Slope transport of suspended particulate matter on the Aquitanian margin of the Bay of Biscay

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Abstract

Spatial and temporal characteristics of the water masses and the dispersion of the suspended particulate matter were investigated using current meter, hydrographic and nephelometric observations, gathered during the ECOFER experiment (1989–1991) in the Cap-Ferret Canyon on the Aquitanian margin of the Bay of Biscay. While characteristics of the deep water masses were stable from one year to another, large hydrographic change in the upper 500 m related to winter renewal induced by poleward advection of warm and saline water along the continental slope. The slope circulation and seasonal eddy activity appear as important dynamical mechanisms that control the entrainment and the dispersion of the suspended particulate matter from the neritic domain to the deep ocean. A predominantly northward along-slope current with occasional reversal characterizes this circulation. The nephelometric structures also showed seasonal changes in the overall suspended particulate matter content, but recurrent features, such as the presence of intermediate nepheloid layers at the shelf-break depth and various depths along the slope (~500, 1000 and 2000 m), were observed. These nepheloid layers extended off the slope to about 10–30 km, but especially laterally along the slope. Their presence indicated that suspended particulate matter exchanges between the shelf and the slope occurred mainly in the head of the canyon and along the southern open slope. The intermediate nepheloid layers around 500 m depth detached from the slope particularly in regions where the bottom slope is close to critical for the M2 internal tide. © 1999 Elsevier Science Ltd. All rights reserved.
1. Introduction

1.1. Sources and transfers of terrigenous material on the Aquitanian margin

Rivers and coastal runoff discharge large quantities of suspended particulate matter (SPM) onto the continental shelf of the Bay of Biscay (Fig. 1A). The Gironde River, with an annual solid discharge of $0.5 - 1.5 \times 10^6$ tons, supplies about 70% of these terrigenous inputs (Castaing and Jouanneau, 1987). SPM expelled from the Gironde estuary in the surface nepheloid layer is dispersed mainly along the coast. The NE–SW orientation of the mud-patches west of the estuary (Jouanneau et al., 1999; Ruch et al., 1993) indicates that the SPM in the bottom nepheloid layer is transported seaward. From estimates of long term accumulation rates (100–1000 yr), Lesueur et al. (1989) estimated that about two thirds of the suspended sediment delivered from the Gironde estuary is stored on the shelf within several mud patches, the remainder being available for export to the slope. Whereas the bottom nepheloid layer (BNL) on the shelf during low river discharge period extends only over the inner shelf, the extension of the BNL in flood period reaches the edge of the shelf. The Cap-Ferret Canyon (Fig. 1B), located southwest of the Gironde estuary, appears to be a probable pathway for material that escapes the shelf and supplies the deep basin of the Bay of Biscay. Ruch et al. (1993) estimated from the distribution of the SPM concentration between the Gironde estuary and the Cap-Ferret Canyon, that 2–3% of the terrigenous SPM flowing from the Gironde estuary during floods is transferred through the Cap-Ferret Canyon.

1.2. Influence of hydrodynamics

Hydrodynamic processes are predominant factors controlling the sediment distribution and the shelf-slope exchanges of suspended particles. Castaing et al. (1999) have shown the winter presence of a mid-shelf thermo-haline front. They suggest that the position of the mid-shelf mud-patches (Fig. 1A) was linked to the long-term stable location of this shelf front. Besides, Castaing et al. (1982) also showed a seaward transfer of detrital quartz grains with an important resuspension on the outer shelf under the combined influence of swell, internal tides and currents. The amplification of mixing processes on the outer shelf was also shown by Ruch et al. (1993), with the progressive thickening of the bottom mixed layer and bottom nepheloid layer from the mid-shelf toward the shelfbreak.

The intensification on critical slope of the internal tides on the continental slope appears to be a major dynamic mechanism controlling the local remobilization of sediment. For critical angles, the incident tidal waves yield to large bottom currents likely to erode the sediment. Dickson and McCave (1986), Thorpe and White (1988) established that the resuspension of sediment at depths of about 600 and 2550 m, respectively, was linked to the critical reflection of the dominant $M_2$ tidal waves on the continental slope of the Porcupine Bank. Thorpe and White (1988) also showed that the resulting deep turbid layer extended laterally along the slope under the influence of a deep and predominantly poleward along-slope current.
Fig. 1. (A) Study area (rectangle) of the Cap-Ferret Canyon on the Aquitanian margin of the Bay of Biscay. The hatched regions delineated the “mud-patches” on the continental shelf (adapted from Jouanneau et al., 1999). (B) Bathymetry of the Cap-Ferret Canyon. Isobaths are labeled in meters. Two sections presented later on are delineated by thick solid lines along the axis of the canyon (A-A') and across the canyon (B-B'). The solid triangles along the axis of the canyon represent the location of the current meter and sediment trap moorings.
Fig. 2. Sketch of surface and slope currents patterns within the Bay of Biscay. (A) Situation during summer and during period of weak flow of the slope current. (B) Winter situation during strong poleward intrusion of the slope current.

