Coupling between the open ocean and the coastal upwelling region off northwest Africa: water recirculation and offshore pumping of organic matter


Facultad de Ciencias del Mar, Universidad de Las Palmas de Gran Canaria, 35017 Las Palmas de Gran Canaria, Spain

Received 24 May 2002; accepted 1 July 2004
Available online 29 September 2004

Abstract

The surface and upper-thermocline waters of the Canary Basin are characterised by very strong coupling between the open ocean and the coastal upwelling region. Such coupling has its origin in water inflow into the upwelling region north of the Canary Islands and its recirculation south along the continental slope, which is the true Canary Current. A portion of this recirculating water is intermittently exported offshore through surface filaments. During late fall, a major diversion takes place at Cape Ghir, allowing the presence of northward flow from Cape Blanc till Cape Yubi. The fraction of water that flows through the Canary Archipelago is the origin of intense mesoscale variability south of the Canary Archipelago, which interacts strongly with the coastal region. These physical characteristics are responsible of intense alongshore and vertical fluxes of nutrients and dissolved inorganic carbon within the upwelling region. Coastal filaments and cyclonic eddies cause localised offshore export of nutrients and organic matter, making possible that respiration be several times larger than production in the open ocean. A major characteristic of the ecosystem comes from the seasonal variation in the current pattern, allowing coastal convergence and intense transfer of coastal properties to the open ocean during late fall.

© 2004 Elsevier B.V. All rights reserved.

Keywords: Coastal upwelling; Canary Current; Water recirculation; Organic matter

1. Introduction

Nutrient supply, primary production, organic matter export and nutrient mineralization constitute the basic sequence for completion of the biogeochemical cycle. The spatial variability of this cycle depends on the physical mechanisms that operate at local and...
regional scales. In some areas the physical environment is capable of supporting large nutrient input that leads to high productivity. The best example of such areas is the upwelling regions in the eastern boundary of all subtropical gyres where the wind regime accelerates the nutrients influx, allowing high rates of primary production, and carbon recycling and export (e.g. Huntsman and Barber, 1977; Codispoti et al., 1982; Schulz, 1982; Jewell, 1994). In upwelling regions, however, additional mechanisms may favour the persistence of biogeochemical transformations, even with weak or no upwelling favourable winds. Localised near-horizontal water exchange between the interior and coastal oceans for example, may bring large amounts of nutrients into the coastal region and provide paths for carbon export. And the specific planktonic community structures may be adapted to the changing physical environment and intermittent nutrient input in the upwelling regime.

Although early ideas of the recirculation of subtropical gyres proposed a weak diffuse eastern flow, more recent circulation maps of the North Atlantic Ocean illustrate an eastward flowing Azores Current that splits into several rather energetic southward branches (Schmitz and McCartney, 1993; Reid, 1994). The number, intensity and location of most of these branches seem to be highly variable, their origin likely linked to developing instabilities of the Azores Current. The easternmost branch, however, always flows east into the coastal upwelling region off northwest Africa and is what we identify as the Canary Current. This branch is possibly the best available example of large-scale interaction between the coastal upwelling and open ocean regions.

Several regional studies, or studies of the Atlantic circulation with enough data resolution in the eastern boundary, have indeed improved our view of the Canary Current (Stramma and Siedler, 1988; Lozier et al., 1995; Siedler and Onken, 1996). These studies have confirmed the existence of water inflow from the open ocean into the coastal upwelling region north of the Canary Islands. This may be interpreted as the coastal upwelling region being the true eastern boundary condition for the subtropical gyre, and the subtropical gyre providing the continuous large-scale forcing on the coastal ocean. How the system is coupled however, is not easily distinguishable in these early studies because they lacked enough spatial resolution in the coastal upwelling region. The importance of this coupling, emphasised by several authors (Pelegrí et al., 1997; Barton, 1998; Barton et al., 1998), lead to its study in a rather large European Union project (CANIGO). In CANIGO several works dealt with the coastal transition zone north of the Canary Archipelago, where the water recirculates south along the continental slope, as well as with specific coastal sites where quasi-permanent filaments stretch offshore (Hernández-Guerra et al., 2001, 2002; Laiz et al., 2001; Pelegrí et al., 2005).

Our objective here is to examine the large-scale and mesoscale processes that develop in the open ocean and coastal upwelling coupled system, and to scrutinise which mechanisms support the biogeochemical processes in the surface and upper thermocline layers of this region. Specifically, we want to examine what the physical (wind forcing, solar radiation, influx from the open ocean, recirculation through the upwelling region, stirring role of the Canary Archipelago) and biological (nutrients availability, plankton community structure, productivity) mechanisms are that control the marine carbon system in the upper thermocline waters off northwest Africa (particulate and dissolved organic carbon, partial pressure of CO₂, total alkalinity, pH, total inorganic carbon). These processes act at different scales, from nutrient influx at the basin scale (Canary Current) to increased productivity and carbon export in mesoscale structures (filaments stretching out of the coastal upwelling region and vortices formed at the lee side of the Canary archipelago). One important point is that local processes are the response to both local and global forcing, the character of the biogeochemical cycle eventually relying on the energy found within the ocean interior and the rapid exchange of properties in the boundary region.

While there have been a number of recent reviews dealing with several specific characteristics in the eastern boundary of the subtropical gyre, our approach is clearly interdisciplinary. We are interested in interconnecting those processes that develop in the surface and upper-thermocline layers, responsible for sustaining the biogeochemical cycle. For this reason, we have not attempted to be exhaustive on each specific component of this complex system and, at each section, the reader will be directed to the relevant references. The geographic region to be examined,
both in the horizontal and vertical directions, has also been rather constrained as compared with other work in the literature. Our region of interest is the surface and upper thermocline waters of the Canary Basin, including their possible association with the poleward undercurrent and the Antarctic Intermediate Water at the eastern boundary. This spans a vertical extension from the sea surface down to approximately 750 m, and a horizontal area from the African Coast to about 20°W and from the Strait of Gibraltar to the lee side of the Canary Islands. These meridional limits include the Canary Current, cruising into and along the African continental slope, and the complex pattern of mesoscale structures that arise south of the Canary Islands because of the interaction of the Canary Current with this archipelago.

2. Air–sea interaction

2.1. Atmospheric forcing

The zonal available potential energy associated with latitudinal temperature gradients is the main energy source in middle latitude systems. In the tropics, on the other hand, the storage of available potential energy is comparably small due to the much weaker temperature gradients, making ocean–atmosphere heat exchange the primary energy source for atmospheric disturbances, especially for those that originate within or near the equatorial zone (e.g. Palmén and Newton, 1969). Here we briefly review the annual cycle of air–sea heat flux in the Canary Archipelago region as obtained from the climate atlas by Isemer and Hase (1987). Fig. 1 shows the annual cycle of the net radiation flux (Net), the oceanic heat loss (Hl) and the net air–sea heat flux (Hf) in W m$^{-2}$. The net radiation flux is the difference between incoming SW and outgoing IR; the oceanic heat loss is calculated from the sum of IR, LE and H; and the net air–sea heat flux is calculated using the data of SW, IR, LE and H. In the Canary Archipelago region, the net heat flux is positive from March to August and negative from September to February.

To calculate SW, we follow Reed's (1977) formulae, applicable for daily averages. The expression uses incident solar radiation at the surface, cloud cover, noon solar altitude and clear sky radiation to evaluate the short-wave radiation. Minima are closely related to maxima of total cloud cover and vice versa (290 W m$^{-2}$ in June, 110 W m$^{-2}$ in December). The second major component determining the net air–sea heat flux is LE. This flux is usually directed upwards, providing for main ocean loss. The patterns of LE are determined by the wind speed and mixing ratio, presenting a relative maximum of 220 W m$^{-2}$ in the trade wind zone. The most constant among all air–sea heat flux components is IR, its magnitude being mainly influenced by total cloud cover, air temperature and humidity, with the air–sea surface temperature difference playing a minor role. In the upwelling region off northwest Africa, the maximum net long-wave radiation (50 W m$^{-2}$) occurs in January, associated to a minimum of cloudiness. Sea surface temperature (SST) and surface wind speed dictate the seasonal pattern of H. This flux is relatively small during the complete annual cycle (30 W m$^{-2}$ in December, 5 W m$^{-2}$ in June), particularly if compared with its value in the western and northwestern part of the Atlantic Ocean.

Schmitt et al. (1989) compiled evaporation and precipitation data for the whole North Atlantic Ocean to calculate the haline density flux into this ocean. In the Canary Basin, mean annual evaporation exceeds precipitation by values ranging between 50 cm year$^{-1}$ in its northern portion and over 100 cm year$^{-1}$ in the southern portion. One important characteristic is that the difference remains almost constant throughout all the Canary Archipelago. Fig. 1 shows the temporal variation of these energy fluxes in the Canary Archipelago region as obtained from the climate atlas by Isemer and Hase (1987).
year, in contrast to what happens over most of the North Atlantic Ocean. Schmitt et al. (1989) further compared the haline and thermal density fluxes and obtained that the thermal density flux exceeds the haline flux by a factor between 2 and 20 in the Canary Basin, with these values increasing towards the African coast. The relatively high evaporation rate in the open ocean is related to the rather large and consistent net air–sea heat fluxes and high sea-surface temperature values. Minimum values occur in May (1.3 cm month$^{-1}$ in Gran Canaria island) and maximum values take place in December and January (1.9 cm month$^{-1}$ in Gran Canaria island). Total rainwater measured at this same location is 16.5 cm year$^{-1}$, confirming that the Canary region is a geographical area with strong evaporation.

The wind regime in the Canary Basin is mainly determined by the location of the Azores high, which causes that the mean wind direction is from the northeast all year long with changes being related to synoptic and mesoscale weather systems (INM, 1988). The most characteristic feature of the summer pattern is the ubiquitous Azores high-pressure elliptical centre that, together with the north Africa low, produces the well-defined northeasterly trade wind regime. In the Canary Archipelago the trade winds reach maximum monthly mean values of 8.7 m s$^{-1}$ in August. During autumn or spring, the Azores anticyclone is not as deep as in summer, typically presenting a quite elongated shape. The winter regime is determined by the southern movement of the Azores high, which causes the passage of low-pressure centres through the Canary Basin to be a common feature. During this season, however, mean winds are still from the northeast, the minimum monthly mean values corresponding to January (2.7 m s$^{-1}$).

