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# Amélioration de la prévisibilité de la dispersion du matériel en dérive en zone côtière par fusion d'informations numériques et observationnelles

Sloane Bertin

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# AMÉLIORATION DE LA PRÉVISIBILITÉ DE LA DISPERSION DU MATÉRIEL EN DÉRIVE EN ZONE CÔTIÈRE PAR FUSION D'INFORMATIONS NUMÉRIQUES ET OBSERVATIONNELLES

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Novembre 2024

Sous la direction de :  
Pr. **Alexei Sentchev**  
Dr. **Anna Rubio**





# Thèse de Doctorat

*Mention Sciences Agronomiques et Écologiques*

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**de l'Université du Littoral et de la Côte d'Opale**

par

**Sloane Bertin**

pour obtenir le grade de Docteur de l'Université du Littoral  
et de la Côte d'Opale

***Amélioration de la prévisibilité de la dispersion du matériel en  
dérive en zone côtière par fusion d'informations numériques et  
observationnelles***

Thèse dirigée par Alexei SENTCHEV et Anna RUBIO

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# Table des matières

<b>Liste des abréviations</b>	<b>5</b>
<b>Introduction générale</b>	<b>6</b>
<b>1 Zones d'étude et jeux de données</b>	<b>15</b>
1.1 Préambule . . . . .	15
1.2 Manche . . . . .	16
1.2.1 Contexte marégraphique et courantologie . . . . .	16
1.2.2 Contexte météorologique . . . . .	18
1.3 Données disponibles en Manche orientale . . . . .	19
1.3.1 Vitesses des courants mesurées par les bouées . . . . .	19
1.3.2 Vitesses modélisées . . . . .	20
1.3.3 Données de vent . . . . .	23
1.4 Golfe de Gascogne . . . . .	23
1.4.1 Contexte marégraphique et courantologie . . . . .	23
1.4.2 Contexte météorologique . . . . .	25
1.5 Données disponibles dans le golfe de Gascogne . . . . .	26
1.5.1 Vitesses des courants mesurées par les bouées et ADCP . . . . .	26
1.5.2 Vitesses radar HF . . . . .	28
1.5.3 Vitesses modélisées par IBI . . . . .	28
1.5.4 Données de vent . . . . .	30
1.5.5 Données satellite . . . . .	31
<b>2 Méthodes</b>	<b>32</b>
2.1 Préambule . . . . .	32
2.2 Interpolation Optimale des mesures de bouées . . . . .	32
2.2.1 Théorie . . . . .	32
2.2.2 Composition des matrices de covariances . . . . .	33
2.2.2.1 Manche orientale . . . . .	33
2.2.2.2 Sud-est du golfe Gascogne . . . . .	34
2.3 Calcul d'erreur . . . . .	35
2.3.1 Erreur relative et quadratique moyenne . . . . .	35

2.3.2	Distance de séparation et distance parcourue . . . . .	35
2.3.3	Erreur lagrangienne . . . . .	36
2.4	Correction de la vitesse induite par le vent . . . . .	37
2.5	Diagnostics lagrangiens . . . . .	37
2.5.1	Dispersion absolue d'un cluster . . . . .	38
2.5.2	Fonction de densité de probabilité . . . . .	39
2.5.3	Exposants de Lyapunov de taille finie . . . . .	39
2.5.4	Divergence lagrangienne dans un domaine fini . . . . .	40
2.6	Méthodes de clustering . . . . .	40
2.6.1	K-Means . . . . .	40
2.6.2	Self-Organizing Maps . . . . .	42
<b>3</b>	<b>Dispersion turbulente en Manche orientale</b>	<b>43</b>
3.1	Introduction . . . . .	43
3.2	Article : <i>Fusion of Lagrangian drifter data and numerical model outputs for improved assessment of turbulent dispersion</i> . . . . .	46
<b>4</b>	<b>Structures de convergence des courants dans le golfe de Gascogne</b>	<b>63</b>
4.1	Introduction . . . . .	63
4.2	Article : <i>Coastal current convergence structures in the Bay of Biscay from optimized high-frequency radar and satellite data</i> . . . . .	66
<b>5</b>	<b>Effet du vent sur les structures de convergence</b>	<b>86</b>
5.1	Introduction . . . . .	86
5.2	Article : <i>The impact of windage on the transport and accumulation of marine litter : a Lagrangian analysis in the Bay of Biscay</i> . . . . .	89
<b>6</b>	<b>Conclusions générales et perspectives</b>	<b>120</b>
	<b>Valorisations du travail</b>	<b>135</b>

# Liste des abréviations

*2dVar* : Two-dimensional Variational interpolation  
ACP : Analyse en Composantes Principales  
ADCP : Acoustic Doppler Current Profiler  
Arpege : Action de Recherche à Petite Echelle Grande Echelle  
CFL : Courant–Friedrichs–Lewy  
Chl-a : Chlorophylle-a  
CMEMS : Copernicus Marine Environment Service  
ECMWF : European Centre for Medium-Range Weather Forecasts  
ENSO : El Niño Southern Oscillation  
FDLD : Finite-Domain Lagrangian Divergence  
FSLE : Finite Size Lyapunov Exponents  
GEBCO : General Bathymetric Chart of the Oceans  
GDOP : Geometric Dilution of Precision  
GRDC : Global Runoff Data Centre  
HF : Haute Fréquence  
IBI : Iberia-Biscay-Ireland  
IO : Interpolation Optimale  
JERICO : Joint European Research Infrastructure for Coastal Observatories  
LCS : Lagrangian Coherent Structure  
LOG : Laboratoire d’Océanologie et Géosciences  
MARC : Modélisation et Analyse pour la Recherche Côtière  
MARS : Model for Applications at Regional Scale  
MERIS : Medium Resolution Imaging Spectrometer  
ODI : Ocean Drifter Irridium  
OLCI : Ocean and Land Color Imager  
OMA : Open-boundary Modal Analysis  
RANS : Reynolds Averaged Navier-Stokes  
RMS : Root Mean Square  
RONIM : Réseau français d’Observation du Niveau de la Mer  
SHOM : Service Hydrographique et Océanographique Français  
SSH : Sea Surface Height  
SWOT : Surface Water and Ocean Topography  
WRF : Weather Research Forecast



# Introduction générale

Les zones côtières, situées entre le continent et l'océan ouvert, sont soumises à de nombreuses pressions anthropiques. Ces dernières sont liées, entre autres, à l'urbanisation, aux activités industrielles, à la pêche et au tourisme et peuvent impliquer différentes problématiques (e. g. pollution, incidence croissante de la prolifération d'algues toxiques et d'espèces invasives). En parallèle de ces pressions anthropiques, ces espaces sont aussi soumis à des forçages naturels divers tels que les aléas extrêmes, l'érosion ou encore le dérèglement climatique.