This intermittent poleward slope current, whose main direction is opposite to the anticyclonic circulation in the deep basin, is part of the general circulation in the Bay of Biscay (Fig. 2). Le Cann and Pingree (1995) as well as Koutsikopoulos and Le Cann (1996) described its structure and variability. Strongly constrained by the bathymetry, this current derives from forcings associated with the density field and the wind. Thorpe (1987) and Frouin et al. (1990) examined the dynamics of the slope current along the Porcupine Bank and the Portuguese coasts and concluded that the motion was principally in geostrophic balance. The annual cycle of the slope current is attributed to the seasonal distribution of the wind off the Iberian peninsula. In the southern part of the Bay of Biscay, the summer period is characterized by upwellings and an equatorwards slope current; in winter the weakening of the southward component of the wind stress leads to a development of the thermohaline poleward slope current. From multi-annual current meter observations, Pingree and Le Cann (1990) measured a poleward residual velocity of a few centimeters per second and an increasing transport to the north (i.e., about \(3.5 \times 10^6\) m\(^3\) s\(^{-1}\)) along the Celtic and
Armorican slopes). The local variation of the surface slope current is rapid, with occasional reversals of its direction related to the tide and the wind (Le Floch, 1969, 1970; Thorpe, 1987; Pingree and Le Cann, 1990). Furthermore, Pingree and Le Cann (1992) observed, near the Cap-Ferret Canyon, the formation of eddies that detached from the slope current. The eddies resulted from perturbations of the slope current caused by the abrupt change of the continental slope’s direction at the latitude of the Cap-Ferret Canyon. These eddies migrated eastward with a velocity of a few centimeters per second entraining the surface ( < 500 m) slope water seaward. The occurrence of these eddies varies from one year to the next depending on the intensity of the slope current. Pingree and Le Cann (1992) and Pingree (1994) showed that the formation of two eddies during the winter of 1989/1990 coincided with a particularly intense poleward transport of surface warm water along the Atlantic coast of the Iberian peninsula. They referred to this episodic intensification of the surface slope current as the Navidad current.

1.3. Study of the transfers of suspended particulate matter on the continental slope

The ECOFER experiment is part of the ECOMARGE (ECOsysstèmes de MARGE continentale) program, which studies the role of continental margins in the cycle of elements (C, N, Si) and estimates the particle flux and energy transfer between continental shelves and deep basins (Monaco et al., 1990). One of the goals of the ECOFER experiment is to study these shelf-slope exchanges on the Bay of Biscay’s continental margin and, in particular, the effective role of the Cap-Ferret Canyon as a conduit of terrigenous material to the deep sea. It included four consecutive deployments, between 1989 and 1991, of two moorings equipped with sediment traps, current meters and transmissometers. Five cruises were conducted to study, on a seasonal basis, the main hydrological and nephelometric features of this margin as well as the major characteristics of bottom sediment and biota. In the framework of this experiment, the present study describes the results of examination of hydrographic, nephelometric and current meter observations gathered in the Cap-Ferret Canyon (Fig. 3). This study focuses mainly on: (1) the description of the slope water masses and the distribution of the suspended particles; (2) their temporal (seasonal) variations and (3) the influence of the tidal waves and the along-slope circulation on the shelf-slope transfers of SPM.

Fig. 3. Timing of the deployment of the different moorings and the CTD cruises. Mooring deployments are numbered with Roman numbers, while cruises are numbered with Arabic numbers.
2. Data and analysis techniques

2.1. Hydrographic, nephelometric and current meter data

The hydrographic and nephelometric data examined here are observations from five cruises conducted during the period between June 1989 and August 1991 (Fig. 3). Additional current meter observations were obtained during the ECOFER II, III and IV deployments (Fig. 3) are also presented.