Fig. 1. Seasonal variation of heat flux in the Canary Island region (after Isemer and Hase, 1987). (a) Net short-wave radiation (SW), net long-wave radiation (IR), sensible heat flux (H), and latent heat flux (LE). (b) Net radiation (Net), oceanic heat loss (Hl), and net air–sea heat flux (Hf).
Plate 1. SST satellite images for (a) August 27, 1998, (b) November 19, 1998, and (c) March 20, 1998. The colour temperature code is shown in a vertical scale at the right of the images (units in °C). The insets show the corresponding surface pressure fields (units in mbar).
Plate 1 presents SST images for 3 days, each corresponding to a different season, that show the latitudinal shift of isotherms in the surface layers as a response of heat accumulation or loss. Insets in this plate show the concurrent surface pressure fields, which illustrate the characteristic local patterns of surface pressure for summer (Plate 1a), autumn or spring (Plate 1b), and winter (Plate 1c). The pressure patterns are consistent with the ubiquitous presence of northeasterly winds from spring into fall (Plate 1a and b), with a maximum intensity during summer. These winds reinforce the Canary Current throughout these months, which attains its maximum intensity by late autumn. During winter the mean wind conditions respond to the high pressure centre located in its southernmost position, still causing northeasterlies but much weaker than during summer (Plate 1c). At this time, it is frequent to find southwesterlies, associated to the passage of low-pressure centres, which cause important forcing fluctuations in relatively short time scales, a few days as compared with a characteristic time scale of 10 days during summer.
2.2. Surface mixed layer

Below 200 m, the temperature distribution experiments only minor seasonal variations. The major changes in temperature take place in the topmost 200 m of the water column, following heat exchange between the ocean and atmosphere and the deepening of the surface mixed layer, the latter mainly resulting from increased summer winds and winter vertical convection. A good description of the seasonal evolution of the temperature in the uppermost layers of the water column has been provided by several authors (Barton et al., 1998; Ratsimandresy et al., 2001). Fig. 2, reproduced from Ratsimandresy et al. (2001), illustrates an idealised sketch of the temporal evolution of the mixed layer. The mixed layer is deep and cold in March, following winter ocean heat loss and deep convection. In June surface heating causes the surface waters are catted with a shallow warm layer, which increases its thickness throughout summer due to the action of the intense winds. After October the ocean starts losing heat and the surface temperature decreases while the surface layer thickness continues to increase.

Fig. 3, also reproduced from Ratsimandresy et al. (2001), illustrates the seasonal evolution of the temperature and temperature anomaly in the uppermost 200 m of a transect from the Canary Islands to Madeira. The seasonal evolution of the surface mixed layer is clearly visible from the temperature plots (Fig. 3a). The temperature anomaly plots, on the other hand, illustrate the relative warming and cooling of the surface layers as compared with the annual mean situation (Fig. 3b). A remarkable feature in the temperature anomaly plots is the subsurface (between 50 and 100 m) maxima easily visible during winter. Siedler et al. (1987) suggested that this water, left below the seasonal thermocline, is the origin of about 1 Sv of Madeira Mode Water that finds its way to greater depths while shifting westward. This amount of modal water, however, is still rather small as compared with the Eighteen-Degree water formed in the western subtropical gyre (Worthington, 1959).

One important aspect in the Canary Basin is that the annual mean surface winds cause water convergence at the sea surface (wind-influence) layer, which leads to negative Ekman pumping and water subduction into the ocean interior. Stommel (1979) indicated that, because of the latitudinal motion of the surface layer, surface waters only escape to the ocean interior during winter. Hence, the amount of water that does escape from the surface depends on both winter mean conditions and the passage of atmospheric frontal systems. As discussed above, despite its seasonal heat gain–loss cycle, the region has a net heat gain. The main reason for such gain is that the Canary Current is a relatively cool current, particularly cold in the upwelling region, susceptible to gain heat. The rate of heat gain in the Canary Basin has been estimated by Pelegrí et al. (1997) to be of order 0.05 PW. The escape of some of this heat from the ocean surface, either through the formation of modal water or by subduction of relatively salty water, could be an efficient mechanism for the ocean interior to absorb heat excess generated in the atmosphere (Bindoff and McDougall, 1994).

3. The open ocean

3.1. Water masses

In the frame of the CANIGO project, a closed box was carried out four times (January 97, September 97, April 98 and July 98) spanning the region between the African coast, the Canary Islands (29°N) and a parallel near Madeira (32°N). The objective was to determine the mass, heat and freshwater balances in this region. Some of these sections have been recently analyzed and the results are progressively being published (Pérez et al., 2001; Knoll et al., 2002). The southern section (the 29°N section in Fig. 4) is used here to describe the principal water masses present in the area. The vertical distribution of the mean isohalines, obtained by averaging all four cruises, is most helpful for this purpose because of the very different salinity signature of the water masses (Fig. 5).

In the surface layer, from 0 to approximately 700-m depth, we find North Atlantic Central Water (NACW). This water mass is characterised by nearly horizontal isohalines (Tomczak, 1981), as observed in Fig. 5. Several authors have differentiated NACW of subtropical and subpolar origin, the latter with salinity in this region over 35.66 (Harvey, 1982; Pollard and...
Pu, 1985; Pérez et al., 2001). From Fig. 5 we appreciate that NACW of subtropical origin occupies the top 500 m while NACW of subpolar origin occupies the depth range between 500 and 750 m. Our main interest here is the surface mixed layer and subsurface waters down to 750 m, and we will refer to these subsurface waters indistinctly as the upper-thermocline or NACW.

At intermediate layers (from approximately 600- to 1500-m depth), two water masses are present: Antarctic Intermediate Water (AAIW), detected by its low salinity (<35.3), and Mediterranean Water (MW), with high salinity values (35.5) (Hernández-Guerra et al., 2001). AAIW is clearly observed in the eastern boundary (up to 600 m) as a rather spread tongue, much wider than the classical alongshore poleward current (Barton, 1998; Pérez et al., 2001). Since Fig. 5 has been obtained using the mean of all four cruises, the existence of these minimum salinity values confirms that AAIW is a permanent feature at
the boundary. The MW is observed in the remaining of the section. Several patches of water with intermediate salinity are presumably due to mixing of AAIW and MW. Under the MW (approximately below 1500 m), we find North Atlantic Deep Water (NADW), with salinity decreasing with depth and the isohalines having a quasihorizontal distribution.

3.2. Chemical properties

Recent measurements of the marine carbon system variables have been carried out in the Canary Region in response to an increased interest in global carbon change and greenhouse warming. The capacity of the oceans to uptake carbon dioxide (CO₂) depends greatly on many factors such as hydrography, water circulation, mixed layer dynamics, wind stress and biological processes in the oceans (Broecker and Peng, 1982). A systematic study of the dissolved inorganic carbon system in the Canary Basin started in October 1995, one year after setting up the ESTOC time series station north of Gran Canaria (29°10' N, 15°30' W). At this location the partial pressure of carbon dioxide (expressed as fugacity, fCO₂) shows a seasonal pattern in the surface waters characterised by CO₂ dissolution from December to May and outgassing from June to November (González-Dávila and Santana-Casiano, 1999; Santana-Casiano et al., 2001). Considering the CO₂ gas transfer coefficient of Wanninkhof (1992) and a mean wind of 7 m s⁻¹, it turns out that on an annual scale the region acts a small sink of CO₂, at a rate of −0.15 mol m⁻² year⁻¹ (Santana-Casiano et al., 2001; González-Dávila et al., 2003). This value is found in the range estimated by Takahashi et al. (1999), between 0 and −1 mol m⁻² year⁻¹.

The distribution of carbon-related variables in the Canary Basin presents several characteristic features associated with the different water masses. AAIW is found close to the African shelf-break, characterised by the lowest pH₅ values (total hydrogen pH scale, normalized at 25 °C) in the Canary region, around 7.62. The signal of this water is traced by its pH along a zonal transect from 20°W to the African Coast (very close to the 29°N section in Fig. 4), and along the coast up to a latitude beyond 30°N, in front of Cape Ghir (Santana-Casiano et al., 2001). The northward transport of AAIW has also been identified by Pérez et al. (2001) from the analysis of nutrients, dissolved oxygen, alkalinity and dissolved inorganic carbon. The influence of the Mediterranean outflow Water (MW) is observed north of the Canary Archipelago at depths about 1100–1200 m, identified by a high value of total alkalinity (around 2360 µmol kg⁻¹) (Pérez et
al., 2001). MW water has also been observed north of La Palma Island in rotating lenses of high salinity and alkalinity (Santana-Casiano et al., 2001).

Nutrients are almost undetectable all year long within the mixed layer, except during winter when deep convective mixing incorporates nutrients from the upper-thermocline. At this time, nitrate concentration may increase to 1–2 μmol kg⁻¹ in the top 150 m (Pérez et al., 2001; Pérez-Rodríguez et al., 2001). Phosphate concentration follows this same pattern while the silicate distribution in surface water shows intermittent high values (about 2 μmol kg⁻¹) due to Sahara Dust events arriving to the region. Below the mixed layer we find the nutricline, where nutrient concentrations roughly increase with potential density except in those regions where the isopycnals approach the surface and nutrients are utilised. This correlation is indicative of southwestward water recirculation along isopycnals, following water subduction in regions where Ekman pumping is negative (Lectma and Bunker, 1978; McClain and Firestone, 1993). As the upper-thermocline waters recirculate, dissolved nutrients increase through the dissolution of falling particles and regeneration of organic matter (Kawase and Sarmiento, 1985). Nitrate is very well linearly correlated with phosphate throughout the whole water column. Nitrate is nonlinearly correlated with silicate within NACW due to redissolution of opal skeletons and decreased regeneration of nutrients with depth, particularly near the upwelling waters at the eastern boundary (Pérez et al., 2001).

Pérez et al. (2001; see also Castro et al., 1998) have described the chemical and nutrient characteristics of the eastern North Atlantic in terms of water masses, with their relative proportions obtained using a mixing model of temperature, salinity, nutrient, oxygen and carbon system variables. Table 1 shows a summary of representative values for the chemical characteristics of some of these water types. In the upper-thermocline, down to about 750 m, there is only NACW of subtropical and subpolar origin (although Pérez et al., 2001 introduce a third water type, which they named H, identified as the limit between these two water types). An exception is the eastern boundary where AAIW may be found below about 500 m, in a maximum percentage of 11%. For this reason, Pérez et al. (2001) introduced a new water type (AA) that contained the influence of NACW of subpolar origin and AAIW, characterised by large values of nutrients. They suggest that this water is rapidly diluted further north where any observed large nutrient concentration is caused by regeneration of organic matter associated to upwelling processes.