Ainsi, les zones côtières, où habite 40% de la population mondiale (SIMPSON & SHARPLES, 2012), font l'objet de nombreuses problématiques scientifiques aux enjeux importants. La proximité de ces zones avec les habitations humaines rend les effets combinés de ces pressions anthropiques et de ces forçages naturels particulièrement préoccupants. Les chercheurs s'efforcent de comprendre et de prévoir les impacts de ces phénomènes sur les écosystèmes côtiers, les moyens de subsistance des populations locales, la sécurité, la santé publique, ainsi que la gestion des problématiques de pollution. De plus, la recherche et sauvetage en mer, ainsi que la compréhension de la dynamique de la faune marine, sont des enjeux cruciaux pour la gestion durable des zones côtières. Ces objectifs visent à assurer la préservation de ces environnements fragiles tout en répondant aux besoins socio-économiques des communautés qui en dépendent.

Les études menées dans le cadre de cette thèse concernent les processus de dispersion turbulente à la surface de l'océan et leur impact sur le transport et l'accumulation des déchets marins flottants, en particulier. Les processus de transport océaniques sont cruciaux car ils déterminent le devenir des quantités physiques, biologiques et chimiques, ainsi que du matériel particulaire dérivant avec les courants océaniques. Ils jouent un rôle clé dans la distribution spatiale de chaleur et de salinité (DONG et al., 2014), de nutriments (WHITNEY et al., 2005), de planctons (MCMANUS & WOODSON, 2012) et de la faune marine (BERLINE et al., 2013), mais aussi de substances polluantes comme les plastiques ou les hydrocarbures (HOOGENBOOM et al., 2024; HORTON, 2022; SUARIA et al., 2022), à l'échelle du globe.

## *Différentes échelles spatio-temporelles*

L'océan constitue un environnement complexe dont la dynamique peut évoluer selon une large gamme d'échelles spatio-temporelles (Fig. 0.1). L'océanographie physique à grande échelle s'intéresse à la dynamique des fluides et utilise la théorie fondamentale étayée par les observations océaniques acquises depuis le début du 20<sup>ième</sup> siècle. La variabilité des grandes échelles spatiales et temporelles est attribuée au changement climatique, aux phénomènes planétaires tels que ENSO, au cycle des saisons ou à la variabilité à l'échelle des bassins qui génèrent une circulation à grande échelle et influencent le climat à l'échelle mondiale (DICKEY & BIDIGARE, 2005). La circulation à grande échelle est alimentée par les vents de surface, générant une dé-

rive dite d'Ekman à la surface de la mer sous l'influence de la rotation de la Terre. Le transport d'Ekman, intégré sur les dizaines de mètres supérieurs de la colonne d'eau, crée des zones de convergence et de divergence des courants en surface, qui à leur tour alimentent le flux géostrophique à l'intérieur de l'océan (PRICE et al., 1987). Ces processus génèrent un flux horizontal à grande échelle permettant des connexions entre éco-régions et le transport à travers les différents bassins mondiaux. Les processus à grande échelle de l'océan ouvert sont généralement connus des chercheurs et sont à l'origine de mécanismes efficaces de transport des quantités physiques, chimiques et biologiques sur de grandes distances à l'échelle mondiale. Ces processus, en particulier les gyres océaniques, contribuent à l'accumulation de déchets marins dans certains bassins océaniques (KERSHAW, 2016).

Entre 10 et 100 kilomètres, correspondant à la mésoéchelle, la variabilité spatiale et temporelle est attribuée aux tourbillons et fronts, aux upwellings et aux marées (DICKEY & BIDIGARE, 2005). Au-delà de la configuration des gyres à l'échelle du bassin, l'océan est rempli de tourbillons. Les tourbillons de mésoéchelle sont des tourbillons en rotation lente, avec des diamètres de plusieurs dizaines à quelques centaines de kilomètres, des profondeurs de quelques dizaines à des centaines de mètres et des durées de vie de quelques semaines à quelques années (CHELTON et al., 2011). Les tourbillons de mésoéchelle forment également des fronts et des filaments entre eux en exerçant une contrainte sur les eaux de surface. Ces structures impactent fortement le transport océanique à l'échelle mondiale de part leur capacité à faire converger, mais aussi parce que la dérive vers l'ouest de ces structures potentiellement durables peut entraîner un transport sur plusieurs milliers de kilomètres (van SEBILLE et al., 2020).

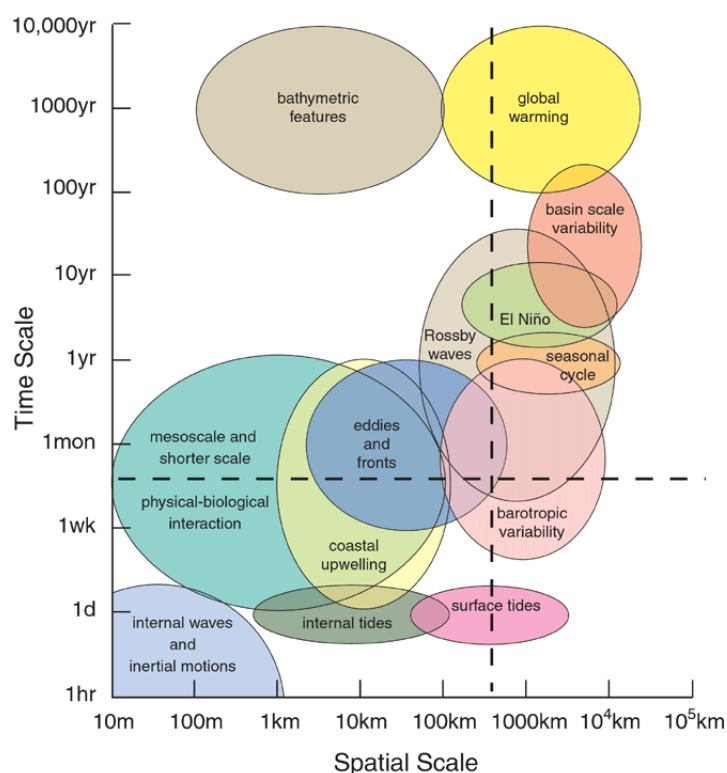


FIGURE 0.1 – Échelles de temps et d'espace de la variabilité océanique (D'après DICKEY et al., 2001)

