak flow (e.g. Haynes and Barton, 1990; Frouin *et al*., 1990; Mazé *et al*.,

1997) that often extends to the surface during winter, has been referred to both as the Iberian Poleward Current (IPC) (Peliz *et al*., 2003a) and the Portugal Coastal Counter Current (PCCC) (Ambar and Fiúza, 1994). Figure 4 shows that north of about 41°N, the IPC is confined to the slope by the southward-flowing PoC, and its recurrent eastern branch.

Off western Iberia, there is considerable water mass variability found along sections with small along-slope variations: Huthnance (1995) suggests that the circulation here may consist of a number of distinct (horizontal) cells, each with poleward flow at the eastern boundary margin, but with limited continuity between them; in this way, individual water parcels (plus eggs/larvae) may be mixed offshore or otherwise recirculated. Mesoscale eddies are common features of the western Iberian region, owing to both the presence of the IPC (see 2.8 above) and the poleward flow of MW (see 3.3.2 below and 2.8 above) interacting with the topography (e.g. Peliz *et al*., 2002; Peliz *et al*., 2003b; Míguez *et al*., 2005; Serra and Ambar, 2002). Peliz *et al*. (2003b) use a numerical model to describe the generation and unstable evolution of eddies in the vicinity of prominent bathymetric features [the Aveiro Canyon (40.5°N) and the Estremadura Promontory (41.5°N)]. The results showed that anticyclones associated with the shedding of cyclonic eddies sometimes remained trapped to the slope, in the lees of the topographic features, for periods of up to 2-3 months. The presence of dipole eddy pairs off Cape St. Vincent has been reported by Peliz *et al*. (2004).

The review article of the oceanography of the Galician Rías Baixas (42-43°N) by Varela *et al*. (2005) mentions recent observations of internal waves: tidal interaction with the shelf-break topography generates internal waves with (relatively high) frequencies of 15-50 minutes and vertical displacements up to 30 m. Associated vertical velocities may reach 3 m min^{-1}, sufficient to modify the thermohaline and biogeochemical structure of the water column. The phenomenon occurs all-year round, with high variability in its frequency characteristics. Internal tides in the western Iberian region have also been reported by Fraga (1996); Barton *et al*. (2001); Small (2002) at Cape Sines, 37.9°N; and da Silva *et al*. (1998) off Porto, 41.0°N (see also 3.1 above).

3.3.1.1 Spring/summer (March/April to September/October)

During spring and summer the predominant, northeasterly trade winds cause persistent upwelling of cooler water from about 100-300 m depth, along the entire western Iberian coastal margin (Fiúza, 1983; Haynes *et al*., 1993; Smyth *et al*., 2001). An associated 30-40-km wide equatorward current develops at the surface (<100 m) that dominates over the continental shelf and slope; in the north this current transports recently upwelled, cold and nutrient-rich $ENACW_P$, while further south colder and relatively nutrient-poor $ENACW_T$ is transported. Upwelling events usually begin, and remain particularly intense, off Cape Finisterre (42.9°N), Cape St. Vincent (37.0°N) and Cabo da Roca (38.8°N). There is evidence for the presence of an upwelling shadow zone in the lee of Cabo da Roca (Moita *et al*., 2003). Upwelling filaments (see 2.4.1 above) often form at these and other capes, extending more than 100 km offshore (Haynes *et al*., 1993); thermal features in satellite images can be tracked to obtain their velocities, which can reach 0.28 m s^{-1} (Smyth *et al*., 2001). The upwelling centre that develops at Cape Finisterre (and extends ~60 km northeast beyond Punta Roncudo) has some particular characteristics: the area may be divided into two sub-regimes according to the wind patterns and also the character of the upwelled water masses. Upwelling to the north of Punta Roncudo is favoured by northeasterly winds, which are more variable and less intense than the northerly winds which favour upwelling further south over the Rías Baixas; off Cape Finisterre is the region of the Galicia Front (see 1.4 above) where $ENACW_T$ meets $ENACW_P$, such that the water mass character of the upwelled water displays considerable temporal and spatial variability (e.g. Ríos *et al*., 1992; Míguez *et al*., 2005).

Small (2002) describes summer observations of large amplitude (up to 50 m in depths of 250 m) internal waves off Cape Sines. SAR images showed that the position of the waves, which approached the coast at between 0.35 and 0.45 m s^{-1} separated by wavelengths of 16-20 km, was linked to the tide. Three sources of generation were identified for the waves: the ~500-km, curved Horseshoe Seamount chain (visible in Figure 4 to the southwest of Cape St. Vincent), which in places rises up to within a few hundred meters of the sea surface (e.g. Richardson *et al*., 2000); the slope region to the north of Cape Sines; and as the product of resonant interactions between the waves generated by the two previous mechanisms.