An important hypothesis posed by Pérez et al. (2001), on the basis of historical and recent data sets, is that during the last decades there has been a progressive decrease in the northward penetration of AAIW. It is possible, however, that the analysis of the scarce available data does not reflect a decadal tendency but simply a seasonal variability. This alternative explanation is based on the relation between the poleward undercurrent and the presence of AAIW and the poleward undercurrent. The hypothesis, to be discussed below, is that there is a seasonal cycle in the poleward undercurrent, responding to seasonal changes in the preferential southward path of the Canary Current.

### 3.3. The seasonal plankton production cycle

In the open ocean waters of the Canary region, the annual plankton production cycle is affected by the quasi-permanent seasonal thermocline. This sharp density transition restricts the vertical flux of inor-
ganic nutrients from upper thermocline waters to the euphotic zone, limiting phytoplankton growth during most of the year. As indicated above, the seasonal thermocline only disappears during winter as the result of surface cooling and the mixed layer reaches its maximum penetration depth (150–200 m, see Fig. 3). The seasonal thermocline begins to reform in April–May, leading to the more common situation of a surface euphotic zone depleted of nutrients. Therefore, it is only during the short mixing period that follows deep convection when phytoplankton will grow fast. In this circumstance, the phytoplankton builds up a “bloom,” the only condition being that cellular growth rate be higher than the grazing rate of zooplankton. The dynamics of this “late-winter bloom” is more similar to the autumn bloom (nutrient-limited) than to the spring bloom (light-limited) in temperate seas.

From late spring to early winter, most chlorophyll is found in a deep maximum (DCM) within the seasonal thermocline, and large phytoplankton cells are almost absent from surface waters. Indeed, small picophytoplankton (<2 μm)—characteristic of oligotrophic regions—is the most abundant autotrophic group, contributing to total chlorophyll and production more than 75% in autumn and up to 50% in spring (Montero, 1993; Aristegui et al., 2001). These cells, together with the heterotrophic bacteria and their consumers, the heterotrophic flagellates, are the main components of the autotrophic and heterotrophic biomasses, respectively, of the planktonic food webs. They are also mainly responsible for the remineralization of organic matter in surface waters (Aristegui and Montero, 2005). Nevertheless, the structure of this microbial community displays seasonal and spatial changes, to a large extent related with mesoscale variability induced by the flow passing through the islands and the nearby coastal upwelling.

Only during late winter peaks in chlorophyll and primary production may be observed in surface waters. These maxima (0.5–1.0 mg m⁻³) are significantly higher than the annual means (<0.2 mg m⁻³; Braun and Real, 1984), but much lower than the maxima (>2 mg m⁻³) observed in temperate waters during the spring bloom. Whether these values are the result of the injection of nutrients into the impoverished euphotic zone (resource control), the effect of the grazing pressure (consumer control), or both, is still a matter of debate. Aristegui et al. (2001) studied the short-term (days to weeks) temporal variation in a coastal planktonic community at Gran Canaria Island, to assess the effect of resource and consumer controls on the variability of the seasonal planktonic cycle. They observed a clear decoupling between the biomass and production of autotrophs as well as an inverse relationship between the biomasses of phytoplankton and mesozooplankton during the early stages of the bloom session. This led them to suggest that primary production could depend on a resource control (nutrients) while biomass could depend on a consumers (grazing) control. In oceanic waters the year round variation in chlorophyll concentrations, integrated over the whole water column, ranges between 15 and 60 mg Chl m⁻². This range of variation is relatively small as compared with the much larger variability in primary production values, at least one order of magnitude greater (Basterretxea, 1994). This supports the hypothesis that phytoplankton biomass is mainly controlled by grazing pressure.

Zooplankton biomass values are rather low during most of the annual cycle, although they depend on the proximity of the sampling site to the African coast and whether it is on the windward or leeward portion of the islands. North of Tenerife Island, the average biomass value between 0- and 200-m depth was 390 mg dry weight (dw)·m⁻² (Braun, 1981), while south of Gran Canaria Island the average value for the annual cycle was 1370 mg dw·m⁻² for the same depth strata (Hernández-León et al., 1984). This sharp difference is related to an island mass effect on the biomass values of plankton. A mesozooplankton bloom appears to coincide with increased mixing during late winter (Hernández-León, 1998; Hernández-León et al., 1984, 1998, 2001c), although an increase of biomass may be observed in July–August coinciding with the maximum intensity of the Trade Winds on the southern shelf of Gran Canaria Island (Hernández-León, 1988).

Mesozooplankton grazing in the Canary Island waters is relatively small, in agreement with observations for other oceanic areas. Based on indices of metabolism and assuming herbivorous feeding, Hernández-León et al. (1998) estimated that this potential grazing accounted for 10% to 46% of...
the primary production. Aristegui et al. (2001) found that mesozooplankton controlled 20% of primary production in a coastal station, while Hernández-León et al. (1999) found a value of 46% in the tropical oceanic waters. However, direct estimations of grazing using the gut fluorescence method showed values ranging from 11% to 22% of primary production (Hernández-León et al., 2001c) while ingestion (obtained from metabolism and growth indices) accounted by 20–37%. Up to 41% of the ingestion by mesozooplankton was non-pigmented food. Therefore, most of the grazing control in these waters should be due to protista. This is in agreement with the calculations made by Dam et al. (1995) for the tropical Pacific where most of the grazing is exerted by microzooplankton.

Finally, the control of zooplankton and protista on primary production can change depending on the bottom-up and top-down control in the rather complex food web of warm water ecosystems. Recently, Hernández-León (1998) and Hernández-León et al. (2001b) found a lunar cycle in zooplankton abundance similar to the one described for freshwater ecosystems (Gliwicks, 1986). A decrease in the abundance of copepodites plus adult copepods was observed from the second to the fourth quarter of the moon. During the full moon the large zooplankton and micronekton of the Deep Scattering layers (DSL) do not reach the upper mixed layer, in order to avoid predation under a relatively high level of illumination (Pinot and Jansá, 2001), and epipelagic zooplankton abundance increases due to the relatively low predatory pressure. On the other hand, during the new moon phase the diel vertical migrants reach the surface waters and prey upon epizooplankton, producing and abundance decrease. This lunar cycle in zooplankton could explain the 30-day period observed in sediment trap data in subtropical waters (Khripounoff et al., 1998). The effect of the diel vertical migrants promotes not only the intensity of the active flux in the ocean (see Hernández-León et al., 2001c for the assessment of this flux around the Canary Islands) but could also drive the variability of the gravitational flux. Moreover, zooplankton is the food of most larval stages of marine organisms and they described variability can also explain the lunar cycle usually observed in fish larvae and invertebrates in warm water ecosystems.

4. Coastal upwelling recirculation

4.1. Along and cross-shore circulation

Recent observations have led to the viewing of the Canary Current as a quite intense and well-localised boundary current, with its behaviour being intimately connected with the coastal upwelling region. The coastal upwelling region is maintained by the presence of northeasterly winds all year long. From Cape Blanc to the Strait of Gibraltar (roughly from 20 to 35°N) the winds are upwelling favourable all year long, although north of the Canary Archipelago they are intense during the Summer months and moderate or weak during all other seasons (Wooster et al., 1976; Speth and Detlefsen, 1982; Nykjaer and Van Camp, 1994). From Cape Ghir to Cape Yubi (roughly 28 to 31°N), the coastline presents a prominent concave shape. This, together with the morphology of the Atlas range in this region, causes the upwelling to be much reduced, even during summer. This breakage in coastal morphology and upwelling favourable winds has, as we will see below, some potential important consequences on the path of the Canary Current.

Upwelling is characterised by the presence of two main dynamic features: a vertical cell and alongshore currents (e.g., Allen, 1973; Brink, 1983). The vertical cell is induced by the alongshore winds; these produce offshore Ekman transport of surface waters that are replaced by relatively cold subsurface waters, approximately flowing along the isopycnals. If strong enough alongshore winds are maintained for a few days, then surface waters are retrieved from the coastal region and a frontal signature, marking a contrast between the surface layer and subsurface waters, shows off away from the coast (Csanady, 1977). The front is less visible in winter, because of the absence of the seasonal thermocline. The upwelling system responds rapidly to the upwelling favourable winds. In times of about only 1 day a baroclinic jet, in approximate geostrophic balance, sets up in the offshore side of the upwelling front. This jet reaches velocities up to 0.5 m s^{-1}, but these are limited to the upper layers (~50 m), so the total transport over a width of the order of the internal Rossby radius of deformation (~20 km) is typically less than 0.5 Sv (Allen, 1973; Brink, 1983; Pelegri and Richman, 1993). The baroclinic jet produces mixing between
the subsurface and surface layers, this vertical exchange for water and momentum setting up an offshore limit for the surface front. If the winds are maintained for times of the order of 1 or several weeks, then a relatively slow barotropic jet will develop in the poorly stratified upwelled waters found onshore of the surface front. This jet has large inertia and will remain even if the upwelling conditions temporally disappear. Furthermore, this barotropic flow may sustain several times the water transport associated to the baroclinic jet (Pelegrí and Richman, 1993).

This idealised two-dimensional picture is broken because of the finite longitude of the coast and the presence of irregularities in coastal morphology. The presence of the Gibraltar Strait interrupts the meridional extension of the upwelling system, so the alongshore baroclinic and barotropic transports require a supply of open ocean water. In the description of the eastern Atlantic circulation, this corresponds to a sink of surface and upper-thermocline water that recurrently appears in different analysis of historical data sets (Stramma and Siedler, 1988; Reid, 1994; Lozier et al., 1995; Siedler and Onken, 1996). An example is shown in Fig. 6, reproduced from Lozier et al. (1995), that shows the mean streamlines at the 26.5 and 27.0 isopycnals, typically found at depths of about 150 (just below the mixed layer) and 400 m in the Canary Basin. The speeds that correspond to these stream functions are about 1.5 and 0.7 cm s\(^{-1}\) into the coastal region, and suggest a transport of several Sv into the coastal upwelling region north of Cape Ghir. This is consistent with some 3 Sv inferred by Stramma (1984) for the top 800 m, and about 1 Sv inferred by Stramma and Siedler (1988) for the top 200 m. The connection between the upper thermocline of the open ocean and the coastal upwelling region is also evident in the distribution of potential vorticity at the 26.5 and 27.0 isopycnals (Fig. 6c and d, obtained from data provided courtesy of Susan Lozier, Duke University; see also Lozier, 1997). The distribution in the deeper isopycnal suggests a homogenization of potential vorticity in the coastal upwelling region north of the Canary Islands, hence possibly allowing the existence of southward recirculation along the slope. These results, however, were mostly obtained from open ocean data so they are still unable to provide the precise location and distribution of the water inflow. More recently, Pelegrí et al. (unpublished manuscript; see Laiz et al., 2001, and Ratsimandresy et al., 2001) have reported a mean flow of about 2 Sv into the coastal upwelling region between Cape Ghir and the Strait of Gibraltar.