The hydrographic data of the ECOFER 1, 2 and 5 cruises, carried out by the Group LIMNOCEANE of the University of Neuchâtel, were obtained with a Neil Brown Mark III CTD. Those of the ECOFER 3 and 4 cruises, collected by the Department of Geology and Oceanography of the University of Bordeaux, were obtained with a Seabird SBE9 CTD. A systematic and constant shift of the salinity measurements, about 0.045 psu, was observed for the ECOFER 4 stations. The lack of bottle samples for salinity measurements did not allow direct correction of the CTD salinity. Since the Neil Brown Mark III CTD was calibrated before each cruise and the consistency of the deep $\theta$-$S$ curve of the ECOFER 1, 2, 3 and 5 cruises, the ECOFER 4 shift very likely relates to a conductivity offsetting problem. This instrumental offset was corrected to ensure a deep $\theta$-$S$ ensure relationship with the other cruises. Dissolved oxygen data ($O_2$ electrode on CTD) were obtained during the ECOFER 3 cruise (Fall 1990).

Nephelometric observations during the ECOFER 1, 2 and 5 cruises were made using a forward light scattering meter (Vangriesheim et al., 1992) attached to the CTD probe. A Seatech transmissometer with a 25 cm beam path length was used during the ECOFER 3 and 4 cruises. The nephelometer was calibrated versus a standard Formazine solution, and the scattered light intensity expressed as equivalent Formazine Turbidity Units (FTU) (Nyfeler and Godet, 1986). Calibration of turbidity with SPM or particulate organic carbon (POC) concentrations showed that the turbidity was relatively well correlated to both parameters (Durrieu de Madron, 1995). These comparisons produced linear relationships: SPM Conc. ($\mu$g/l) = $0.27 \times$ Turbidity (mFTU) + 329 with a correlation coefficient of 0.52, and POC Conc. ($\mu$g/l) = $0.119 \times$ Turbidity (mFTU) + 75 with a correlation coefficient of 0.68.

2.2. Sampling strategy

The Cap-Ferret Canyon runs E–W across the Aquitanian continental slope (Fig. 1B). The general orientation of the continental slope south of the canyon is N–S, and north of the canyon it is oriented NW–SE. The canyon is about 30 km wide at the shelfbreak depth and 2500 m deep at the bottom of the slope. The coverage of the hydrographic cruises consisted of a dense grid of stations in and around the canyon to gain a general picture of the distribution of the water masses and the dispersion of the SPM. Each cast extended to within a few meters of the bottom. The hydrographic coverage was largely restricted to the axis of the canyon during the ECOFER 3 and to the surface layer (< 250 m) during the ECOFER 4 cruise. Therefore, the hydrographic and nephelometric spatial structures will mainly be illustrated by results from the ECOFER 1, 2 and 5 cruises.
The moorings were deployed in the middle part (site MS1) and the lower part (site MS2) of the canyon (Fig. 4). The MS1 mooring was equipped with PPS3 sediment traps together with Aanderaa RCM 5, 7 or 8 current meters, at four levels (380, 1370, 1900 and 2250 m). An additional current meter was localized at 750 m deep. The MS2 mooring was equipped with two sediment traps coupled with current meters, at 1900 and 2250 m, respectively. Current meters recorded pressure, temperature, speed and direction at 30 min intervals. The depth of the instruments was calculated from the pressure data. During the three deployment periods, the total length of record for each current meter ranged between 152 and 447 days.

3. Results

3.1. Hydrography

The character of the water masses in the Cap-Ferret Canyon is illustrated in the $\theta$–$S$ diagrams shown in Fig. 5. The density structures along and across the axis of the canyon are displayed in Figs. 9–11. Potential temperature ($\theta$) will hereafter be referred to as temperature.

3.1.1. Characterization of the water masses

Four bodies of water were differentiated for each cruise: (1) A low salinity water mass spanned depth between 200 and 600 m. The salinity minimum occurred at $\theta \sim 10.8^\circ$C, $S \sim 35.51$–$35.54$ psu, $\sigma_\theta \sim 27.22$ and core depth of $\sim 500$ m. This water mass characterized a more narrow body than that included in the original definition of North Atlantic Central Water (Sverdrup et al., 1942). Following the nomenclature of Harvey (1982), we refer to this water mass as Eastern North Atlantic Water (ENAW). According to Tsuchiya et al. (1992) it is possible that, besides its origin in the Labrador current region, this low-salinity water probably contains a significant amount of Antarctic Intermediate Water (AAIW) transported to the Northeast Atlantic by the Gulf Stream-North Atlantic Current.

(2) The high saline core of Mediterranean Overflow Water (MOW) was clearly seen between 700 and 1300 m deep. The MOW that flows northward along the European coast is characterized in the eastern Bay of Biscay by a temperature and salinity maximum ($\theta \sim 9.8^\circ$C, $S \sim 35.73$–$35.76$ psu, $\sigma_\theta \sim 27.6$) and an oxygen minimum at a depth of $\sim 1000$ m.