Aux échelles inférieures à 10 km, correspondant à la sous-mésoéchelle, divers processus physiques peuvent expliquer la variabilité des courants et générer de petites structures tourbillonnaires. Ces processus de sous-mésoéchelle sont connus pour être très importants localement pour le transport de quantités physiques, chimiques et biologiques qui s'accumulent dans les fronts de densité et les tourbillons cycloniques (MCWILLIAMS, 2016 ; POJE et al., 2014). Différents modèles et observations ont démontré que les échanges verticaux induits à sous-mésoéchelle sont concentrés près des fronts, des jets et des tourbillons à l'échelle du kilomètre (D'ASARO et al., 2018 ; DAUHAJRE et al., 2017 ; HAZA et al., 2008 ; MCWILLIAMS, 2016). Plusieurs mécanismes sont responsables de la formation de zones frontales à cette échelle, tels que les interactions des courants avec la pente continentale, les upwellings, les panaches fluviaux, les tourbillons ou encore les ondes internes (AYOUCHE et al., 2020 ; MANN & LAZIER, 2005). Ces fronts à petite échelle sont associés à la convergence intense du flux horizontal induit par la dynamique à sous-mésoéchelle, et peuvent avoir un impact sur la distribution du matériel flottant, le piégeant ainsi comme le montrent D'ASARO et al. (2018), HUNTLEY et al. (2015), et HERNÁNDEZ-CARRASCO et ORFILA (2018). Les structures océaniques superficielles de sous-mésoéchelle présentent un défi tant pour les modèles numériques de circulation que pour les systèmes d'observations en raison du manque de résolution spatiale, particulièrement prononcé dans les zones côtières. Le caractère chaotique de ces structures rend leur modélisation extrêmement complexe, même avec une résolution appropriée.

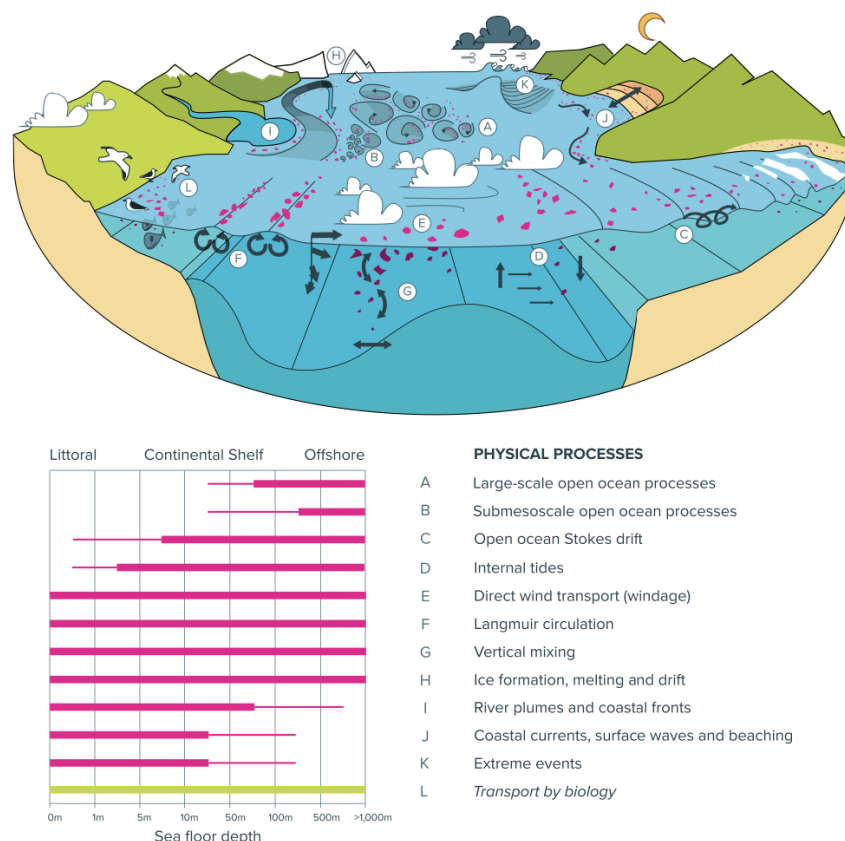


FIGURE 0.2 – Schéma des processus physiques qui affectent le transport du plastique (éléments roses) dans l'océan (panneau supérieur). Le tableau (panneau inférieur) identifie les régions dans lesquelles les différents processus sont importants (van SEBILLE et al., 2020).

A travers toutes ces échelles spatio-temporelles de variabilité, de nombreux processus contribuent au transport océanique de matière particulaire, et spécifiquement de déchets plastiques. van SEBILLE et al. (2020) a retracé les différents processus océaniques à l'origine du transport de plastique dans la Figure 0.2. Cette thèse se focalise sur les processus physiques B (*submesoscale processes*), E (*direct wind transport*), et I (*river plumes and coastal fronts*).

### Etat de l'art

À la sous-mésoscale et en zone côtière, la détection des structures de convergence des courants constitue un défi. En effet, les modèles numériques de circulation et les moyens d'observation peinent à identifier les structures de convergence côtière en raison de leur résolution spatio-temporelle limitée et du caractère chaotique de ces structures.

Afin de comprendre les processus sous-jacents au transport et à la dispersion de matière particulaire à la surface des océans et d'identifier les structures frontales, il est courant de se placer dans le référentiel lagrangien (Fig. 0.3). Ainsi, le transport de matière peut facilement être quantifié à l'aide d'un ensemble de trajectoires d'un grand nombre de particules individuelles, tandis que la dispersion et le mélange de matière sont généralement quantifiés en considérant le déplacement de paires de particules.

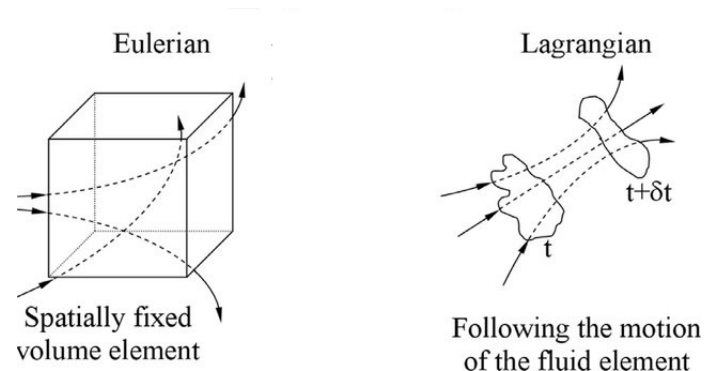


FIGURE 0.3 – Schéma des points de vue eulérien et lagrangien (SHADLOO et al., 2016)

Les structures frontales à la surface de la mer peuvent être identifiées en calculant des indicateurs lagrangiens spécifiques qui représentent des fonctions des trajectoires des particules de fluide. Les "fronts lagrangiens" sont déduits des maxima des indicateurs lagrangiens (PRANTS et al., 2012, 2014a, 2014b).