The southward extension of the upwelling system is limited by both the existence of an inflexion in the African coastline, where winds are locally modified, and the presence of the Canary Archipelago. These two factors are probably responsible for the existence of the Ghir filament, a quasi-permanent surface structure that stretches offshore over 100 km near Cape Ghir (Hagen et al., 1996; Pelegrí et al., 2005). Pelegrí et al. (2005) estimate its net offshore transport to be about 0.5 and 1 Sv at times when the filament was weak and moderate, respectively. This appears to be only a fraction of the total inflow north of Cape Ghir so the conclusion is that most of the transport flows south, probably through the eastern Canary Islands or between the Canary Islands and the African coastline. The flow of the Canary Current through the eastern Canary Islands and the African coast has been estimated from several years of XBT and moored current meter data (Hernández-Guerra et al., 2002; Knoll et al., 2002), as well as from historical tide gauge and hydrographic data (Navarro-Pérez and Barton, 2001). The transport of the Canary Current flowing between Gran Canaria Island and the African coast is 1.8 ± 1.4 Sv southward. Approximately half of this transport flows through the channel between the islands of Gran Canaria and Fuerteventura and the other half through the channel between Fuerteventura and the African coast. The transport presents a significant seasonal variability, ranging between 1.2 ± 0.3 Sv in May and 2.6 ± 0.1 Sv in January. The transport through the channels presents reversals (northward flow) during May (between Gran Canaria and Fuerteventura) and November (between Fuerteventura and Africa).

The major picture arising from the above description is that the Canary Current flows south along the upwelling transition zone past the Canary Islands which constitutes the real eastern boundary condition for the open ocean. Fig. 7 shows the trajectories of surface buoys, dragged at a depth of
10 m, that were deployed at the Gulf of Cádiz on February 1988. It is remarkable that these buoys did not enter the Mediterranean Sea and followed a swift southwestward trajectory, with speeds typically over 0.1 m s\(^{-1}\), along the upwelling zone. This trajectory was not modified by the offshore Ekman transport associated to the northeasterlies at least until past the Canary Archipelago. South of the

Fig. 6. Mean stream function for the (a) 26.5 and (b) 27.0 isopycnals (reproduced from Lozier et al., 1995). Solid contour intervals in the Canary Basin are drawn every 0.5 m\(^2\) s\(^{-2}\). Thick lines indicate the locations where these isopycnals surface during winter and summer. (c) Logarithmic distribution of the potential vorticity for the 26.5 isopycnal. (d) Logarithmic distribution of the potential vorticity distribution for the 27.0 isopycnal. The numbers provide the exponents of base 10 in units of 10\(^{-12}\) s\(^{-1}\) bar, so a given value \(p\) would correspond to 10\(^p\) \times 10\(^{-12}\) s\(^{-1}\) bar. Data for c and d have been provided courtesy of Susan Lozier.
Canary Archipelago the buoys slowly drifted offshore, probably related to the great intensity of the Ekman vertical cell. The trajectories in this southern region are similar to trajectories of buoys deployed at Cape Yubi in August 1999, which followed south slowly drifting offshore, although when trapped by coastal filaments the offshore movement became very fast (Des Barton, personal communication).

A conceptual difficulty could arise when trying to connect this quite permanent picture with the intermittent character of upwelling. The permanence of the southward flow probably lies on the mixed baroclinic and barotropic character of the upwelling jet, with the barotropic component being capable of sustaining the southward transport even when upwelling weakens. The intermittent character of the upwelling zone is vividly illustrated by Plate 1, which presents several satellite images of the SST and the concurrent surface pressure fields. Plate 1a illustrates how the upwelling signature (coastal cold waters) is broken between Cape Yubi and Cape Ghir, despite surface pressure maps suggest strong upwelling favourable winds throughout the region. In Plate 1b the upwelling signal is very well defined from Cape Ghir until the west coast of the Iberian Peninsula. In the spring image (Plate 1c), despite the relatively weak winds, the SST signal provides a sense of continuity along the coast.

The observations of winter flow reversals in the upwelling zone near the Canary Archipelago are supported by a study of seasonal variability by Navarro-Pérez and Barton (2001). These authors have used hydrographic and tidal gauge data to describe a winter offshore diversion of the southward Canary Current as it approaches the Canary Archipelago. Seasonal changes in regional circulation are based on the repartition of the Canary Current in baroclinic and barotropic contributions. The baroclinic flow responds to the seasonal pattern of northeasterly winds, with maximum values taking place during summer. At this time this flow may often become unstable and depart from the coast as

![Fig. 7. Trajectories of surface buoys illustrating the coastal jet (original data courtesy of Pierro Zanasca, Saclante).](image)
surface filaments. On the other hand, the barotropic component slowly builds up, from spring till fall, with its maximum rate of change occurring during the strong summer winds. By late fall, southward transport is maximum and the flow becomes unstable and separates from the coast, retaking its southward motion further west between the central or western Canary Islands. At this time the eastern channels sustain northward flow, having its origin on a large cyclonic loop originated at Cape Ghir and the related surfacing of the poleward undercurrent as it flows north without opposition.

Fig. 8 is a schematic representation of the main geostrophic circulation in this region, as described above, from the mooring data and 3 years of XBT transects in the region. Fig. 8a shows the situation predominant most of the year and Fig. 8b represents the large loop that may form in late fall. The exact location of this loop, and how much of it flows north along the upwelling zone, is indeed speculative and needs to be verified through future measurements in the region. Note that these schematic diagrams do not include export of surface water through Ekman transport or upwelling filaments. The main modifications produced by the filaments cell would be localised distortions such as small loops, while the vertical upwelling cell would cause an offshore drift of the streamlines, mainly south of the Canary Archipelago.

4.2. Dissolved inorganic carbon and nutrient fluxes

In November 1995 the Canary Islands area acted as a source of CO₂, with the maximum CO₂ fluxes occurring in the northwest coast of Fuerteventura, where localised upwelling was taking place (Santana-Casiano et al., 2001). The effect of upwelling on the CO₂ system may also be observed in studies done during October 1999 between the Canary Islands and Cape Ghir. The introduction of subsurface water, rich in nutrients and with relatively high f/CO₂ values, to the nutrient-depleted ocean surface favours primary production, and in turn reduces the amount of dissolved inorganic carbon (Cᵢ) through photosynthesis. Using the above measurements, we may estimate a decrease of about 26 μmol kg⁻¹ in the surface waters between the upwelling and open ocean regions (from 2114 μmol kg⁻¹ at station 615 to 2088 μmol kg⁻¹ at station 619), approximately following the trajectory of surface water parcels in this region (Fig. 9a).

Phytoplankton uptake in the highly productive upwelling waters moving offshore explains the normalized partial pressure of carbon dioxide (NICO₂) distribution in Fig. 9c, with maximum values above 400 μatm in the upwelled waters at the northeastern extreme of the region. However, it is the temperature field in the oligotrophic region that controls the f/CO₂ values (Fig. 9b), with the offshore area acting during this time of the year as a source of CO₂ to the atmosphere with an average value of 5 mmol C m⁻² y⁻¹ (González-Dávila et al., 2003).

The vertical circulation cell associated to coastal upwelling transports coastal properties towards the open ocean. Wooster et al. (1976) and Nykjaer and Van Camp (1994) have obtained characteristic values of cross-shore Ekman transport between the Canary Island and the Strait of Gibraltar between 0 and 1.5 m² s⁻¹ (per meter of coastline), depending on the season. If we multiply this figure by 5 mmol m⁻³ (a typical mean value of the nitrate concentration within this layer in the upwelling band; Minas et al., 1982), we obtain that the maximum offshore nitrate transport is about 0.7 mmol s⁻¹. Over a distance of 1000 km this would account for a net nitrate export up to some 7 kmol s⁻¹ during the maximum upwelling season. This value may be compared with the amount of nitrate transported alongshore by the Canary Current imbedded in the upwelling system. Assuming a depth integrated mean nitrate concentration of 8 mmol m⁻³ in the top 400 m (Minas et al., 1982) and multiplying this value by a southward water transport of 2 Sv leads to a southward nitrate transport of some 16 kmol s⁻¹. Despite the involved approximations and even taking into account the possibility that a significant fraction of nutrients is lost at the filaments (acting as point sinks), it is clear that an important fraction of the nutrient finds its way south beyond the Canary Islands. These results may be compared with calculations of southward alongshore nitrate transport at about 32°N (north of Cape Blanc) by Marrero and Pelegrí (1995), between 5 and 7 kmol s⁻¹, which indicates the meridional continuity of the southward nutrient transport in the upwelling system off northwest Africa.
Fig. 8. Schematic diagram illustrating the mean geostrophic transport of upper-thermocline waters (down to about 700 m) in the region. Each line roughly corresponds to one Sverdrup. (a) Predominant situation during most of the year; (b) situation in late fall.
Fig. 9. (a) Temperature, (b) $f\text{CO}_2$, and (c) NfCO$_2$ distributions in the upwelling transition zone north of the Canary Archipelago during October 1999. The location of stations mentioned in the text is shown in a; see also Fig. 4.
4.3. Plankton productivity in upwelling waters

Coastal upwelling off northwest Africa is characterised by high wind stress that forces relatively deep mixing (Longhurst, 1998), such that nutrient differences in the upwelling source water set the potential of primary production. Nutrient-rich South Atlantic Central Water (SACW) upwells south of Cape Blanc, with some interleaving across the meridional front (Barton, 1987), while north of Cape Blanc the upwelling water is typically NACW. Typical nutrient contrast between NACW and SACW (at a level immediately below the seasonal thermocline, \( \sigma_t \approx 26.5 \)) is rather large, for nitrate decreasing northwards from over 20 to 5 mmol m\(^{-3}\) (Zenk et al., 1991; Pérez-Rodríguez et al., 2001). Compiled productivity values in the whole region (Table 2) yields an average annual estimate of 2.4 g C m\(^{-2}\) day\(^{-1}\), a value close to that computed from satellite chlorophyll and averaged photosynthetic parameters (Longhurst et al., 1995). Nevertheless, the highest annual rates are probably produced between Cape Barbas and Cape Blanc, where SACW is available as source water and upwelling is produced all year long.