(3) Evidence of Labrador Sea Water (LSW) was observed with an oxygen maximum at $\sigma_2 \sim 36.90$ and a depth of $\sim 1900$ m. This water was also generally characterized by a salinity minimum. Due to the mixing with the overlying saline MOW, this salinity minimum lessens as the water migrates to the East and eventually disappears into the Bay of Biscay (Byun, 1980).

(4) The characterization of the bottom deep water (BW in Fig. 5A), underlying the LSW, suggests influence of both Northeast Atlantic Deep Water (NEADW) and Antarctic Bottom Water (AABW). Beneath the LSW, the relative salinity maximum characteristic of the NEADW (Tsuchiya et al., 1992) was not observed. However, this
Fig. 4. (A) Design and localization of the moorings at site 1 (upper canyon) and site 2 (lower canyon). Four and two sediment traps-current meters’ pairs were moored at mooring sites 1 (MS1) and 2 (MS2) respectively. (B) Typical temperature, salinity and turbidity profiles at MS1 and MS2.
Fig. 5. $\theta$–$S$ diagrams for all stations of each cruise (A) and expanded diagram for the upper part of the water column (B): (ENAW) eastern North Atlantic Water; (MOW): Mediterranean Overflow Water; (LSW) Labrador Sea Water; and (BW) Bottom Waters. The asterisk indicates the $\theta$–$S$ core characteristics of the eddy F90A described by Pingree and Le Cann (1992).
salinity maximum is also likely to vanish in the eastern Bay of Biscay as a result of the MOW influence in increasing the salinity and suppressing the salinity minimum of the LSW. The similarity of the $\theta$–S characteristics of the NEADW ($\theta \sim 2.9^\circ$C, $S \sim 34.95$ psu, $\sigma_2 \sim 37$ at a depth of $\sim 2700$ m) west of the Bay of Biscay (Tsuchiya et al., 1992) to that of the water observed at depths of 2500–2600 m in the Cap-Ferret Canyon supports the influence of NEADW in the region. Moreover, the bottom-most water observed in the Cap-Ferret Canyon ($\theta \sim 2.5^\circ$C, $S \sim 34.92$ psu, $\sigma_2 \sim 37.03$) at a depth of $\sim 3000$ m and the increase in silicate concentration of the deepest water (Valencia, personal communication) indicate the possible influence of AABW of southern origin.

3.1.2. Temporal variation of the surface waters

Whereas general characteristics of the hydrography for depths greater than 500 m ($\sigma_0 > 27.22$) were similar for all the cruises (Fig. 5A), important annual variations occurred in the overlying layer (Fig. 5B). Salinity and temperature shallower than 200 m were higher in spring and fall periods of 1990 than in spring and summer periods of 1989 and 1991. At a depth of 100 m, the increase in temperature and salinity ranged between 0.5–0.8°C and 0.06–0.13 psu, respectively. Thus during 1990 there was evidence of warmer and saltier water advection. This advection of warm and salty water most likely resulted from an enhancement of the poleward winter transport of surface water along the continental slope off Portugal and Spain (Frouin et al., 1990; Haynes and Barton, 1990). During the winter of 1989–1990, an exceptional warm flow was observed in the southern Bay of Biscay, with a large northward extension along the French continental slope (Pingree and Le Cann, 1992; Pingree, 1994). The presence of abnormally warm and salty surface water observed in May and October 1990 in the Cap-Ferret Canyon indicated that the signal of this intrusion of water remained there for almost one year. The surface water for 1991 was colder and denser than that of the two previous years. The similarity of the $\theta$–S characteristics between spring and summer 1991 also indicated a persistence of this new surface water signal over at least four months. Thus, the inter-annual variability of the surface water characteristics probably related to a winter renewal, which in turn seems primarily modulated by the northward spreading of the Navidad current.