L'avantage des indicateurs lagrangiens réside dans leur capacité à analyser le mouvement des particules fluides sur une période de temps, ce qui permet de mieux capturer les processus de transport et de mélange ainsi que d'identifier des structures cohérentes dans le flux. Contrairement à cela, les indicateurs eulériens fournissent une vue instantanée de la dynamique locale du flux, en mettant en évidence des zones de rotation, comme avec la vorticité, ou de déformation, comme avec le critère d'Okubo-Weiss, mais ils peinent à suivre l'évolution temporelle des structures et le transport des particules.

Ainsi, cette étude emploie les indicateurs lagrangiens suivants :

- Les exposants de Lyapunov de taille finie (FSLE) : définis comme l'inverse du temps requis pour que deux particules de fluide en écoulement se séparent d'une distance initiale  $\delta_0$  à une distance finale  $\delta_f$  fixées (D'OVIDIO et al., 2004 ; HERNÁNDEZ-CARRASCO et al., 2011a ; LACASCE, 2008).
- La divergence lagrangienne (FDLD) : définie en intégrant et en moyennant la divergence horizontale dans le temps le long d'une trajectoire dans le domaine fini (HERNÁNDEZ-CARRASCO, ORFILA et al., 2018).

Ces diagnostics lagrangiens sont déduits des trajectoires de bouées dérivantes réelles ou de particules virtuelles. Ils permettent de visualiser des lieux et de caractériser l'intensité d'entraînement des particules dans un écoulement turbulent, ce qui est essentiel pour la compréhension de processus du transport et de la dispersion de quantités physiques, chimiques et biologiques.

Cependant, la détection de structures convergentes dans les cadres eulérien et lagrangien est fortement tributaire des champs de courants de surface utilisés pour les calculs.

Bien que certaines techniques d'observations telles que les bouées dérivantes (RYPINA et al., 2021), les flotteurs argo (MCKEE et al., 2023), ou encore les véhicules autonomes (MCCAMMON et al., 2021), permettent de détecter des structures frontales à fine échelle, ces données restent ponctuelles et ne permettent pas de cartographier la dispersion turbulente à fine échelle.

Les fronts de surface sont généralement détectés par satellite à l'aide de capteurs utilisant des longueurs d'ondes infrarouges, visibles et micro-ondes (PRANTS, 2022). Ils sont identifiés sur la base du calcul des gradients en identifiant les changements rapides des propriétés physiques de l'eau, telles que la température, la salinité ou la densité (BELKIN & O'REILLY, 2009 ; CHAPMAN et al., 2020 ; XING et al., 2023 ; ZAINUDDIN et al., 2006). Toutefois, ces techniques sont limitées par les résolutions spatiales des images satellites, par le temps de revisite et par la présence aléatoire d'une couverture nuageuse, selon le produit.

Les données issues de modèles de circulation océaniques sont aussi utilisées pour détecter les structures frontales comme la température de surface ou la Chlorophylle-a (Chl-a) (BAUDENA et al., 2021 ; CHAPMAN et al., 2020 ; COLBERG & REASON, 2006 ; TAGUCHI et al., 2012 ; TRAN et al., 2022). Cependant, ces études sont généralement menées en utilisant des champs de vitesses de faible résolution, le plus souvent insuffisante pour capter les structures à sous-mésoéchelle, ou concernent des régions d'études assez conséquentes (Océan Pacifique nord, golfe du Mexique, Mer de Chine. . .). De plus, la validation de ces données nécessite diverses observations parfois indisponibles dans certaines zones.

En zone côtière, l'un des moyens les plus efficaces et performants pour cartographier les structures convergentes est l'utilisation des radars (Radio Detection and Ranging) Haute-Fréquence (HF). La technologie radar HF est largement répandue. Elle utilise des ondes électromagnétiques pour mesurer la vitesse et la direction des courants océaniques en détectant le décalage Doppler des signaux radar réfléchis par les ondulations de la surface de l'eau. Sur nos côtes, de nombreuses stations radar font face à l'océan, permettant des observations pluriannuelles à haute résolution des zones côtières (PADUAN & ROSENFELD, 1996 ; ROARTY et al., 2019). Leur portée peut atteindre 300 km. La présence de radars HF permet de détecter des structures frontales avec une fine résolution spatio-temporelle (BOURG et al., 2024 ; TRAN et al., 2022).

*Objectif de la thèse et zones d'étude*

L'objectif de cette thèse est de proposer une méthode de caractérisation des structures de convergence des courants marins en zone côtière, ainsi que de leur distribution spatiale et de leur variabilité temporelle. La dispersion de matière dérivant avec les courants, ou leur accumulation, se produit au sein de ces structures, leur conférant une importance capitale. La méthode emploie une approche lagrangienne dont la pertinence dépend fortement de la qualité des champs de vitesse de surface utilisés en entrée. Pour accroître la fiabilité de la méthode et rendre plus réaliste l'identification des structures de convergence, une technique de fusion de données hétérogènes a été mise en œuvre et validée dans deux régions géographiques très différentes par leur dynamique : la Manche orientale et le sud-est du golfe de Gascogne.

La première région d'intérêt de cette étude couvre la **Manche orientale**. En premier lieu, la Manche est une région d'eaux peu profondes du plateau continental nord-ouest européen reliant l'Océan Atlantique à la Mer du Nord (Fig.0.4). C'est un carrefour biogéographique et économique important, siège de nombreuses activités maritimes. La Manche couvre une superficie d'environ 76 000 km<sup>2</sup> et forme un corridor de 500 km de long et de 250 km de large au maximum, 35 km de large au minimum dans le détroit du Pas de Calais, en Manche orientale. Les marnages (différence de hauteur d'eau entre une pleine mer et une basse mer consécutive) y sont inférieurs à 5 m.