Significant changes in the nutrient regime and primary production are observed on short-term scales, correlated with variations in the local winds (Grall et al., 1982). The centre of the upwelled water migrates from the inner shelf to the shelf break as upwelling progresses. The high wind stress and strong equatorward and cross-shelf currents may prevent accumulation of organic matter in sediments, where aerobic respiration dominates. Concentrations of particulate organic matter nearshore are in the low end among all coastal upwelling regions, with a predominance of <50-µm particles (Lenz, 1982). However, mineralization rates and nutrient fluxes are as high as in other upwelling regions (Minas et al., 1982). Fluxes of nutrients from sediments into the well-mixed water column are appreciable, and silica dissolution may exceed rates of silica uptake by phytoplankton (Codispoti et al., 1982). High ammonia concentrations are homogeneously distributed in the turbid inshore waters, where ammonia uptake by phytoplankton is inhibited. Turbidity results from wind mixing and aeolian dust deposition blown into the sea from the Sahara desert. High nutrient regeneration rates may result also from organic matter recirculated by mesoscale features into the nearshore environment. Fish populations contribute mainly to regeneration of phosphate and urea, while ammonia is equally regenerated both by zooplankton and fish (Smith and Whitledge, 1982).

Organic carbon flux rates near the shelf bottom may exceed primary production rates, indicating that a large fraction of the nearshore material is resuspended. This material may be lost out from the system by cross-shelf advective transport (Longhurst, 1998). Large filaments of cool water like those stretching from Cape Ghir and Cape Blanc (Van Camp et al., 1991) may export large proportions of coastal produced organic matter hundreds of kilometers into

<table>
<thead>
<tr>
<th>Region</th>
<th>Latitude (N)</th>
<th>Season</th>
<th>Daily primary production (g C m(^{-2}) day(^{-1}))</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Average</td>
<td>Min.</td>
</tr>
<tr>
<td>C.Sim–C.Ghir</td>
<td>31.5–30.5</td>
<td>Summer</td>
<td>2.4</td>
<td>0</td>
</tr>
<tr>
<td>C.Sim–C.Ghir</td>
<td>31.5–30.5</td>
<td>Winter</td>
<td>1.3</td>
<td>0.2</td>
</tr>
<tr>
<td>C.Sim–C.Ghir</td>
<td>31.5–30.5</td>
<td>Summer</td>
<td>1.3</td>
<td>0.6</td>
</tr>
<tr>
<td>C.Sim–C.Ghir</td>
<td>31.5–30.5</td>
<td>Autumn</td>
<td>1.5</td>
<td>1.0</td>
</tr>
<tr>
<td>C.Jubi–C.Bojador</td>
<td>28.5–26.5</td>
<td>Summer</td>
<td>3.1</td>
<td>1.3</td>
</tr>
<tr>
<td>C.Jubi–C.Bojador</td>
<td>28.5–26.5</td>
<td>Summer</td>
<td>2.4</td>
<td>1.3</td>
</tr>
<tr>
<td>C.Corveiro–C.Blanc</td>
<td>22.0–21.0</td>
<td>Spring</td>
<td>2.4</td>
<td>0.8</td>
</tr>
<tr>
<td>C.Corveiro–C.Blanc</td>
<td>22.0–21.0</td>
<td>Spring</td>
<td>2.4</td>
<td>1.1</td>
</tr>
<tr>
<td>C. Blanc</td>
<td>21.0</td>
<td>Spring</td>
<td>1.15</td>
<td>0.3</td>
</tr>
<tr>
<td>C. Blanc</td>
<td>21.0</td>
<td>Spring</td>
<td>0.96</td>
<td>0.2</td>
</tr>
<tr>
<td>C. Blanc</td>
<td>21.0</td>
<td>Summer</td>
<td>1.62</td>
<td>0.4</td>
</tr>
<tr>
<td>C. Timiris–Nouakchott</td>
<td>19.5–18.0</td>
<td>Spring</td>
<td>3.9</td>
<td>1.6</td>
</tr>
<tr>
<td>NW Africa upwelling</td>
<td>31.5–18.0</td>
<td></td>
<td>2.4±1.5</td>
<td></td>
</tr>
</tbody>
</table>
the ocean domain (Gabric et al., 1993; García-Muñoz et al., 2005). Mineralization of organic matter along filaments must constitute an important nutrient source for new production in nearby oceanic communities.

Zooplankton biomass in the upwelling area varies with latitude and time, in relation with changes in wind intensity (Postel, 1990). The highest values are found near Cape Blanc along the entire annual cycle. To the north the maximum values are observed in summer and the minimum in winter, the opposite situation occurring south of Cape Blanc. The increase in zooplankton biomass appears to be uncoupled with the maximum of phytoplankton biomass and changes in zooplankton composition and growth are observed as the water mass is transported offshore. The species composition varies depending on the seasonal upwelling cycle of the upwelling (Thiriot, 1978). Calanoides carinatus is the most common species during the upwelling season. During the non-upwelling season it remains in overwintering at mesopelagic depth (Vives, 1975), while during the upwelling season it appears at the surface where it grazes the high diatom densities usually found over the African shelf. Maximum abundance is found near the upwelling centres coinciding with high values of primary production. As the water masses are transported offshore, C. carinatus is observed at increasing depth coinciding with the decrease in primary production and phytoplankton biomass. Euphausiids are also observed offshore as strong diel vertical migrants. Euphausia khronii is the most abundant species (80–90% of the euphausiids) north of Cape Blanc. Other organisms such as Thalia democratica are observed over the slope waters forming high-density swarms.

5. Upwelling filaments

5.1. Structure and dynamics

Considerable knowledge on the dynamics and structure of filaments came after the Coastal Transition Zone interdisciplinary program off California. A summary of the program may be found in Brink and Cowles (1991) and a description of the major physical characteristics of the observed filaments was given by Strub et al. (1991). Coastal filaments are relatively narrow (typical widths of a few tens of kilometers) near surface (usually no deeper than 150 m) structures that stretch offshore (up to several hundred kilometers) from near the coast. The core of a filament flows offshore with rather large velocities, usually between 0.25 and 0.5 m s⁻¹, and it is common to find onshore flow at both sides of the filament associated to one or two vortices. These structures have been observed in all eastern boundary currents of subtropical gyres and are almost always associated to upwelling favourable conditions, such that they contain relatively cold water of subsurface origin.

According to Strub et al. (1991) and Ramp et al. (1991), filaments may arise from one or a combination of several factors: baroclinic instability of the coastal current, irregularities in coastline and bottom topography, coastal convergence caused by wind stress, and the interaction of the offshore eddies with the coastal region. One important element to consider is that filaments are a very common feature off California during the upwelling season, when the California Current flows south. During winter the winds reverse, upwelling ceases and the current (now the Davidson Current) flows north, filaments being very uncommon. Other clues on the origin of filaments are that they contain nutrient-rich cold water of subsurface origin and that their location sometimes appears to be related to major coastal or topographical features, but in other instances there is no clear association.

A similar situation occurs at the eastern boundary of the North Atlantic subtropical gyre. Off the Iberian Peninsula, where the alongshore current reverses seasonally, filaments are a common feature during the upwelling season (Haynes et al., 1993) but they are absent when the water flows north (Haynes and Barton, 1990; Barton, 1998). South of the Strait of Gibraltar upwelling takes place all year long, although with a clear summer maximum, and the Canary Current continuously flows south. Filaments are here a recurrent feature all year long, although the small winter surface temperature contrast between coastal and open waters makes it hard to appreciate their SST signal during this season.

In any satellite SST image, we usually appreciate a number of filaments emanating from the northwest African coast, two of them being major permanent structures: the Cape Ghir and the Cape Blanc filaments. The latter is a giant filament that typically
stretches nearly 500 km off Cape Blanc. It lies south of our region of interest, its great size and persistence being linked to the convergence of North Atlantic and South Atlantic upper thermocline waters at the Cape Verde frontal zone (Gabric et al., 1993). The Cape Ghir filament is a rather ubiquitous feature that usually stretches offshore over 150 km. It is visible in most colour and SST satellite images of the region (e.g. Plate 1; see also Van Camp et al., 1991 and Hernández-Guerra et al., 1993) and even appreciable from the mean surface temperature fields, obtained from either hydrographic (Mittelstaedt, 1991) or satellite data (Hernández-Guerra and Nykjaer, 1997). Several hydrographic cruises off Cape Ghir have been documented by Hagen et al. (1996) and Pelegrí et al. (2005). These cruises covered the Cape Ghir plateau, off the northern flank of Cape Ghir, and the Agadir submarine canyon, immediately south of this Cape. The filament was always present, even during non-favourable upwelling winds, and consisted in a shallow feature that did not extend beyond about 200 m. Water flows offshore in the central portion of the filament, with characteristic speeds over 0.25 m s\(^{-1}\), and recirculates onshore at both sides of the filament as two oppositely rotating eddies. South of the filament a deep anticyclonic eddy, its origin probably linked to the poleward undercurrent over the slope, was found in two October cruises. Fig. 10a, reproduced from Pelegrí et al. (2005), presents the temperature field at a depth of 20 m and the velocity field obtained from ADCP measurements in the 25- to 75-m layer at a time of a relatively weak filament that resulted from non-favourable upwelling winds.

South of Cape Ghir, it is usual to find other filaments emanating from the coast. Hernández-Guerra et al. (1993) showed different satellite images with filaments either detaching from Cape Yubi or Cape Bojador, and Barton (1998) and Barton et al. (1998) discussed a situation when a large filament emanated from somewhere between these two Capes. There are many instances, however, when there are no apparent filaments in this region (Nykjaer, 1988; Van Camp et al., 1991; Arístegui et al., 1994). Other minor filaments have also been observed in the northwest coast of Fuerteventura in the Canary Islands (Real et al., 1981; Molina and Laatzen, 1989).