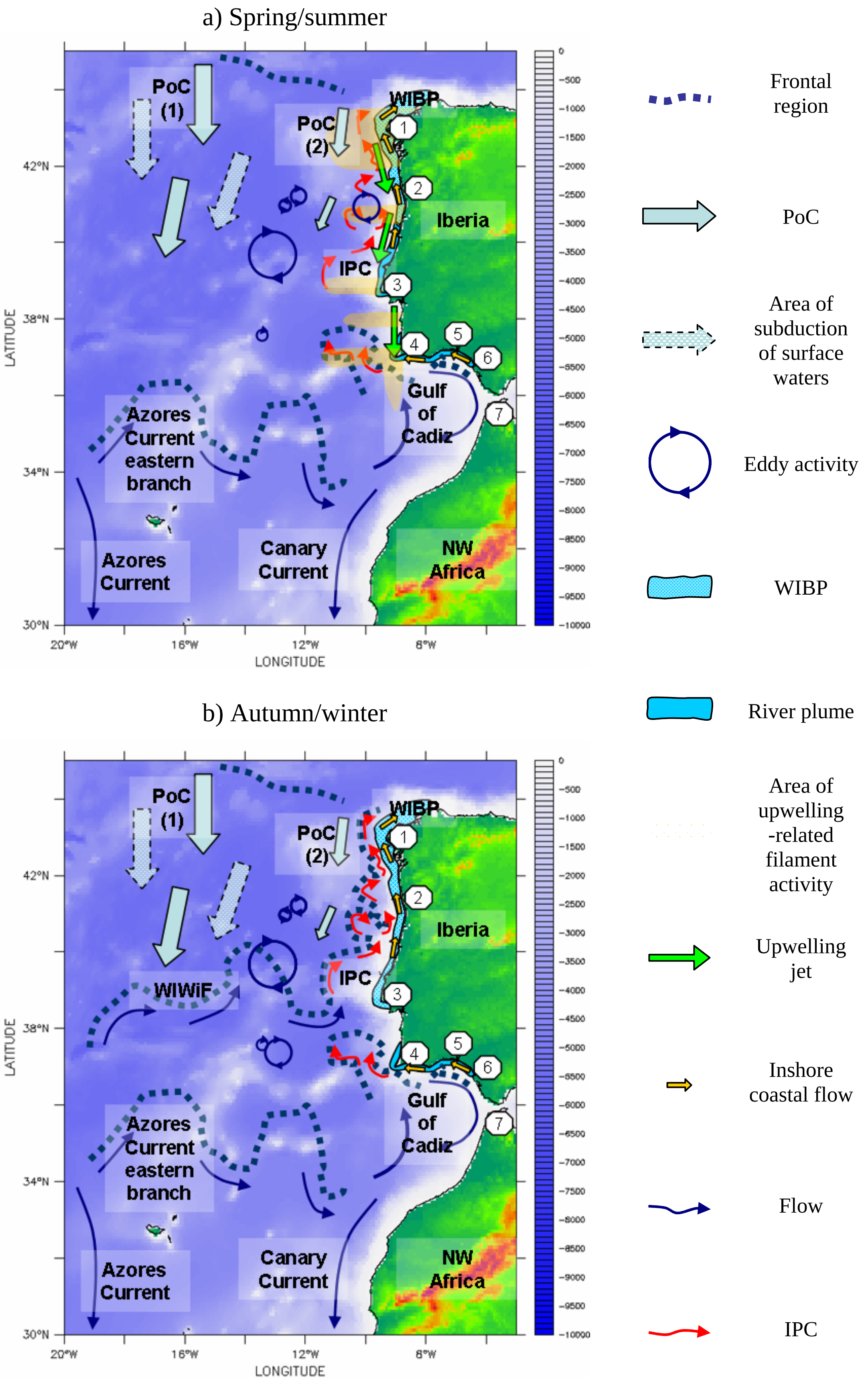

Figure 4. The western Iberia and Gulf of Cadiz regimes in a) spring and summer, and b) autumn and winter. 1) Cape Finisterre; 2) River Douro; 3) Cabo da Roca; 4) Cape St. Vincent; 5) Guadiana River; 6) Guadalquivir River; 7) Strait of Gibraltar. Adapted from Peliz *et al.* (2002; 2005).