La Manche est une mer méga-tidale caractérisée par des marées de grande amplitude. Bien que les courants résultant de l'action des vagues et du vent soient forts, les courants de marée semi-diurne prédominent la circulation locale et structurent fortement les écosystèmes et la bathymétrie (REYNAUD et al., 2003).

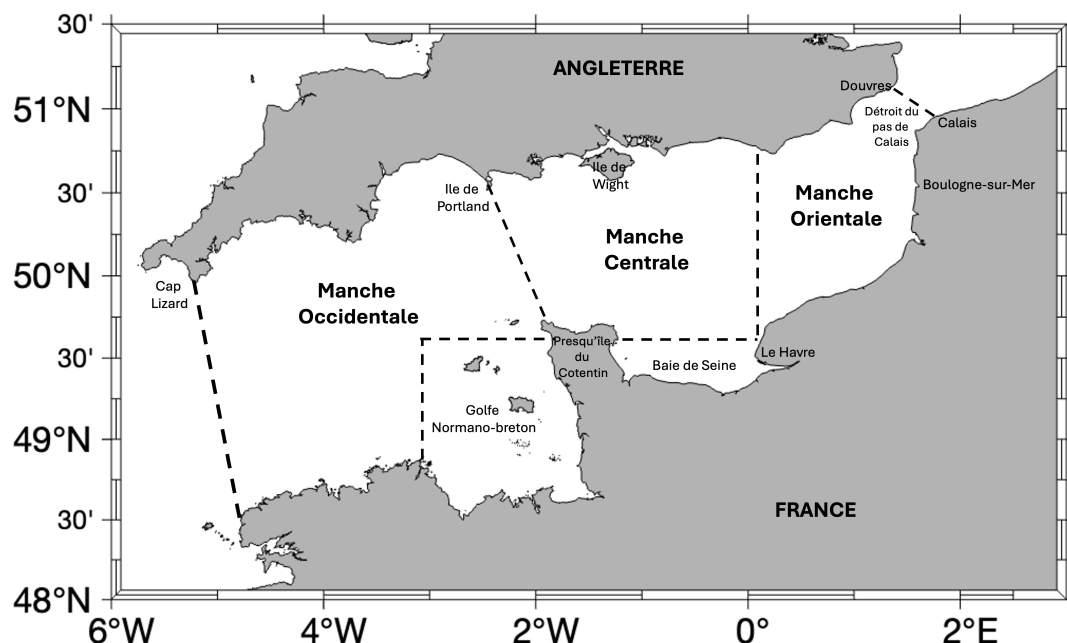


FIGURE 0.4 – Domaine d'étude - La Manche.

L'étude de la salinité et du transport des particules en Manche a révélé un transport général des eaux depuis l'Atlantique vers la Mer du Nord (SALOMON & BRETON, 1993), en concordance avec la direction des courants résiduels moyens. En effet, une circulation à grande échelle dans l'Océan Atlantique Nord génère une différence de niveau de la mer entre la Manche et la Mer du Nord, entraînant un flux résiduel de l'Océan Atlantique vers la Mer du Nord.

D'un point de vue hydrologique, il existe deux bassins en Manche : la Manche occidentale, située à l'ouest, et à l'est, le bassin oriental divisé en Manche centrale et Manche orientale (Fig. 0.4). La Manche occidentale est plus profonde et fortement influencée par les eaux provenant de l'Océan Atlantique tandis que la Manche centrale et la Manche orientale sont peu profondes et fortement influencées par le déversement de la Seine.

La seconde région d'intérêt de cette étude couvre le **sud-est du golfe de Gascogne**. Le golfe de Gascogne est situé dans la zone tempérée de l'Océan Atlantique nord-est, entre le nord-ouest de la France (au large de la Bretagne) et le nord-ouest de l'Espagne (Galice). De la même manière que la Manche, le golfe de Gascogne est un carrefour biogéographique et économique important, siège de nombreuses activités maritimes. Au total, la superficie du golfe occupe environ 175 000 km<sup>2</sup>. Dans cette région, le marnage est inférieur à 2 m.

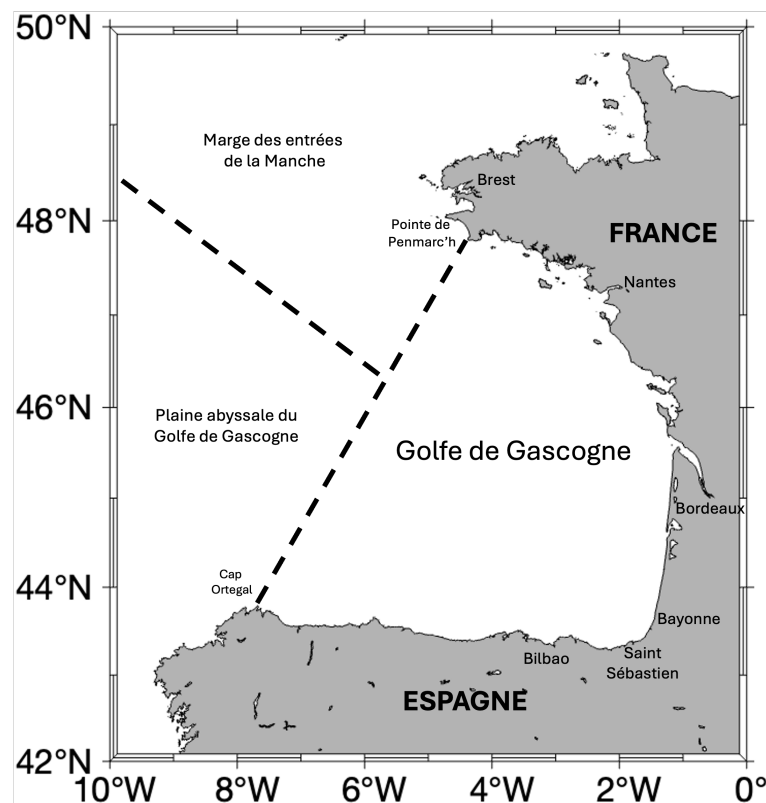


FIGURE 0.5 – Domaine d'étude - le golfe de Gascogne.

Le golfe de Gascogne est marqué par la séparation du domaine en plateau continental, plaine abyssale et à leur limite, le talus continental. Ce talus est entaillé par de nombreux canyons, le plus visible étant le canyon de Capbreton, où l'isobathe de 1000 m se trouve à seulement 3 km de la côte. Les vallées profondes permettent le transport des sédiments continentaux vers les

bassins océaniques à partir des principaux fleuves (Vilaine, Loire, Gironde et Adour), tous situés en France, tandis que les fleuves du nord de l'Espagne sont plus courts et ont un plus faible débit.