3.2. Flow patterns in the canyon

Direct current measurements made in the axis of the canyon demonstrated the major patterns of the flow in the canyon (Fig. 6). The currents were predominantly oriented along the slope bathymetry and periodic reversals for periods about 1 month occurred. Filtered velocities, where tidal and inertial signals were removed using Thompson’s (1983) ten days filter, indicated a decrease of the current speed with depth. In the middle canyon (Site MS1), maximum N–S filtered velocities at 380 m depth reached 25 cm s$^{-1}$. They were as high as 20 and 10 cm s$^{-1}$ at 750 and 1370 m depth, respectively. The deep and near bottom currents at 1900 and 2250 m depth were predominantly oriented along the canyon’s axis, with a weak upslope net flow, and a N–S component lower than 3 cm s$^{-1}$. In the lower canyon (Site MS2), N–S
filtered velocities at 1900 and 2950 m deep were generally less than 6 cm s\(^{-1}\). At both mooring sites, the net flows inferred from the three deployment periods indicated a systematic poleward flow with residual northward velocities ranging between 0.2 and 1.4 cm s\(^{-1}\). This net flow is compatible with the residual along-slope currents described by Pingree and Le Cann (1989).

The occasional opposition of the current direction at 380 and 750 m depth (Fig. 6) revealed a strong baroclinic component. The distribution of the surface geopotential anomalies relative to 2000 dbar indicated an along-slope current flowing northwardly during the ECOFER 2 (May 1990) (Fig. 7) and southwardly during the ECOFER 1 (June 1989) and 5 (August 1991) cruises. During the ECOFER 2 cruise, the mooring lines were deployed before the hydrographic survey. The direction of the currents measured at 380 and 750 m depth during the period of the cruise was consistent with the direction of the baroclinic flow.

Spectral analysis of the half-hourly data (Fig. 8) showed that the semi-diurnal tidal frequencies (M\(_2\) and to a lesser extent S\(_2\)) are the major tidal components. Smaller peaks of energy appear at the inertial frequency at 380 m depth and at the quarter-diurnal tide (M\(_4\)) for deep near-bottom currents. The comparison of the current’s spectral density at 1900 m deep and near the bottom, within (site MS1) and outside (site MS2) the head of the canyon, shows a redistribution of the energy of the currents at the predominant semi-diurnal tidal frequency (M\(_2\)). The semi-diurnal flow variability was predominantly associated with the N–S flow component in the lower canyon, but was almost completely linked to the E–W flow component in the head of the canyon. This redistribution showed a channelling of the deep tidal currents along the axis of the canyon. The coherence results between the bottommost instruments at MS1 and MS2 indicated a high coherence with a phase lag about 140\(^\circ\) in the semi-diurnal E–W currents. This coherence was again indicated when comparing E–W currents at 1900 m between MS1 and MS2 but with a smaller phase lag (40\(^\circ\)). Vertical coherence between E–W currents from instruments at 1900 m deep and near the bottom on the same mooring also indicated a phase lag at both sites. Thus, the phase lags in the horizontal and the vertical between MS1 and MS2 suggested that the semi-diurnal tide propagated upcanyon. The decrease of the bottommost current’s energy toward the canyon head furthermore indicated no upcanyon intensification of the tide between the two mooring sites.

3.3. Spatial distribution of the nephelometric structures

The vertical distribution of the nephelometric structures is depicted in the sections along and across the axis of the canyon (Figs. 9–11). The horizontal dispersion of the SPM in the intermediate nepheloid layer around 500 m deep is displayed in Fig. 12.

3.3.1. Recurrent features

The overall patterns of the nephelometric structures were rather similar for all the cruises. Below the surface nepheloid layer (SNL) that covers evenly the shelf and slope region, the main intermediate nepheloid layers (INLs) detached from the continental slope at the shelf-break depth, and around a depth of 500 m. The offshore extension
Fig. 6. Stick plots of the daily average currents and current ellipses at mooring sites 1 (canyon head) and 2 (lower canyon) for the ECOFER II, III and IV deployments. The distance of the deep current meters to the bottom is indicated in meter above bottom (mab). See Figs. 1 and 4 for mooring’s location and design. While the nominal depth of the sub-surface currentmeters at MS1 was 380 and 750 m, pressure sensors indicated depths of 280 and 610 m during the first deployment (ECOFER II).
and the intensity of the turbid structures in the canyon head decreased with greater depths down to about 1200 m. This later represents the approximate depth of the clear water minimum near the slope. Offshore, the clearest waters ( ~ 20–30 mFTU) were observed at depths between 700 and 1800 m, within the Mediterranean and Labrador
Sea waters. The slight increase in turbidity between the clear water minimum and the bottom defines a deep nepheloid layer around 2000 m depth, which was observed throughout in the canyon (Figs. 4–11). Horizontally, these different INLs have different expanse:

1. **INL at the shelfbreak depth.** The turbid bottom mixed layer on the shelf, which is about 50 m thick, indicated a transport of SPM by the bottom currents. The shelf bottom nepheloid layer (BNL) became detached at the shelf break depth (~ 200 m) to form an INL. This shelfbreak depth INL appeared mostly in the canyon’s head. Its off-shelf extension was small (≤ 10 km).