There is considerable zonal variability in the upwelling-related flow field off western Iberia, as illustrated in Figure 4a, which is a schematic diagram adapted from Figure 8 of Peliz *et al*. (2002), who reported the results of a September 1998 survey off the northern Portuguese coast at 41°N. Two important counter-flows are visible in Figure 4a: the first is the slope poleward flow, the IPC; the second is a coastal counter-flow that is attached to the coast. Peliz *et al*. (2002) speculate that this second feature is partly attributable to buoyancy input from the many regional rivers (the Douro, Minho and Mondego Rivers, other smaller rivers, and the Rías Baixas). They named this low-salinity water lens, a year-round feature which extends all along the coast as a narrow band, the Western Iberia Buoyant Plume (WIBP). The WIBP influences the structures related to the upwelling by increasing stratification over the shelf (reducing offshore Ekman transport by reducing the thickness of the Ekman layer), and by the creation of an inshore frontal region that promotes northward baroclinic transport. The review article by Varela *et al*. (2005) also refers to the presence of the WIBP further to the north in the region of the Rías Baixas. The southern section of the western Iberian coast has two important rivers, the Tagus and the Sado; these rivers are both located to the northern part of the southern section, and so the associated dynamical effects may not be so prominent further south. Figure 4a indicates that the summer western Iberia upwelling circulation is complex: it is a double-frontal system, subject to alongshore and cross-shore accelerations, influenced by the wind, by the topography, and by the stratification input from the WIBP.

3.3.1.2 Autumn/winter (October/November to March)

Rather little is known about the wintertime circulation off western Iberia (Figure 4b). However, a recent review by Peliz *et al*. (2005) has clarified the various results and conclusions from previous studies. At about 39-40°N, winter SST imagery reveals a recurrent meandering frontal system, named the Western Iberia Winter Front (WIWiF), which represents the transition to the southern area of the Iberian Basin, where the PoC is less influential. The WIWiF supports the eastward advection of relatively warm and salty water, and this front is considered to be the main generating mechanism of the poleward flowing IPC (Peliz *et al*., 2005): near to the coast the front is deflected northward, and so becomes the IPC. Further south, 35-36°N, the easternmost extension of the AC is responsible for a second frontal system. A warm southerly flow from the Gulf of Cadiz is a further recurrent feature found in SST imagery; this is likely to be a northward recirculation of the AC, possibly connected with the IPC. In contrast to the predominant summer upwelling, the wind regime is variable in autumn and winter with the occurrence of both upwelling- and downwelling-favourable wind patterns (e.g. Santos *et al.*, 2004; Mason *et al.*, 2005). Santos *et al*. (2004) demonstrated a retention mechanism for sardine larvae produced by the interaction of a strong winter upwelling event, the IPC and the buoyant (WIBP) river plume off western Iberia in February 2000. The WIBP (see 3.3.1.1 above) is a particularly important feature owing to the maximum regional rainfalls that characterise the winter months. In some cases, the WIBP may be associated with strong poleward transport over the shelf.

3.3.2 The Gulf of Cadiz

The Gulf of Cadiz (GoC), where the trend of the coastline is at a large angle with respect to the dominant trade wind direction, effectively separates the extensive CUE (Iberia and northwest Africa) into two parts, between which continuity of flow is thought to be largely absent (Figure 4; Barton, 1998). The surface circulation of the GoC remains relatively understudied, particularly for the wintertime, in comparison with aspects of its deeper circulation, and the dynamics of the wider Iberian region. The GoC is enclosed to the north and east by the IP and the Atlantic coast of northwest Africa. Open boundaries to the south and west ensure that the GoC is a conduit for the exchange of water masses between the Mediterranean Sea and the North Atlantic Ocean; indeed, this exchange dominates the circulation of the GoC. Intensive observational and modelling research has been directed towards understanding of the outflow and subsequent spreading of high-salinity (>37.00) MW from the 300-m deep Strait of Gibraltar, which is located at the eastern extreme of the GoC (e.g. Price *et al.*, 1993; Price and Baringer, 1994; Ambar *et al.*, 2002). The fate of this deep-flowing MW is mentioned in sections 1.3 and 2.8 above, and 3.3.3 below; meanwhile, in the upper waters of the Strait of Gibraltar, there is a corresponding inflow to the Mediterranean Sea of $ENACW_T$. Consequently, the GoC is often characterised as having a generally eastward surface circulation (e.g. Vargas *et al.*, 2003; Sánchez and Relvas, 2003). It has been suggested that this gentle eastward flow represents the last meander of the AC (Johnson and Stevens, 2000). A modelling study by Jia (2000) indicates that entrainment processes associated with the Mediterranean outflow in the GoC (a significant volume of overlying $ENACW_T$ is entrained into the sinking MW) may actually be a generating mechanism in the formation of the AC.