Contrairement à la Manche, la dynamique dans le golfe de Gascogne présente une variabilité caractérisée par une grande diversité de processus à l'origine des courants marins, tels que la marée, les vents, les apports d'eau douce, ainsi que la circulation à grande échelle dans l'Atlantique (VANDERMEIRSCH et al., 2012). La bathymétrie, elle aussi très variable (plateau, talus continental, plaine abyssale), contraint et influence fortement les courants marins. Tous ces phénomènes donnent lieu à une forte variabilité saisonnière de la circulation générale dans le golfe de Gascogne (CHARRIA et al., 2013; RUBIO et al., 2019; SOLABARRIETA et al., 2014; VANDERMEIRSCH et al., 2012).

### *Méthode d'optimisation*

Dans chacune des régions d'étude, différents jeux de données sont disponibles : champs de courants de surface issus de modèles de circulation océanique côtière, issus de radars HF (seulement dans le golfe de Gascogne), images satellites, trajectoires de bouées dérivantes. Dans le cadre de cette thèse, la recherche s'est focalisée sur la combinaison de données la plus optimale pour étudier la dispersion à sous-mésoéchelle et ses effets sur le transport particulaire en Manche orientale et dans le sud-est du golfe de Gascogne.

Pour fusionner les données, l'Interpolation Optimale (IO) a été utilisée. Développée par Gandin (1963) et appliquée à la modélisation atmosphérique, l'IO a été largement utilisée dans différents domaines des géosciences pour la cartographie de la température de surface de la mer (BRETHERTON et al., 1976), l'optimisation de la vitesse des courants modélisés (MOLCARD et al., 2003; SENTCHEV & YAREMCHUK, 2015) ou l'optimisation de la topographie (WU et al., 2021).

Comparée à d'autres approches d'optimisation de la circulation océanique telles que les méthodes variationnelles (par exemple, KALNAY, 2002; SENTCHEV et YAREMCHUK, 1999; WIKLE, 2005), l'IO présente plusieurs avantages. Tout d'abord, la méthode est simple à mettre en œuvre et garantit un équilibre raisonnable entre la complexité de calcul et la cohérence statistique des écarts entre les grandeurs utilisées lors que la fusion de données. Deuxièmement, la précision du champ de vitesse reconstruit peut être évaluée à chaque pas de temps.

### *Étapes de la thèse*

La **première étape** de la thèse vise à acquérir et traiter des données observationnelles. Ces dernières sont des mesures lagrangiennes issues du déploiement de bouées dérivantes dans les deux zones d'étude. En Manche orientale, deux campagnes ont été menées, avant le début de la thèse, consistant au déploiement de deux bouées en novembre 2020, et quatre bouées en mai 2021. Dans le sud-est du golfe de Gascogne, deux campagnes ont été menées durant la thèse, auxquelles j'ai participé. Ces campagnes ont consisté au déploiement de vingt-trois bouées en avril 2022 et neuf bouées en mai 2022.

L'utilité de l'acquisition de données observationnelles est multiple. Ces données sont employées d'une part pour étudier la fiabilité des instruments de mesure et modèles à représenter



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la dynamique océanique dans les deux zones d'étude considérées via différents diagnostics de calcul d'erreur (erreur quadratique moyenne, distance de séparation entre trajectoire réelle et simulée, erreur lagrangienne...), ce travail constituant la **deuxième étape** de cette thèse.

D'autre part, les données observationnelles sont employées pour obtenir des champs de vitesses optimisés dans chaque région d'étude en utilisant une technique de fusion de données, l'IO, constituant la **troisième étape** de la thèse. Cette méthode de fusion de données hétérogènes et d'observations lagrangiennes est mise en place dans chaque région d'étude afin de permettre une nette amélioration de la représentation de la dynamique et des phénomènes de transport et de dispersion inhérents à chaque zone d'étude.

La **quatrième étape** de la thèse consiste à caractériser la dynamique et la dispersion dans les deux zones d'étude via différents diagnostics lagrangiens. Les champs de vitesse optimisés par fusion de données sont utilisés pour fournir une estimation de la dispersion et du transport de matière particulaire dans chaque région d'étude, avec une application aux déchets marins dans le golfe de Gascogne.

### *Structure du manuscrit*

En conformité avec l'objectif de la thèse énoncé précédemment, la structure de ce manuscrit est la suivante : le chapitre 1 fournit une description des deux régions d'étude, la Manche orientale et le sud-est du golfe de Gascogne en détaillant leurs problématiques, leurs caractéristiques géographiques, forçages physiques, et processus hydrodynamiques. Le chapitre 1 aborde aussi les différents jeux de données utilisés en Manche orientale et dans le sud-est du golfe de Gascogne : les données issues de modèle, de radar HF, de satellite et de bouées dérivantes.

Le chapitre 2 présente les différentes méthodes de traitement et d'analyse de données utilisées dans cette thèse. Le calcul des différentes quantités lagrangiennes utilisées pour estimer le transport et la dispersion de matière particulaire est aussi abordé dans ce chapitre, avec pour objectif de représenter la distribution spatiale des structures de convergence des courants marins.

Le chapitre 3 présente les résultats de fusion des données de dérive lagrangienne et des sorties d'un modèle haute résolution pour une meilleure estimation de la dispersion turbulente en Manche orientale.

Le chapitre 4 aborde la question de l'identification des structures de convergence des courants marins dans le sud-est du golfe de Gascogne, à l'origine de l'accumulation potentielle de déchets marins grâce à la fusion de données de dérive lagrangienne et de champs de courants issus de radars HF.

Le chapitre 5 présente l'influence du vent sur les structures de convergence des courants susceptibles d'accumuler les déchets marins flottants dans le sud-est du golfe de Gascogne.

Le chapitre 6 résume les principaux résultats de cette thèse, les questions restant ouvertes et les perspectives inhérentes à ces travaux.

# Chapitre 1

## Zones d'étude et jeux de données

### 1.1 Préambule

La Manche orientale et le sud-est du golfe de Gascogne sont deux régions côtières européennes importantes, accueillant le tourisme, la pêche et un trafic maritime dense, en particulier en Manche. Cependant, la compréhension de la dynamique à sous-mésoéchelle dans ces zones reste limitée, ce qui entrave la résolution efficace des problèmes de pollution, de recherche et de sauvetage en mer, ainsi que la connaissance de la dynamique de la faune marine.