2. **INLs around 500 m depth.** A mid-slope INL centered at depth of around 500 m was observed during all the cruises. This INL was located either on the flanks of the canyon and in the head of the canyon. It extended along the slope rather than across the slope with great continuity (Fig. 12). The entrainment and the subsequent dilution of the SPM entering the vein of the slope current resulted in a rapid off-slope decrease of the turbidity. The off-slope extension of the INL’s varied between 10 and 30 km.

3. **Deep nepheloid layer around 2000 m deep.** A BNL, up to 500 m thick, appeared at the base of the canyon head, below the wedge of clear water that penetrated the canyon (e.g., station at MS1 in Fig. 4). This BNL indicated resuspension in the canyon head. It detached from the bottom around 2300 m deep, and formed a deep INL extending seaward along the axis of the canyon (station at MS2 in Fig. 4). The currents at 1900 m depth and near the bottom in the axis of the canyon indicated a channeling with a predominant upcanyon net flow. These measurements suggested that no net down canyon transport of SPM induced by the mean currents occurred in the deep canyon down to 50 mab. However, they did not prejudge the net transport direction closer to the bottom, that is believed to occur downcanyon as suggested by sedimentological results (Cremer et al., 1999).

### 3.3.2. Seasonal variations of SPM concentration

The intensity of the nephelometric measurements was rather similar between late spring 1990 (ECOFER 2) and early summer 1989 (ECOFER 1). Nevertheless waters between 200 and 1000 m depth showed higher turbidity values (~25%) in early summer (ECOFER 1) than in late spring (ECOFER 2). The overall turbidity values for these later cruises were about two times higher than those observed in late summer 1991 (ECOFER 5). The decrease of the SPM concentration in the upper 200 m during late summer was probably a consequence of a reduced primary productivity in the surface (euphotic) layer in comparison with the planktonic bloom period that occurs...
Fig. 9. Profiles of turbidity (mFTU) and potential density anomaly (kg m\(^{-3}\)) along and across the axis of the Cap-Ferret Canyon during the ECOFER 1 cruise (June 1989). Potential density anomaly is referenced to 0 and 2000 dbar (below 1000 m). Sections are designated as transect A-A' and B-B' in Fig. 1. The crossing point of the two transects is indicated by a cross. The inverted triangles indicate the location of the mooring sites.
Fig. 10. Profiles of turbidity (mFTU) and potential density anomaly (kg m$^{-3}$) along and across the axis of the Cap-Ferret Canyon during the ECOFER 2 cruise (May 1990). Potential density anomaly is referenced to 0 and 2000 dbar (below 1000 m). Sections are designated as transect A-A’ and B-B’ in Fig. 1. The crossing point of the two transects is indicated by a cross. The inverted triangles indicate the location of the mooring sites.
Fig. 11. Profiles of turbidity (mFTU) and potential density anomaly (kg m\(^{-3}\)) along and across the axis of the Cap-Ferret Canyon during the ECOFER 5 cruise (August 1991). Potential density anomaly is referenced to 0 and 2000 dbar (below 1000 m). Sections are designated as transect A-A’ and B-B’ in Fig. 1. The crossing point of the two transects is indicated by a cross. The inverted triangles indicate the location of the mooring sites.
in spring and early summer (Heussner et al., 1999). Besides, the decrease of the SPM concentration below 200 m probably related to reduced terrigenous inputs from the shelf as a result of low river discharge in summer (Durrieu de Madron et al., 1993; Froidefond et al., 1999).

4. Discussion

We discuss here the influence of two dynamical processes on the transfer of SPM on the continental slope, namely the internal tides and the along-slope current.

4.1. Influence of the internal tides

The detachment of INLs from the slope suggests (1) a localized origin of sediment laden layers mainly on the open slope and at the base of the canyon head, and (2) the occurrence of bottom currents sufficient to carry recent sediments into suspension. Lead-210 analysis of sediment traps samples indicated the influence of two sources of particulate matter, localized in the surface layer and on the upper slope (Radakovitch and Heussner, 1999). Sediment trap data also showed a downward increase of the mean annual mass flux within the head of the canyon. These trends were representative of all sampling periods (between 16 and 27 days), except during winter 1990/1991 where the mass flux profiles showed occasional mid-depth (trap at 1350 m deep)
minimum or maximum. Heussner et al. (1999) interpreted these results as indicative of a lateral input of particles at depths between 380 and 1350 m.