Recent studies to focus on the surface circulation within the GoC include Folkard *et al*. (1997), Relvas and Barton (2002), Vargas *et al*. (2003), Sánchez and Relvas (2003) and Sánchez *et al*. (2005). Sánchez and Relvas (2003) use climatological data to strengthen the case for the anticyclonic circulation being fed from the west by the AC, and also suggest there to be a significant seasonal contribution from the western Iberian upwelling equatorward current in summer. Folkard *et al*. (1997) conclude that the surface circulation in the GoC is primarily wind-forced; although, onshore of the shelf edge, the surface responses to the wind forcing may be complicated by the varying trend of the GoC coastline. The wind field over the GoC displays an east-west tendency (in comparison with the north-south tendency off western Iberia) that results from the constraining effect of the local orography (Folkard *et al*., 1997). In general, upwelling and downwelling responses tend to be weaker, less uniform and more localised than they are over western Iberia. Using *in situ* data and infrared imagery, Stevenson (1977) described a warm coastal current flowing along the Spanish and Portuguese coast, counter to the mean anticyclonic circulation (i.e. poleward), and named the frontal region separating these two flows the Huelva Front (Figure 4). The Huelva Front is an extensive feature, sometimes visible over the length of the shelf between the two major GoC rivers, the Guadalquivir and the Guadiana. Buoyant river plumes from these and other smaller rivers, for instance the Tinto-Odiel River System at Huelva, are a persistent inshore feature of the GoC; the Guadalquivir contributes a significant outflow in spring as a result of melting snow in the Sierra Morena (Peliz and Fiúza, 1999). Stevenson (1977) also identified a region of high current shear to the southeast of Cape St. Maria (westward of the Guadiana River mouth), which he named the Stafford Shear. The Stafford Shear is associated with an upwelling-related cold-warm frontal region. Rapid current reversals are known to occur over the shelf region, as Sánchez *et al*. (2005) observed over the Portuguese shelf in the northern GoC. This event, which took place in less than 48 hours, was not correlated with the local wind stress; instead, the authors showed large-scale remote atmospheric forcing to be the likely responsible mechanism.

The exchange of $ENACW_T$ and MW that takes place at the Strait of Gibraltar is essentially a 2-way exchange, which is density-driven: warm, high-salinity MW flows in to the GoC at depth, whilst cooler, fresher, less dense Atlantic water flows out at the surface (e.g. Bryden and Stommel, 1984). This scheme suggests one-way transport only (west to east) for surface water-borne particles. However, a recent study by Stanichny *et al*. (2005) demonstrates that certain wind patterns can induce net westward surface transport in the Strait of Gibraltar. Easterly winds promote upwelling off the African coast, which gives rise to an associated westward surface current in the southern portion of the Strait [see Figure 3 of Stanichny *et al*. (2005) for a sequence of satellite SST images over the duration of such an event]. The resulting upwelling plume may be advected ~100 km into the GoC, where it interacts with the mesoscale circulation. These observations demonstrate a mechanism whereby mixing between coastal surface waters in the Gulf of Cadiz, both Iberian and African, and the interior of the GoC, may occur. However, this mixing appears to be at a rather limited scale, driven by random transient events, and so the claim by Barton (1998) for negligible continuity of meridional flow through the GoC must stand.

The GoC represents a major inter-regional boundary between the seasonal upwelling of western Iberia and the continuous Canary upwelling to the south, and this merits a short discussion within the context of a recent study by Gaylord and Gaines (2000). These authors, noting that there is a commonly held assumption that variations in properties (usually temperature) at oceanographic discontinuities are responsible for generating biological distributional patterns, looked at the potential for ocean flows to perform this function. Their results indicate that the simple, common flow fields often observed in association with biogeographic boundaries may have the potential to constrain a species' geographic range, even in cases where suitable habitat outside the range is abundant. The authors used model simulations to show that one- or two-way barriers may limit range expansion, and these may be differentially permeable, with boundary "leakiness" depending on life-history characteristics of the species concerned, and the degree of temporal variability in the near-shore flow field. According to Gaylord and Gaines (2000), a region where an onshore current meets the coastal boundary provides a suitable scenario, and this indeed appears to be the case in the Gulf of Cadiz: the eastward-flowing branch of the AC meets the continental shelf in the GoC region (see 3.3 above for further details of this circulation). This may be a one-way boundary, such that transport occurs from western Iberia into the Gulf of Cadiz, but not (or rarely, see Relvas and Barton, 2002; 2005) vice-versa.