Ces deux régions présentent des caractéristiques géographiques, topographiques et dynamiques distinctes, ce qui les différencie nettement. Cette disparité souligne la nécessité d'adopter des stratégies différentes en ce qui concerne l'acquisition des données océanographiques.

Par exemple, le déploiement de bouées dérivantes en Manche orientale se fait sur des durées moins importantes que dans le golfe de Gascogne. Ceci s'explique par les forts courants présents dans la zone, pouvant atteindre 2 m/s au jusant, et par le trafic maritime soutenu pouvant engendrer la destruction du matériel.

En outre, les côtes espagnoles sont équipées de radars HF qui permettent d'obtenir en temps réel les courants océaniques de surface sur place, tandis que la Manche orientale en est dépourvue.

Ce premier chapitre présente le contexte marégraphique et la courantologie des deux régions d'études, ainsi que les différents jeux de données disponibles.

## 1.2 Manche

### 1.2.1 Contexte marégraphique et courantologie

La Manche, située entre l'Angleterre au nord et la France au sud, est un carrefour biogéographique et économique majeur, accueillant de nombreuses activités maritimes (Fig.0.4). En Manche, la Manche orientale spécifiquement est une région d'eaux peu profondes du plateau continental nord-ouest européen reliant la Manche à la Mer du Nord (Fig.0.4 - rectangle rouge).

La carte de la bathymétrie en Manche est présentée en Figure 1.1. Les profondeurs diminuent progressivement d'ouest en est. La profondeur maximale de 174 m est atteinte dans la fosse centrale, au nord-ouest de la presqu'île du Cotentin. La profondeur moyenne est d'une cinquantaine de mètres. La région est caractérisée par une bathymétrie très irrégulière, avec la présence de nombreux bancs de sable, grossièrement orientés dans la direction du courant dominant, avec une profondeur pouvant atteindre quelques mètres seulement à marée basse.

La Manche est une mer méga-tidale où les courants de marée semi-diurne prédominent la circulation locale (REYNAUD et al., 2003) et structurent fortement les écosystèmes et la bathymétrie. La distribution spatiale des courants est étroitement liée à la bathymétrie, et certaines configurations morphologiques côtières induisent des phénomènes distincts tels que l'augmentation de la vitesse des courants dans les goulets et au niveau des caps, le remplissage et le vidage des baies, l'asymétrie du flot et du jusant à l'embouchure des estuaires, ainsi que l'émergence de mouvements cycloniques.

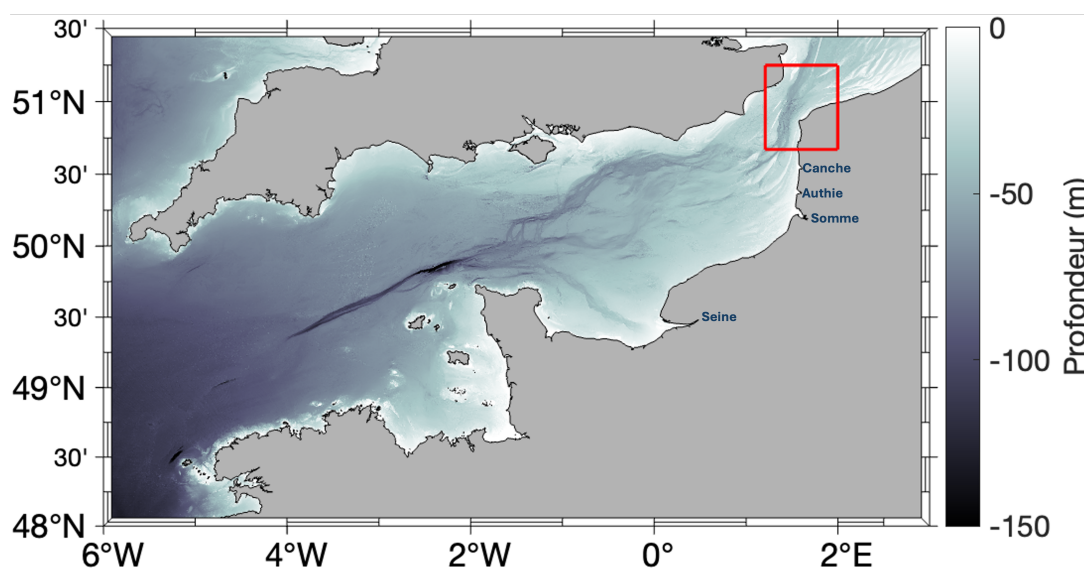


FIGURE 1.1 – Carte des profondeurs en Manche. Les données bathymétriques proviennent de SHOM, 2015. MNT Bathymétrie de façade Atlantique (Projet Homonim) ([http://dx.doi.org/10.17183/MNT\\_ATL100m\\_HOMONIM\\_WGS84](http://dx.doi.org/10.17183/MNT_ATL100m_HOMONIM_WGS84)). Les noms des principales rivières pouvant influencer la zone d'étude sont indiqués en bleu. La région d'étude, la Manche orientale, est indiquée par un rectangle rouge.

La carte de la circulation moyenne en Manche est présentée en Figure 1.2. En Manche occidentale, l'entrée des eaux Atlantique s'effectue le long des côtes de la Bretagne par une branche du courant résiduel, tandis qu'une seconde branche est dirigée vers l'extrémité sud de l'Angleterre. Dans le golfe Normano-breton, la circulation est marquée par des structures tourbillonnaires autour des différents archipels.

La circulation en Manche centrale est influencée par la présence d'un vaste tourbillon anti-cyclonique, orienté dans le sens des aiguilles d'une montre, formé dans le sillage du Cotentin, connu sous le nom de tourbillon de Barfleur. Les vitesses de circulation atteignent environ 2 à 3 cm/s en l'absence de vent (LAZURE & DESMARE, 2012). À l'est, l'estuaire externe de la Seine se distingue par des écarts de courants significatifs entre la surface (dirigée vers l'ouest puis le nord) et le fond (orienté vers l'estuaire).