The current meters were located too far from the bottom to properly study current’s potential to resuspend recent sediments. Nevertheless, the current observations at 380 and 750 m deep in the head of the canyon showed current up to 37 cm s\(^{-1}\) and about 2% of the velocity measurements higher than 20 cm s\(^{-1}\). Only one current meter record at 380 m deep showed stronger currents, with 20% of the measurements higher than 20 cm s\(^{-1}\), during the May–October 1990 period. Except for the latter period, mid-depth currents were probably insufficient to resuspend recent sediments on the upper slope. Similarly, speeds recorded by the bottommost current meters (50 mab) in the canyon’s head (MS1) suggested that current-induced resuspension was not likely to occur. Speeds never exceeded 14 cm s\(^{-1}\) and average velocities were less than 4 cm s\(^{-1}\).

Several studies indicated the occurrence of sediment resuspension along the slope down to 2500 m deep at critical levels for the internal tides (Cacchione and Drake, 1986; Dickson and McCave, 1986; Thorpe and White, 1988; Gardner, 1989a). Experimental studies (De Silva et al., 1997) further indicate that mixing along a sloping bottom by waves breaking could be an important means of isopycnic transfer of matter off the slope. Current measurements in the Cap-Ferret Canyon clearly indicate that the semi-diurnal tide (M\(_2\)) is the dominant frequency (Fig. 8). It is possible from the hydrological data to determine if the influence of this wave is conceivable.

The dispersion equation for internal waves in a stratified rotating fluid is:

\[
\sigma^2 = N^2 \sin^2 \theta + f^2 \cos^2 \theta, \\
\]

where \(\sigma\) is the wave frequency, \(\theta\) is the angle of the wave vector with respect to the horizontal, \(N\) the local mean buoyancy frequency, and \(f\) the local inertial frequency (\(\sim 1.03 \times 10^{-4} \text{ s}^{-1}\)). An on-shore going internal wave can be reflected forward (shoreward) or backward (seaward) by the continental slope if the angle of the incident wave is larger or, respectively, smaller than the angle of the seabed, \(\alpha\) (also measured with respect to the horizontal). The critical reflection (propagation of the group velocity of the wave along the sloping bottom) is obtained when the angle of the incident internal wave matches the angle of the seabed. The frequency of the M\(_2\) tide is \(\sigma_{M2} = 1.405 \times 10^{-4} \text{ s}^{-1}\). Therefore, the regions of the continental slope likely to induce a critical reflection of the semi-diurnal internal tide can be estimated to have the following sea bed angle:

\[
\alpha = \arctan \left[ \left( \frac{\sigma_{M2}^2 - f^2}{N^2 - \sigma_{M2}^2} \right)^{1/2} \right].
\]

Mean values of \(N\) for all the cruises range from 1.7 to 3.5 \(\times 10^{-3} \text{ s}^{-1}\), and give values for \(\theta\) which range from 3.2 to 1.5°, respectively. The bottom angle of the upper canyon head, between 10 and 15°, is generally steeper than the characteristic angle (\(\theta\)) of the M\(_2\) tidal wave (Fig. 13). The lower part of the canyon, the base of the slope around 2000 m deep in the canyon head, and parts of the southern open slope are less steep (between 1.5 and 3.2°). Locally, the angle of the slope (\(\alpha\)) approached and even
matched that of the $M_2$ tidal wave. The base of the slope in the canyon head and the southern open slope represented thus potential sites of sediment resuspension. These later results together with the nephelometric and sediment trap measurements clearly support that resuspension by internal tides and subsequent transport of sediment occurred on the southern open slope. Resuspension at the base of the canyon head, although suggested by nephelometric measurements and the potential role of tidal waves, remains hypothetical.

The study of Gardner (1989a) emphasized the relationship between resuspension and internal tides whose characteristic angle matched the bottom slope of the canyon along with the funneling and amplification of the tide between the canyon walls. In the Cap-Ferret Canyon, we observed instead an upcanyon decrease of the near-bottom currents between the two mooring sites. The steep walls of the canyon are indeed more favorable to reflect backward the incident tidal wave rather than to reflect it forward toward the canyon head. The morphology of the canyon is of major importance for the influence of the tidal currents on the resuspension and the transport of sediment. The Cap-Ferret Canyon is much more like the broad-head of the Quinault canyon (Hickey et al., 1986), where the up-down canyon flow was much less intense than in Baltimore and Hudson canyons that are sharp V-shaped canyons (Hotchkiss and Wunsch, 1982; Gardner, 1989a).