3.3.2.1 Spring/summer (March/April to September/October)

Folkard *et al*. (1997) report that spatial SST variability in the GoC is typically high in summer, though less so in spring. For the spring and summer periods, they demonstrated the contrasting effects of westerly and easterly winds on SST fields over the region: westerly winds resulted in cool surface water extending from the western IP around Cape St. Vincent, flowing eastward as far as the Guadiana River [such flows are also described by Fiúza (1983), Relvas and Barton (2002) and Vargas *et al*. (2003)]; in the case of easterly winds, Folkard *et al*. (1997) found the eastward extension of cooler water beyond Cape St. Vincent to be less pronounced, with warmer water occupying a correspondingly larger westward extent of the shelf, and GoC as a whole. Westerly winds also support local upwelling along the southern Portuguese coast, particularly downwind of Cape St. Maria (which is the location of the Stafford Shear); García *et al*. (2002) conducted a cruise in the region in July 1997 and found anomalously low SST, together with enhanced nutrient and chlorophyll concentrations. Vargas *et al*. (2003) also report that easterlies produce strong upwelling along the Cape Trafalgar coast and heating of the shelf waters north of Cadiz.

3.3.2.2 Autumn/winter (October/November to March)

In autumn and winter there are a number of contrasts to the summer situations observed by Folkard *et al*. (1997). Notably, the entire Iberian GoC coastline area is occupied by cooler waters, with warmer waters confined to the interior. SST variability tends to be very low in winter, but high in autumn; this contrast is attributed to the increased surface mixing that occurs with the onset of winter. Álvarez *et al*. (1999) used statistical analyses of a 1-year time series of surface currents (measured by the Cadiz buoy at 36.47°N, 6.95°W) to show that, between December-February, the flow was opposite to the otherwise anticyclonic circulation; these results indicate that a (mean) winter cyclonic circulation regime may occur, although further investigation of a longer time series is clearly necessary.

3.3.3 Northwest Africa

The northwest African section of the CUE is subject to year-round upwelling, apart from at its northern (above ~30°N) and southern (below ~20°N) limits where the phenomenon is seasonal (January through May; Wooster *et al.*, 1976; Nykjaer and Van Camp, 1994; Figure 5). It is an extensive region, with nearly 2500 km (Dakar to Gibraltar) of coastline featuring several large bays and capes which introduce significant local structure along its length, usually an enhancement of the upwelling (e.g. Van Camp *et al.*, 1991; Nykjaer and Van Camp, 1994; Santos *et al.*, 2005). These modifications to the circulation often have a seasonal element. To the north, the region is bounded by the Strait of Gibraltar at 36°N, where MW empties into the GoC. A major topographic feature is the Canary Island archipelago, which lies at the transition zone between coastal and oceanic waters. This location results in localised hydrodynamic effects, such as filament formation, through the interaction of the upwelling and the alongshore southward-flowing Canary Current (CaC). The CaC, fed by a branch of the Azores Current (AC; see also Figure 1) which passes north of and around Madeira at 33°N, transports around 5 Sv in two branches separated by the island of Madeira (Stramma, 2001; Barton, 2001). This figure may be compared with an estimated 6- to 9-Sv off-shelf flow through filaments, which are located roughly every 200 km along the coast (Barton *et al*., 2004) and are often associated with capes or headlands, although most of this flow recirculates shoreward. The CaC, which extends to around 500 m depth and is up to 1000 km wide, is generally cool through the entrainment of upwelled water. The CaC separates from the African coast in the region of Cape Blanco at 21°N, and by 15°N (Cape Verde) all of the flow has turned westward to supply the North Equatorial Current. It should be noted that the described current structure concerns average values, and that at short timescales considerable variability is evident.