All the above features point at the relation between filaments and coastal upwelling, and suggest that under some circumstances they may indeed be linked to the baroclinic instability of an intense coastal current, probably triggered by topographic features. The reasoning behind this argument is that during the upwelling season the coastal jet is intense enough such that it becomes unstable. The potential energy available in the eastern boundary system would justify that filaments appear even when the upwelling favourable conditions weaken. This would also justify that no filaments are present when the eastern boundary flows north, since these flows have downwelling isopycnals at the coast and no potential energy is available from this system. It seems surprising, however, that filaments are absent in intense western boundary currents. This suggests that an additional mechanism, based on the importance of bottom friction acting over steep slopes, could also be relevant. Lee et al. (2001) have proved that the main source of relative vorticity over the continental slope is the bottom stress acting over the seaward deepening water column. In a southward current flowing over the continental slope, this mechanism would provide positive vorticity so the flow would not be capable to adjust to the local planetary vorticity. When this happens we could expect that it would move away from the slope. This mechanism relies on the existence of a bottom boundary layer over the continental slope but this has indeed been observed to be important in the region (Badan-Dangon et al., 1986).

### 5.2. Distribution of nutrients and carbon

Several authors (Minas et al., 1982; Basterretxea, 1994; García-Muñoz et al., 2005) have reported the existence of relatively high levels of nutrients at the base of a filament, in concordance with the subsurface origin of these waters. Nutrient concentrations decrease in the offshore direction and become rather exhausted at the offshore extreme of the filament. The nutrient field is coherent with the temperature field and maintains an inverse linear relation with chlorophyll valies. These characteristics indicate that nutrients are being utilised as the water parcels move offshore, and suggest that the direct export of nutrients by the filaments to the open ocean may be relatively small, at least compared with the export of organic matter. The regeneration of exported organic matter,
Fig. 10. Distribution of properties in the Cape Ghir region from September 28 to October 1, 1997. (a) Temperature field at 20-m depth (left) and velocity vectors in the 25- to 75-m layer (right) adapted from Pelegrí et al. (2005). (b) Mean depth (top 100 m) distribution of Chlorophyll $a$ (Chl $a$, mg m$^{-2}$), particulate organic carbon (POC, mmol m$^{-2}$), dissolved organic carbon (DOC, mmol m$^{-2}$) and nitrate+nitrite (mmol m$^{-2}$). Reproduced from García-Muñoz et al. (2005).
however, causes an indirect export of both nutrients and dissolved organic carbon, reflected in a high correlation between these quantities (Minas et al., 1982; Pérez et al., 2001). Fig. 10b, adapted from García-Muñoz et al. (2005), illustrates some of these features at a time of a relatively weak filament.

Despite most nutrients appear to be utilised along the offshore excursion of water parcels, it is unquestionable that filaments represent a point source of nutrients along the coast of magnitude much larger than the source of nutrients. García-Muñoz et al. (2005) have computed a net export of nitrate+nitrite of 2.6 kmol s$^{-1}$ by the Cape Ghir filament, most of it related to significant nutrient concentrations and offshore velocities at depths between 50 and 100 m. This value is indeed much larger than the offshore export that would take place in an upwelling system along a shore distance of only a few tens of kilometers, equivalent to the filament width. Using a maximum water Ekman transport of 1.5 m$^2$ s$^{-1}$ (per meter of coastline) and a typical value of 3 mmol m$^{-3}$ for nitrate concentration at the base of the seasonal thermocline, we may estimate that the nitrate Ekman transport through a 20-km wide filament would be about 0.1 kmol s$^{-1}$, less than 5% the actual advective transport estimated by García-Muñoz et al. (2005). This clearly points at the character of filaments as locations of coastal flux convergence and export to the open ocean.

The effect of upwelling filaments on the CO$_2$ system may also be observed from the measurements taken in the area near Cape Ghir during October 1999 (Fig. 9). Filaments stretch into the open ocean and introduce a perturbation in the distribution of chemical characteristics, the most visible feature being a change in the partial pressure of CO$_2$ in surface waters. We may use stations 615 (within the Cape Ghir filament) and 612 (just out of the filament) to visualise what happens to the carbon content of a water parcel exported away of the upwelling region through a filament. Total dissolved inorganic carbon and partial pressure of CO$_2$ increase by 13 µmol kg$^{-1}$ and 19 µatm, respectively (Fig. 9b). These values, however, cannot be directly used to conclude whether the region acts as a source or sink of CO$_2$. Before doing so, we must consider the thermodynamic (temperature and salinity change) and air–sea exchange effects. Our observations indicate that the surface temperature, salinity and pH increase from the upwelling region to the open ocean. Temperature increases by 3.9 °C and salinity and pH increase by about 0.2 and 0.05 units, respectively. Normalisation to a constant temperature of 21.5 °C (NiCO$_2$) indicates that the fCO$_2$ experiments a large increase in the upwelling region because of the sole thermodynamic effect in CO$_2$ solubility (Fig. 9c). This is consistent with the fact that in the upwelling region there is upward pumping of subsurface CO$_2$ rich waters, i.e. fCO$_2$ augments by some 40 µatm at a depth of 100 m.

For a water parcel moving from station 615 to station 612, for example, the thermodynamic effect would increase fCO$_2$ by 64 µatm while the air–sea exchange effect on the mixing layer would produce an increase of less than 1 µatm. Hence, in the absence of biological effects, we would expect that the water parcel experimented changes much larger than actually observed. This suggests that there is significant nutrient utilisation and biological production in the area. For example, to consume 28 µmol kg$^{-1}$ of dissolved inorganic carbon (corresponding to a change of 45 µatm in fCO$_2$, or the difference between the thermodynamic imbalance and the actual increase between our reference stations) requires 3.9 µmol kg$^{-1}$ of nitrate (C$_{ORG}$:N=7.2±0.8, Körtsgen et al., 2001). The time scale for utilization of this nitrate is relatively low, typically of the order of a few days (Louanchi and Najjar, 2001, Agustí et al., 2001). Using an outward velocity value of 0.1 m s$^{-1}$ and considering that the two stations are separated by a distance of 73.5 km, we may estimate a utilization time of 8.5 days. For a 70 m deep euphotic layer, this corresponds to a mean nitrate utilization rate of about 1 mmol m$^{-3}$ day$^{-1}$. This estimate is of the same order as the Agustí et al. (2001) calculations for the downward average flux of dissolved organic nitrogen in the eastern subtropical Atlantic, 15 mmol m$^{-2}$ day$^{-1}$, pointing at coastal filaments as a major input of organic nitrogen and carbon to this region.

The mean C$_T$ over the upper 150 m decreases from 2114 µmol kg$^{-1}$ at station 615 to 2100 µmol kg$^{-1}$ at station 612. This decrease in such a small distance suggests that the biological utilization of the injected nutrients inside the filaments decreases the concentration of dissolved inorganic carbon. The upward and
offshore supply of inorganic nutrients, however, does not necessarily lead to a net uptake of atmospheric carbon, since this nutrient supply is associated with a supply of dissolved inorganic carbon. It is the low temperature of the upwelled water and the concentration of inorganic carbon in the source water that determine whether the upwelling area will behave as a net CO$_2$ sink or source. At the time of our observations, during an intense upwelling event, we may conclude that physico-chemical processes dominated over the biological pump in determining that the upwelling region behaved as a net CO$_2$ sink. These results are confirmed by observations of an average carbon flux of 0.5 mmol m$^{-2}$ day$^{-1}$ from the atmosphere to the ocean in the upwelling region (González-Dávila et al., 2003). A similar finding was described by Agustí et al. (2001) around this area with a net upward flux of dissolved inorganic carbon and downward flux of organic carbon.

5.3. Plankton distribution and organic matter transport

Upwelling filaments may enhance the exchange of biological properties between upwelling and open ocean waters, since their offshore transport is usually significantly larger than Ekman transport (Kostianoy and Zatsepin, 1996; Navarro-Pérez and Barton, 1998). Nevertheless, the return flow of filament water into the coastal upwelling jet may reduce the impact of the overall offshore advective transport. In August 1993 an upwelling filament stretching between Cape Jubi and Cape Bojador was surveyed. The filament extended approximately 150 km offshore where it wrapped around a cyclonic eddy of 100-km diameter, before returning shoreward. Waters with high chlorophyll content, dominated by phytoplankton cells larger than 2 μm, were mainly restricted to the shelf and to a few stations along the filament. Although grazing could actively control the phytoplankton abundance along the filament (Hernández-León et al., 2002), the rapid warming of surface water with distance offshore probably induces mass sedimentation of diatoms in slope waters, and a drastic change in the phytoplankton community structure (e.g. Brink and Cowles, 1991). Indeed, a sharp drop in chlorophyll ascribed to cells larger than 2 μm was evident in the boundary between the upwelling and filament waters (Basterretxea and Aristegui, 2000), coinciding with a change from a diatom-dominated community in onshore waters to a cyanobacteria-dominated community in filament waters (Van Lenning, 2000). This scenario is similar to the one described by Joint et al. (2001), who observed drastic changes in plankton community structure within a filament of the Iberian upwelling system as it progressed offshore. Near the coast they found high chlorophyll concentrations associated with a diatom-dominated assemblage, which included high biomass of heterotrophic dinoflagellates and ciliates, while in the offshore extension of the filament they observed low chlorophyll values associated with a picoplankton community.

Basterretxea and Aristegui (2000) have found that phytoplankton assimilation numbers, and thus primary productivity, show a marked gradient from upwelling to open ocean waters. The assimilation numbers changed over one order of magnitude, paralleling changes in biomass and phytoplankton community structure. During their study they found that the filament was clearly acting as a recirculation loop. Seaward biomass transport occurred mainly in the mixed layer, whereas shoreward transport presented maxima at deeper layers, the net transport being onshore. This situation, however, differs from other field studies (Aristegui et al., 1997; García-Muñoz et al., 2005; Aristegui and Montero, 2005), or from studies based on ocean colour images (e.g. Hernández-Guerra et al., 1993), in the same region, where high organic matter concentrations transported by filaments were observed. Aristegui and Montero (2005) showed that coastal upwelling waters invading the eastern Canary region in the form of filaments can transport either water with low plankton respiration and large phytoplankton cells or water with high respiratory rates associated with small cells. Hence, the nature of the filament and its genesis, as well as the stage of development of the filament structure, are critical for the biological consequences of the filament in terms of plankton productivity and export of carbon to the open ocean.