4.2. Influence of the along-slope circulation

Because of density stratification, the up- or down-slope displacement of water and resuspended material along the bottom is limited. Fluid in the bottom layer can then
escape along a surface of the same density that intersects the sloping bottom. The detachment and the isopycnal dispersion of the INL’s illustrate the predominant influence of horizontal advection.

The entrainment by the slope current of the sediment resuspended on the southern open slope is clearly seen with the spreading of the mid-slope INL across the canyons head. The extension of this turbid plume was limited to the upper slope in June 1989 (ECOFER 1, Fig. 9) and August 1991 (ECOFER 5, Fig. 11) while the slope current is southward. The INL extended across the entire width of canyon in May 1990 (ECOFER 2, Fig. 10) when the current was northward. An INL, detaching from the northern open slope, appeared during the ECOFER 5 cruise (August 1991) while the slope current was southward (Fig. 11). Thus the slope current appears as a major factor controlling the off-slope dispersion of SPM: it swept away the SPM escaping from the shelf and the upper slope, and transported it mostly parallel to the slope (Fig. 12).

The seasonal variability of this current is expected to induce large changes in the SPM transport: the winter intensification of the slope current may flush the shelf and upper slope waters south of the Cap-Ferret Canyon and increased the northward transport of associated SPM along the slope. This effect was confirmed by the particulate mass fluxes measured within the head of the canyon during the winter 1990/1991 (Heussner et al., 1999). During this period they observed the largest mass fluxes with a maximum at mid-depth (1350 m).

One must also consider the potential, though occasional, influence of the eddies detaching from the slope current, which by entraining the turbid surface slope water westward, intensify and prolong the seaward exportation of SPM. Between November 1990 and March 1991, the daily averaged currents at 380 m and 750 m deep within the canyon head (MS1) indicated currents rotating anticyclonically with a 10–14 days’ period (Fig. 6). These flow patterns suggested the occurrence of winter eddy activity. Pingree and Le Cann (1992) showed that the formation of eddies in the vicinity of the Cap-Ferret Canyon occurred for five years during the period 1979–1991, because of a northward propagation of warm surface water along the Spanish and French continental slopes. This process was particularly active during the winter of 1989–1990, which showed the formation of three eddies. Pingree and Le Cann (1992) estimated that one of these eddies (F90A) transported about 400 km$^3$ of warm and salty slope water (Fig. 5B) and associated SPM throughout the Bay of Biscay. Considering that the minimum turbidity of the surface slope water is about 50 mFTU, which is equivalent to a SPM concentration of 0.343 mg/l and a POC concentration of 0.081 mg/l, this eddy would have yielded an exportation of at least $0.14\times10^6$ Tons of SPM and $0.03\times10^6$ Tons of POC.

5. Conclusions

Some conclusions may be drawn from the ECOFER study of the particulate matter transfers as inferred from nephelometric, hydrologic and current measurements.

It clearly shows that advective processes play a prominent role in transferring particulate matter on the Cap-Ferret Canyon margin. Currents were directly
controlled by the canyon morphology and followed the continental slope contours. Their direction is predominantly northward, but with some reversals; over the period from May 1990 to August 1991, the currents presented a weak net flow to the North. The along-slope current provokes a confinement of the suspended particulate matter, deriving from the shelf and the upper slope, in the head of the canyon. It subsequently exports this matter along the slope by entrainment of the shelf and slope waters. The inter-annual evolution of the water masses characteristics clearly illustrates the influence of winter intensification of the along-slope current in the renewal of surface water. This seasonal process probably plays an important role on the SPM export, because it flushed the shelf water upstream of the canyon and the associated eddy activity is likely to transport part of this water offshore.

The influence of the tidal waves impinging the slope was further shown in their potential role in resuspending the sediment along the gently sloping southern open-slope, and together with the along-slope circulation in generating intermediate nepheloid layers. Within the deep canyon, the currents are generally directed cross-slope and are believed to be not strong enough to generate resuspension. Although deep turbidity structures developed along the canyon axis, no downcanyon net transport of suspended matter is evidenced from current observations at 50 m above the bottom.

These results emphasize similar controls of shelf-slope exchanges of SPM by the tidal waves and/or the along-slope current observed on various continental margins (e.g., Baker and Hickey, 1986; Gardner, 1989b; Palanques and Biscaye, 1992; Durrieu de Madron, 1994). Whereas tidal influence is strongly dependent of the tidal amplitude and the slope morphology, along-slope currents are a characteristic feature of the circulation on the oceanic boundaries and thus should be considered as one ubiquitous factor controlling the shelf/slope exchanges.

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