The dominant surface water mass (<600 m) is nutrient-poor $ENACW_T$. In the south of the region, beyond Cape Blanco, the influence of fresher South Atlantic Central Water (SACW), which is also richer in nutrients, is evident. SACW is transported into the region between at ~7-9°N within the eastward-flowing North Equatorial Counter Current (NECC). A major frontal region between the two water masses, $ENACW_T$ and SACW, occurs at the confluence between the NECC flow and the

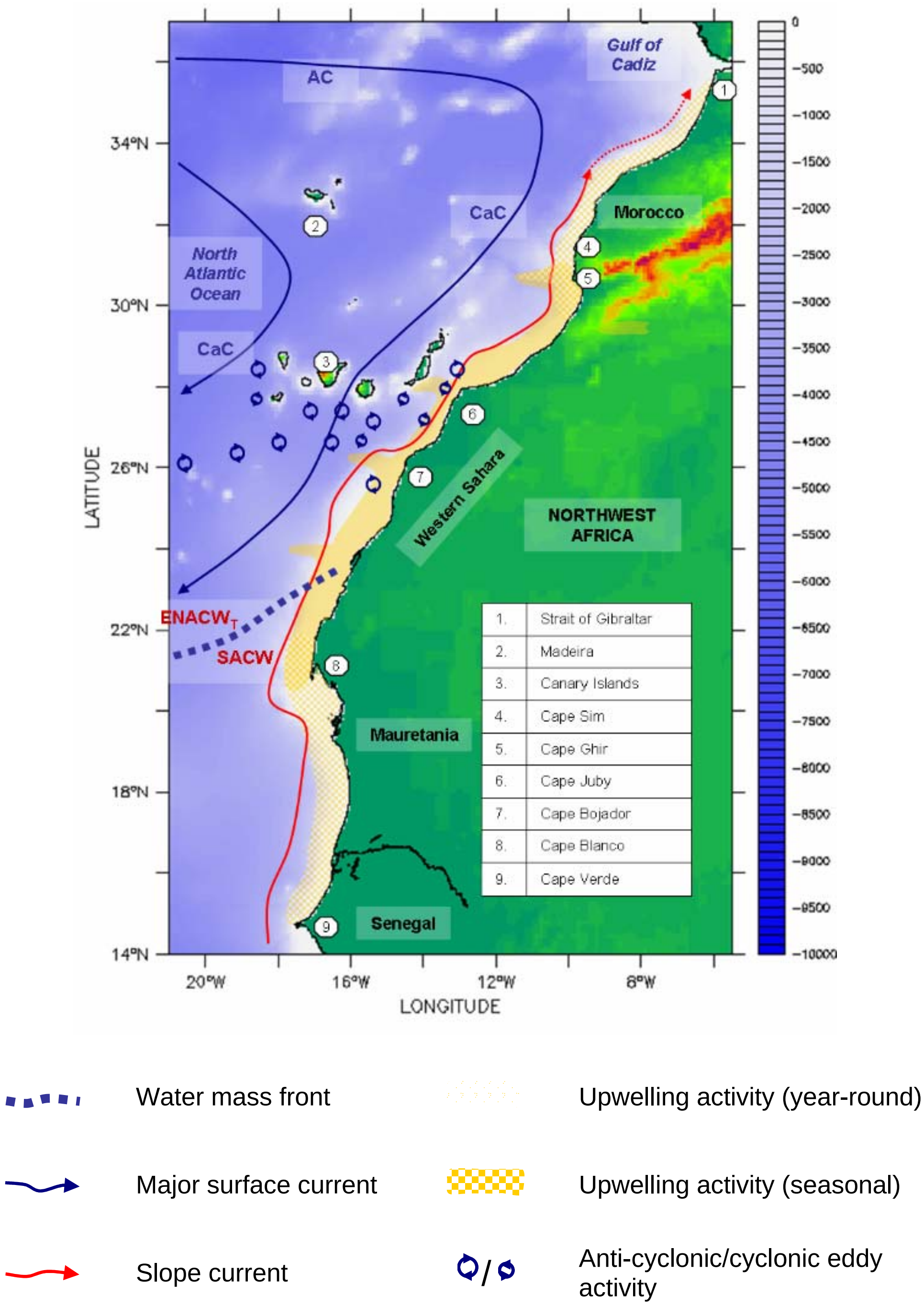

Figure 5. The northwest African upwelling region. Upwelling continues all year down to about 20°N; further south it is seasonal (January-May). Note the filament activity associated with the major capes, and the extensive and significant mesoscale eddy activity. AC-Azores Current, CaC-Canary Current, ENACW$_T$-Eastern North Atlantic Central Water of sub-tropical origin, SACW-South Atlantic Central Water.

southward-flowing CaC. Intense mixing takes place at this boundary, which is convoluted and variable in position. Upwelling to the south of Cape Blanco is more productive than that further north as a result of the higher SACW nutrient content.