A close relationship between the structure of the filament and zooplankton biomass distribution has been identified by Hernández-León et al. (2002). They find high biomass, specific gut fluorescence and electron transfer activity (ETS) along the filament structure in contrast to relatively low values of these
parameters at the centre of a cyclonic eddy enclosed by the upwelling filament. High values of zooplankton biomass are observed in all the filament structure in contrast to chlorophyll and primary production, which decrease towards the African shelf. The longevity of these organisms as well as a change from a plant-based to a non-pigmented (protista) diet allows these organisms to move up to 100 km from the coast. Of great interest is the transport of fish larvae observed in the filament structure (Rodríguez et al., 1999). Neritic species such *Sardina pilchardus* are found along the filament signature, while oceanic species are excluded. These neritic species are good tracers of the filament physical structure, which transports them into oceanic waters and near the Canary Islands (Bécognée et al., unpublished manuscript), where they have an important impact in the local fisheries.

6. Island-generated mesoscale structures

6.1. Generation and dynamics

The Canary Archipelago acts as a barrier to the prevailing winds and currents, representing an exceptional natural example of flow perturbation by obstacles. Two main features arise as a consequence of flow perturbation: cyclonic and anticyclonic eddies and warm lee wakes. Cyclonic eddies, detaching from the western flank of the islands, have been identified downstream of the islands of La Palma, Gomera and Tenerife from satellite observations (Hernández-Guerra et al., 1993; Pacheco and Hernández-Guerra, 1999; Barton et al., 2000), and downstream of Gran Canaria island from both satellite and in situ observations (Aristegui et al., 1994, 1997; Hernández-Guerra et al., 1993; Pacheco and Hernández-Guerra, 1999; Barton et al., 1998, 2000; Marrero-Díaz et al., 2001; Basterretxea et al., 2002). Anticyclonic eddies, detaching from the eastern flank of the islands, have been observed downstream of Tenerife and Gran Canaria islands from both satellite and in situ observations (Aristegui et al., 1994, 1997; Pacheco and Hernández-Guerra, 1999; Barton et al., 1998, 2000; Molina et al., 1996). SST images have also shown anomalous high surface temperature filaments that stretch southwest at the lee side of most archipelago islands, in the direction of the mean velocity surface field (Hernández-Guerra, 1990; Van Camp et al., 1991). These warm oceanic wakes, or lee filaments, are surface structures whose origin lies in the intense day-time warming of the unstirred surface in the wind-sheltered area of the islands. They occur, to some extent, in all the islands of the archipelago but are particularly clear downwind of the higher islands: La Palma, Gomera, El Hierro, Tenerife and Gran Canaria (Barton et al., 2000, 2001; Basterretxea et al., 2002).

Eddies are a recurrent mesoscale feature related to the Canary Current perturbation by Gran Canaria. Pacheco and Hernández-Guerra (1999) analyzed 8 years of satellite colour (Coastal Zone Colour Scanner, CZCS) images and observed the signal of cyclonic eddies associated with Gran Canaria, Gomera, and La Palma and anticyclonic eddies associated with Gran Canaria all during the year. Eddies have also been detected in most surveys conducted south of Gran Canaria, capable of resolving mesoscalar structures (Aristegui et al., 1994, 1997; Barton et al., 1998; Basterretxea et al., 2002). Table 3 summarises the number of eddies sampled during different surveys carried out on Gran Canaria waters, with eddies observed during all seasons. Cyclonic eddies were located southwest of Gran Canaria, either attached to the coast or at approximately one island diameter (50 km) from it. Anticyclonic eddies have been located either attached to the south coast or approximately at one island diameter from it. The high spatial coverage provided by satellite images indicates that the simultaneously occurrence of cyclonic and anticyclonic eddies downstream of Gran Canaria, forming a von Kármán like vortex street, is a quite common feature.

### Table 3

<table>
<thead>
<tr>
<th>Survey</th>
<th>Date</th>
<th>Cyclonic</th>
<th>Anticyclonic</th>
</tr>
</thead>
<tbody>
<tr>
<td>AXBT-1</td>
<td>5 May, 1989</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>AXBT-2</td>
<td>20 February, 1990</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>AXBT-3</td>
<td>24 May, 1990</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>EMIAC 9006</td>
<td>6–12 June, 1990</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>EMIAC 9103</td>
<td>9–17 March, 1991</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>CANARIAS 9110</td>
<td>25 Oct–5 Nov, 1991</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>MAST 9308</td>
<td>18–24 Agust, 1993</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>FRENTES 9806</td>
<td>6 June–8 July, 1998</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>
The mean radius of eddies generated by an island is comparable to the island size. These eddies usually exhibit an elliptical shape, in a similar fashion as observed by Patzert (1969) for eddies shed from Hawaii Island. Patzert suggested that the elliptical shape indicates that the eddy is at an early stage, and argued that an eddy’s eccentricity decreases with increasing distance of the island. Pacheco and Hernández-Guerra (1999) used CZCS images to calculate the eccentricity as a function of distance from the island, for an anticyclonic eddy associated with Gran Canaria and for a cyclonic eddy associated with Gomera. The results show that the Gran Canaria anticyclonic eddy eccentricity decreases when the eddy moves further away from the island. However, the Gomera cyclonic eddy behaves in a completely inverse way. They explained this anomalous behaviour, pointing out that the vicinity of El Hierro Island may disturb the normal evolution of the eddy.

As a consequence of geostrophic adjustment, the cyclonic eddies have a central uplifting of the isotherms, being cold core water structures, whereas anticyclonic eddies cause the central downlifting of the isotherms, becoming warm water core structures. The perturbation of the temperature field by an eddy is noticeable from the near surface layer until at least 600 m (Sangrà, 1995). Intense cyclonic eddies may uplift the isotherms by some 60 m, introducing temperature negative anomalies at about 2.5 °C. On the other hand, the sampled anticyclonic eddies seem to be less intense, causing a depression of the isotherms of the order of 45 m and introducing positive temperature anomalies about 1 °C. Secondary ageostrophic circulation associated to cyclonic (anticyclonic) eddies appears to cause upwelling (downwelling) of deep (surface), nutrient-rich (-poor), waters in their cores as a results of outward (inward) advective flux of surface water. From the outward trajectory of an Argos buoy, deployed in a cyclonic eddy shed by Gran Canaria, Barton et al. (1998) estimated an upwelling velocity higher than 50 m day\(^{-1}\).

Three Argos buoys where deployed on an anticyclonic eddy spun off by Gran Canaria in early June 1998 (Sangrà et al., 2004). Fig. 11 shows the whole trajectory of one buoy, an anticyclonic eddy-like path lasting during 199 days, which suggests that Gran Canaria eddies could last as a coherent structures for many months. The rotation period of all three buoys increased from 3 to 6 days, which indicates that the rate of rotation of the eddy decreases along its lifetime. The initial radius of the eddy, derived from an XBT survey conducted previous to the buoy deployment, was about 25 km, while the final radius, computed from the buoy’s orbit, was about 50 km. This suggests that the radius of the eddy also increases

![Fig. 11](image-url)
along its life time. The orbital trajectories (computed subtracting the mean eddy translation) also show an oscillation of the radius around its mean value, which suggests that the eddy evolves on a pulsating mode.

One plausible mechanism for the generation of eddies is the perturbation of the Canary Current by the islands, but this depends on whether the Canary Current is energetic enough to produce vortex shedding. Sangrà (1995) demonstrated numerically that if the Canary Current remains above 0.1 m s\(^{-1}\) then a von Kármán street is formed downstream of Gran Canaria. Direct measurements and geostrophic calculations indicate that a mean velocity of 0.2 m s\(^{-1}\) may occasionally be attained (Sangrà, 1995) but, according to Navarro-Pérez and Barton (2001), the maximum mean geostrophic value of the Canary Current through the Canary Archipelago is relatively low, roughly 0.05 m s\(^{-1}\). Therefore, only during certain times of the year the Canary Current is energetic enough to generate eddies. During low current velocity and moderate trade wind periods, another mechanism that favours eddy generation may be Ekman pumping originated through wind shear at the islands wakes (Barton et al., 1998, 2000, 2001; Basterretxea et al., 2002). Recent numerical simulation of eddy shedding by Gran Canaria, which includes wind forcing, shows that the vorticity input by the wind through Ekman pumping at the island wake helps to produce vortex shedding at relatively low incident current velocity (Jiménez and Sangrà, 1999). For example, for a wind of 10 m s\(^{-1}\) vortex shedding is produced when the incoming current is above just 0.04 m s\(^{-1}\), which approaches the mean values reported by Navarro-Pérez and Barton (2001).

6.2. Source of nutrients and dissolved inorganic carbon

Eddies generated at the lee of the Canary Islands produce an important perturbation in the distribution of chemical properties in the surface euphotic layers. Cyclonic eddies are a source of subsurface nutrient-rich water into the euphotic layers (Aristegui et al., 1994, 1997). There are several mechanisms through which this may occur. First, the central core of cyclonic eddies in geostrophic balance is characterised by the upwelling of isopycnals, hence physically bringing the subsurface layers close to the surface (Section 6.1). Second, there appears to be actual ageostrophic isopycnal flow into the eddy core, particularly during the early stages of eddy development when geostrophic equilibrium has not yet being fully attained, which leads to near surface central divergence (Barton et al., 1998). And third, diapycnal mixing likely takes place at different eddy phases and locations, probably linked to a localised increase in diapycnal shear (Rodríguez-Santana et al., 1999, 2001). The combination of these mechanisms is probably what makes these cyclonic eddies to be an efficient source of nutrients for the surface euphotic layers. Water parcels in early stages of anticyclonic eddies, on the other hand, appear to undergo radial motions towards the eddy centre of order 0.01 m/s, as suggested from the analysis of surface buoy trajectories (Fig. 11). This radial convergence leads to surface central waters low in nutrient concentration but with high accumulation of biomass.

Although cyclonic and anticyclonic eddies have a quite opposite behaviour, which leads to characteristic different nutrient and chlorophyll distributions, they both appear to undergo intense diapycnal mixing at intermediate distances between their centre and periphery, where diapycnal shear is maximum. Vertical (diapycnal) velocities in these regions may be as large as 10\(^{-5}\) m s\(^{-1}\) (A. Rodríguez-Santana, personal communication), which would explain the intermittent character of nutrient anomalies observed at some stations (Aristegui et al., 1997).