The northwest African shelf is typically 50 km wide (varying between 20-150 km; wider than Iberia, narrower than Biscay) and of shallow depth (~80 m), and the water column is weakly stratified. Santos *et al*. (2005) demonstrated the intensity of the cyclonic wind-stress curl (which produces upwelling) at two known centres of upwelling, both located adjacent to capes: Cape Sim at 31-32°N and Cape Bojador at 24-25°N (Parrish *et al*., 1983; Bakun and Nelson, 1991). A persistent cyclonic eddy is located off central Morocco (Head *et al*., 1996), which is associated with the local bathymetry (Mittelstaedt, 1991). The influence of river runoff is generally low over the northwestern African shelf, apart from in the far south (~18°N) where seasonal (winter) equatorial discharges are high.

A subsurface poleward current flows along the northwest African slope (Barton, 1989; see also 2.5 above). The undercurrent is generally less than 100 km wide with a vertical extent of ~700 m, and is composed of nutrient-rich SACW. Maximum speeds in the core of the flow, at about 300 m depth, are ~10 cm s^{-1}. The current is evident in hydrographic sections throughout the upwelling region, apart from the area south of 20°N where the regime is different. Off northern Morocco there are few observations of the undercurrent, and an unresolved question concerns whether this flow is continuous with the IPC to the north of the GoC (e.g. Barton, 2001). At ~1200 m depth, along the extent of the northwest African slope, there is a southward-spreading layer of MW (e.g. Knoll *et al*., 2002). A continuous northwest Africa-Iberia poleward flow would imply considerable interaction with the outflowing MW in the eastern GoC, but at present observations of such mixing are missing. The relatively small amplitude of tidal motions off the coast of northwest Africa indicates that tides are unlikely to play a significant role in the variability over the shelf (Barton, 1998).

3.3.3.1 The Canary Island archipelago

The Canary Islands are a group of 7 separate volcanic islands, which lie between 90 and 500 km from the African coast. Deep water (>1000 m) separates the islands from the continental landmass. The islands' abrupt topography, unique amongst the sub-tropical eastern boundary currents, creates a downstream wake in the flow of the CaC, and also in the trade winds (the highest mountain peak is 3718 m). Wind speeds, strongest in summer, are typically 5-10 m s^{-1}, although channelling between the islands can double these values. Mesoscale turbulence is significant as eddies, both cyclonic and anticyclonic, are shed from the islands, sometimes drawing filaments of colder upwelled water from the African shelf region several hundreds of kilometres offshore. The eddies are most likely to arise from disturbances of the current flow, but wind perturbation may also have an influence (e.g. Arístegui *et al*., 1994; Barton *et al*.,1998; Barton *et al*., 2000). Pingree and Garcia-Soto (2004) describe seasonal eddy generation that takes place to the south of the Canary Islands: positive sea-level anomalies (anticyclonic eddy structures) are generally formed toward the end of the summer (August), these structures then travel zonally westwards along 26°N at a speed of about 1° of longitude per month, reaching as far as ~50°W after nearly 3 years. The eddies are associated with increased chlorophyll-a concentrations.

Barton *et al*. (2004) provides the physical background to a series of biological and biogeochemical surveys that were carried out in the region of Cape Juby (28°N, Morocco) and the Canary Islands in August 1999 (Barton and Arístegui, 2004). The aim of the surveys was to investigate the offshore flow of upwelled waters through a recurrent filament system found between Cape Juby and Cape Bojador at about 27°N (see also Navarro-Pérez and Barton, 1998). The recurrent filament is associated with a quasi-permanent cyclonic eddy located over a bathymetric trough south of the island of Fuerteventura. The circulation within and around the filament system was revealed to be highly complex, with location of origin and the presence of eddies being the primary determiners between retention (of water parcels) within the upwelling zone or dispersion into the ocean. It was shown that there were multiple possible paths for retention or export offshore. Dual filament structures were also reported, providing further indication of the complexity of the circulation. In these cases, the water within the individual filaments was shown to have distinct temperature and salinity properties, suggesting different origins. Barton *et al*. (2004) obtained transport estimates of up to 1.3 Sv for the filaments (over a fifth of the total CaC transport estimate). See Rodríguez *et al*. (2004) for a discussion of the effects of the mesoscale circulation in the vicinity of the Canary Islands on fish larval distributions.

3.3.3.2 Seasonal descriptions

In comparison with the regions previously described in section 3, the northwest Africa regions displays rather weak seasonal variability. The main features are the winter peak in upwelling that occurs south of ~20°N, the southward meridional extent of the upwelling which is regulated by the seasonally-varying position of the Azores High, and slight changes in the position of the near-shore branch of the CaC. Small-scale upwelling events may take place at any location along the coast at any time of the year.