The area between La Palma, La Gomera and Tenerife is characterised by coexisting cold- and warm-core eddies, which contribute to exchanging CO\(_2\) between the atmosphere and the ocean (González-Dávila et al., 2005). The cold-core cyclonic eddies have surface waters with high partial pressure of CO\(_2\) and nutrient concentrations inside the eddy, while relatively low partial pressure of CO\(_2\) and nutrient values are observed near the eddy limits. Warm-core anticyclonic eddies are associated with low partial pressure of CO\(_2\) and relatively high chlorophyll concentration inside the eddy. Surface values of fCO\(_2\) show that this area acts as a source of CO\(_2\) to the atmosphere, but with a value that is only half (5.4±2.0 mmol m\(^{-2}\) day\(^{-1}\)) of that determined at the same time north of the Canary Islands and during similar wind conditions (González-Dávila et al., 2005). This is due to the
effects of both the solubility and the biological pumps associated with the cyclonic eddies where chlorophyll enhancement follows a decrease in $C_T$.

6.3. Plankton response to mesoscale variability

Nutrient pumping into the euphotic zone by cyclonic eddies enhances phytoplankton primary production and consequently increases chlorophyll concentrations (Aristegui et al., 1997; Basterretxea, 1994; Barton et al., 1998; Basterretxea and Aristegui, 2000). Conversely, anticyclonic eddies concentrate warm surface water and organic matter in their cores, deepening the mixed layer and sinking organic matter below the euphotic zone (Aristegui et al., 1997; Aristegui and Montero, 2005). Counter-paired cyclonic and anticyclonic eddies act therefore as a two-way biological pump, increasing plankton production in cyclonic eddies and accelerating the vertical transport of organic matter out of the photic zone within anticyclonic eddies (Aristegui and Montero, 2005).

Anticyclonic eddies associated with Gran Canaria are revealed by chlorophyll-like pigments (CZCS) images, thanks to their interaction with upwelling filaments stretching from the African coast (Hernández-Guerra et al., 1993; Pacheco and Hernández-Guerra, 1999). Upwelling filaments export water with relatively low temperature and high chlorophyll content that is entrained by the anticyclonic eddies drawing its contours. Therefore, anticyclonic eddies may play a major role on the export of high-chlorophyll upwelling waters into the interior oligotrophic ocean. Satellite images and in situ data also indicate that high-chlorophyll water, resulting from island stirring or local upwelling in the western flanks of the islands, is incorporated into the cyclonic eddies during their early stages and subsequently advected downstream by the Canary Current (Hernández-Guerra et al., 1993; Aristegui et al., 1997; Pacheco and Hernández-Guerra, 1999).

Wind/current shear at the flanks of the islands may enhance plankton metabolism through different effects. Microplankton and zooplankton respiratory activity is enhanced at coastal sites, where velocity shear is maximum, presumably due to an increase in turbulence (Hernández-León, 1988, 1991; Aristegui and Montero, 2005). Ekman pumping on the wind shear boundaries of the islands produces convergence and divergence fronts which affect plankton distribution and productivity. Divergence fronts at the lee of the island induce upwelling of thermocline water, leading to an increase in primary production (Aristegui et al., 1989; Basterretxea et al., 2002). At these fronts, zooplankton is accumulated either because of the availability of large phytoplankton cells (Aristegui et al., 1989) or through passive accumulation of drifting zooplankton (Hernández-León, 1988, 1991; Hernández-León et al., 2001a).

7. Conclusions

The surface and upper-thermocline waters of the Canary Basin exemplify a southward flowing eastern boundary system, exhibiting features common to these systems but also displaying several quite specific characteristics. The dominating open ocean current is the all year long southward flowing Canary Current, the easternmost extension of the Azores Current. Recent results have indicated that the intensity and location of the Canary Current are intimately linked to the coastal upwelling system off northwest Africa, which is the true boundary condition for the interior subtropical gyre. A striking feature is that the Canary Current flows south in a narrow band, typically no wider than 200 km, linked to the coastal upwelling jet that develops north of Cape Ghir. It is not yet clear how this coastal current accommodates about 2 Sv of surface and upper-thermocline waters inflowing from the open ocean. A likely hypothesis is that this flux is shared between the baroclinic jet, of relatively fast response that increases to a maximum during summer (the maximum upwelling season), and the barotropic jet, of great inertia that keeps increasing until late fall.

The baroclinic jet is a relatively shallow feature with a width of the order of the internal Rossby radius of deformation, about 100 km in the upwelling region. This feature is present all year long but intensifies in summer, during the maximum upwelling season, distinguishable in SST images as a frontal region (even in the mean SST field, Hernández-Guerra and Nykjaer, 1997). Coastal filaments, mainly generated through baroclinic instability, are largest
and easiest to identify during this season. These filaments export water and coastal characteristics into the adjacent upwelling transition zone but it is likely that a large fraction recirculate towards the coast. The barotropic jet, on the other hand, probably reaches its maximum intensity in late fall and becomes unstable because of the inability of such a current flowing south to adapt to the decreasing planetary vorticity. At this time the Canary Current separates from the coast and flows south through the Canary Archipelago, showing a meander-like large-scale offshore diversion. When this happens the poleward undercurrent reaches the surface and northward flow may be observed between the Canary Archipelago and the African coast.

These features cause the upwelling transition zone in the Canary Basin to be characterised by two cells transporting subsurface upwelled waters into the open ocean. The first one is the standard vertical cell, present in all upwelling systems, with Ekman offshore transport responding to the wind intermittent (synoptic and seasonal) variations. The second one is the horizontal circulation cell that has its origin on open waters impinging into the coastal region north of Cape Ghir and is closed by water exported through several filaments and the seasonal flow diversion at Cape Ghir. Most of the year, the current system has meridional continuity—thanks to the coastal upwelling system—and transports large amounts of nutrients (order 10 kmol s$^{-1}$ of nitrate and proportional quantities of other nutrients) that sustain the fisheries in the region (independent on the intermittency of upwelling winds). The horizontal cell that detaches at Cape Ghir, however, is a very peculiar seasonal characteristic of the upwelling region in the Canary Basin and provides a diversion of the nutrient transport into the open ocean. The joint action of both cells causes this coastal upwelling region to be characterised by relatively low productivity values, as compared with other upwelling regions, and by a substantial export of nutrients and organic matter into the open ocean.

South of the Canary Archipelago, the Canary Current is greatly perturbed because of vortices generated at the islands and their interaction with a number of coastal upwelling filaments. In contrast with the Cape Ghir filament, the position of these filaments is variable, which gives rise to the hypothesis that their origin is linked to the intensity of the southward flowing coastal current and their interaction with the eddies. The detaching vortices are a rather recurrent feature that appears to be generated topographically by the impinging current, although the northeasterly winds may help their development. Vortices also are temporally coherent structures, which have been followed for a time period over 6 months while drifting some 500 km, so they must have a quite important impact on the distribution of properties in the whole eastern boundary system. Cyclonic (cold core) vortices are characterised by the central raising of the isotherms, and actual radial convergence at least during the early stages of their development, while the opposite situation is true for anticyclonic (warm core) vortices. Vortices also appear to undergo intermittent diapycnal mixing, its overall effect being the raising of nutrients to the surface layers, which likely helps to maintain some level of productivity at more mature stages of their development.

From a biological point of view, the Canary region behaves as a carbon source, both increasing the concentration of CO$_2$ in surface waters and exporting organic carbon to the oceanic domain. Compiled gross primary production (Pg) and community respiration (R) data from the region (Duarte et al., 2001) indicate that R is several times higher than Pg most of the year. This metabolic imbalance can be only explained by allochthonous inputs of organic matter from nearby regions. Coastal upwelling filaments, with high-chlorophyll concentration, are frequently observed extending up to several hundred kilometers offshore from the northwest Africa coastal upwelling system. Eventually, island eddies may interact with the offshore limit of some of these filaments, entraining water with organic matter and transporting it, through a complex eddy field, to the oligotrophic open ocean. Although some of the filament water may recirculate back to the coastal jet, the regular analysis of SST and surface chlorophyll from satellite images suggests that upwelling filaments spread organic matter in a recurrent fashion into the Canary region. A major seasonal feature is the diversion of water from the coastal upwelling region into the open ocean in the Canary Archipelago region, during late fall and early
winter. Such diversion would be responsible for the observations of high content of organic carbon during this season in the open ocean, much larger than actual production values (Aristegui et al., 1997). Despite the high respiration rates measured in the upwelling region, some of the organic matter is indeed transported to the oligotrophic waters of the subtropical gyre. Relatively high dissolved organic carbon values have been measured throughout the whole transition zone of the Canary region, south of the archipelago, indicating that this semi-refractory carbon is available to be advected out of the region while it is slowly oxidized.

An additional source of organic carbon is due to cyclonic eddies, which increase primary production several times with respect to ambient waters, as a response to nutrient pumping. Barton et al. (1998) estimated that island eddies may contribute to the nitrogen flux to the Canary region as much as coastal upwelling. These eddies would have an antagonist effect on the carbon fluxes in the region, when cold, nutrient-rich and CO₂-rich waters upwell at their cores. The surface warming of the recently upwelled water would produce an increase in surface pCO₂. Conversely, the biological fixation of CO₂ by phytoplankton would drop the CO₂ surface concentrations. Overall, however, the net balance is governed by the physical processes, and the cyclonic eddies behave as net sources of CO₂ to the atmosphere. Finally, the fresh organic matter produced by eddies may either be partly oxidized in short times (increasing R in surface waters of the region), exported to the oligotrophic subtropical gyre by the eddy field (increasing R in the open ocean), or transported downwards to the deep ocean, where carbon would be sequestered for long periods. One of the main goals for future research must be to elucidate which of the above carbon paths prevails in this region.

Acknowledgements

This work has been supported through the European Union (project CANIGO, number MAS3-CT96-0060) and the Spanish government (project COCA, number REN2000-U471-C02-02-MAR). Part of this work was written while the first author was at University of Wisconsin-Madison with funding from the Secretaria de Estado de Educacion y Universidades of the Spanish Government. The satellite images used in this study were received and processed at the Universidad de Las Palmas de Gran Canaria. We are grateful to Susan Lozier and Pierro Zanasca for providing the data to produce Figs. 6c, d and 7. We also wish to thank our reviewers for a number of useful comments and suggestions, in particular one of them for his exhaustive revision of the original manuscript that greatly helped to improve it.

References