3.3.3.2.1 Spring/summer (March-August)

During the summer months the trade wind belt displays a strong alongshore component along the northwest African coast between 35-20°N, which drives upwelling (e.g. Wooster *et al*., 1976; Fiúza *et al*.., 1982), for which a corresponding surface water temperature anomaly (i.e. colder water) has been demonstrated by Wooster *et al*. (1976). The Moroccan coast (35-30°N) typically experiences weaker upwelling than further south, due to the orientation of the coastline; coastal winds here tend to blow onshore and are relatively weak. Between 30 and 20°N winds blow parallel to the coast. The winds are strongest in summer, with peak intensities in July-August. A six-week springtime survey just south of 22°N, undertaken in 1974, revealed a series of intense upwelling-favourable wind events of 7-10 days duration, separated by 1-2 days of weak or poleward wind (Halpern, 1977). The ocean response, which lagged the wind forcing by about 1 day, included strong surface cooling and a raising of isopycnals within 50 km of the coast, and the establishment of an equatorward surface jet with velocities exceeding 40 cm s^{-1}; when the winds weakened rapid warming was observed at the surface. The poleward slope undercurrent, which here is found at about 250 m depth, persisted at about 10 cm s^{-1} throughout the study period. South of Cape Blanco at 21°N, the region is characterised by weak monsoon winds, often with a southerly component; a warm mostly poleward flow is observed at the surface (Mittelstaedt, 1991; Van Camp *et al*., 1991). Indeed, coastal SST here is characteristically warmer than in the offshore oceanic water in summer; although Santos *et al*. (2005) found this not to be the case in 1982-1990, they speculated decadal variability of the monsoon winds to be the cause.

3.3.3.2.2 Autumn/winter (September-February)

In winter the coastal upwelling reaches its maximum intensity in the southern region of northwest Africa, extending beyond Cape Blanco (21°N) to about 12°N as a result of the southward shift of the Azores High. Drifter experiments by Barton *et al*. (2004) in the Canaries-Africa region between August-December 1999 showed a northward and offshore advection of a single drifter between October-December. This observation provides further evidence for the suggestion that a major autumn flow reversal occurs in this region (e.g. Stramma and Siedler, 1988; Navarro-Pérez and Barton, 2001; Knoll *et al*., 2002; Hernández-Guerra *et al*., 2003). The near-shore branch of the CaC migrates away from the African coast in winter (Barton, 2001).

4 CONCLUDING REMARKS

The eastern North Atlantic boundary is a highly complex region in comparison with other major eastern boundaries such as the U.S. West Coast, the Peru-Chile upwelling system or the Benguela upwelling system. The primary difference between the present region and the two examples mentioned above is the considerably greater irregularity of the European and northwest African coastline and topography. It has been shown that whilst the coastline orientation is largely meridionally orientated, there are significant zonal and otherwise anomalous stretches of coastline orientation, particularly in the Bay of Biscay and the Gulf of Cadiz. Two further unique topographic features are the Strait of Gibraltar, where dense Mediterranean Water enters and passes through the Gulf of Cadiz from the Mediterranean Sea, and the Canary Island archipelago, which disturbs the prevailing oceanic (and atmospheric) flows producing significant downstream variability on the scale of the islands. These features are, individually and in sum, major contributors to the complex and variable circulation system that has been described in the preceding sections of this review, onto which is superimposed multi-scale seasonal and inter-annual variations in atmospheric forcing, heating, and input of buoyancy through river discharges.

Amongst the implications of this complexity for the fisheries, biological and chemical oceanographic communities involved in investigating the eastern north Atlantic region, is the necessity for particularly close working relationships with physical oceanographers. Generalisations of eastern boundary circulation schemes seem less reliable when applied to the northeastern Atlantic region than they may be for the other eastern boundary regimes, so that closer considerations of local processes are necessary. Examples of such approaches are many in this report, such as the work by Santos *et al*. (2004) described in section 3.3.1.2; here an upwelling event in winter was hypothesized to have a retentive effect on sardine larvae distributions offshore of Aveiro, through interactions with the poleward current (the IPC) and a buoyant river plume (the WIBP). The 'generalised' view of upwelling is that it occurs in the summertime, and it is dispersive (in the offshore direction) rather than retentive; there is a clear conflict between this view and the results of studies such as Santos *et al.* (2004). The problem of defining to what extent generalisations may be confidently applied can only be addressed once better knowledge of local processes is obtained, and this is no trivial task.

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