La circulation en Manche orientale est dirigée vers le nord-est, les courants étant renforcés par le rétrécissement du détroit du Pas de Calais. La vitesse du courant peut atteindre 2 m/s durant la marée de vive-eau, générant des courants résiduels variant entre 5 et 10 cm/s au niveau du détroit du Pas de Calais orientés vers le nord-est, liés à la propagation de la marée et à l'effet du frottement (LAZURE & DESMARE, 2012). Ces courants sont beaucoup plus faibles que les courants instantanés mais leur influence sur le transport à long terme des masses d'eau est déterminante car ils sont permanents, leur intensité étant modulée par les cycles vives eaux- mortes eaux. Le courant de marée perd environ 20% de son intensité durant la marée descendante. En Manche orientale, l'augmentation de la vitesses des courants de marée couplée à la diminution de la profondeur, occasionne un mélange sur toute la colonne d'eau (AGOUMI, 1982).

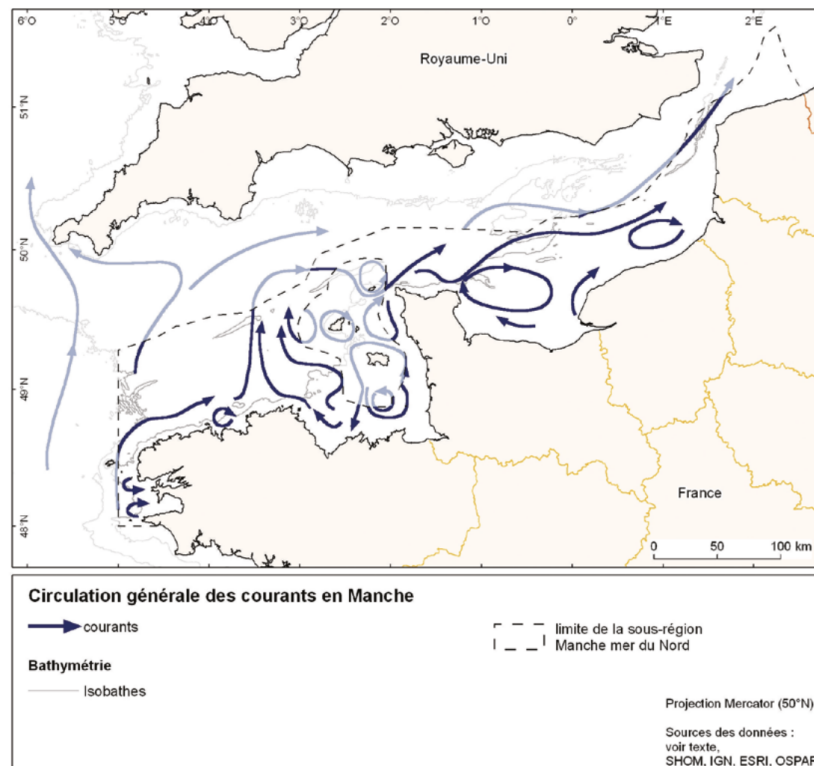


FIGURE 1.2 – Circulation moyenne en Manche (Sources : IGN, ESRI, SHOM, OSPAR, 2010).

BRYLINSKI et al. (1991) ont montré l'existence d'un fleuve central appelé "rivière Manche" constitué des eaux provenant de l'Atlantique. Cette veine d'eau provoque l'apparition de gradients côte-large fortement marqués concernant différents paramètres hydrologiques et biologiques : salinité, turbidité, concentration en sels nutritifs, phytoplanctons... (KELLY GERREYN et al., 2006 ; VISSER et al., 1996).

### 1.2.2 Contexte météorologique

#### *Le régime des vents*

En Manche, le vent affecte de manière significative la circulation locale. Les vents du sud-ouest peuvent augmenter le flux moyen vers l'est tandis que les vents du nord-ouest et du nord-est peuvent réduire le flux de marée s'opposant au vent et même l'inverser (LAZURE & DESMARE, 2012). Sous l'effet conjoint de la tension de vent et de la force de Coriolis, les eaux de surface se mettent en mouvement conformément à la théorie d'Ekman. Le vent est responsable de 3/4 du transport des particules d'eau entre la Manche et la Mer du Nord (SALOMON & BRETON, 1993).

#### *Le régime des houles*

La houle qui se forme au centre de l'Atlantique perd en intensité en approchant la Manche orientale, généralement inférieure à 1.25 m. Environ 75% des hauteurs de vagues annuelles sont inférieures à 1 m (JOUANNEAU, 2013). Les périodes de houle, majoritairement entre 5 et 9 s, caractérisent des "mers de vent". Bien que des périodes plus longues puissent être présentes, les faibles hauteurs de vagues estivales et les importantes amplitudes de marée suggèrent que la houle n'a qu'un impact limité sur la dynamique en Manche.

#### *Les apports fluviaux*

L'hydrodynamisme particulier des mers à marées et des détroits (accélération des courants, formation de front, augmentation de la turbulence, structuration des eaux en bandes parallèles, effet de cap...) agit sur le devenir des apports fluviaux, soit en les diluant, soit en les confinant à la côte. Depuis la baie de Seine jusqu'au Pas-de-Calais, plusieurs fleuves se jettent en Manche orientale fournissant eau douce, nutriments, matières en suspension et matières organiques à la zone côtière : la Seine, la Somme, la Canche et l'Authie (Fig. 1.1) principalement, qui sont à l'origine de la dessalure de la bande littorale. Les apports d'eau douce provenant de l'Angleterre sont estimés à  $327 \text{ m}^3 \cdot \text{s}^{-1}$  tandis que  $665 \text{ m}^3 \cdot \text{s}^{-1}$  proviennent des fleuves français ( $410 \text{ m}^3 \cdot \text{s}^{-1}$  concernant la Seine) (DAUVIN & LOZACHMEUR, 2006). Ces apports d'eau douce sont à l'origine de la création d'un fleuve côtier se mélangeant peu avec la veine atlantique (BRYLINSKI et al., 1991).

Cependant, la Manche orientale n'est que peu influencée par les apports fluviaux. Seuls les plus grands fleuves sont susceptibles de créer une circulation spécifique. À l'échelle de la Manche, seule la circulation induite par les apports en eaux douces de la Seine est capable de créer des différences de courants significatives entre la surface et le fond, différences n'atteignant pas ou peu la Manche orientale (LAZURE & DESMARE, 2012).

### 1.3 Données disponibles en Manche orientale

#### 1.3.1 Vitesses des courants mesurées par les bouées

Au total, 6 bouées lagrangiennes ont été déployées dans le détroit du Pas de Calais durant deux campagnes de mesures C1 et C2, destinées à estimer la dispersion turbulente en Manche (Fig. 1.3a, b). Les caractéristiques des différentes campagnes menées en Manche sont présentées dans le tableau 1.1.

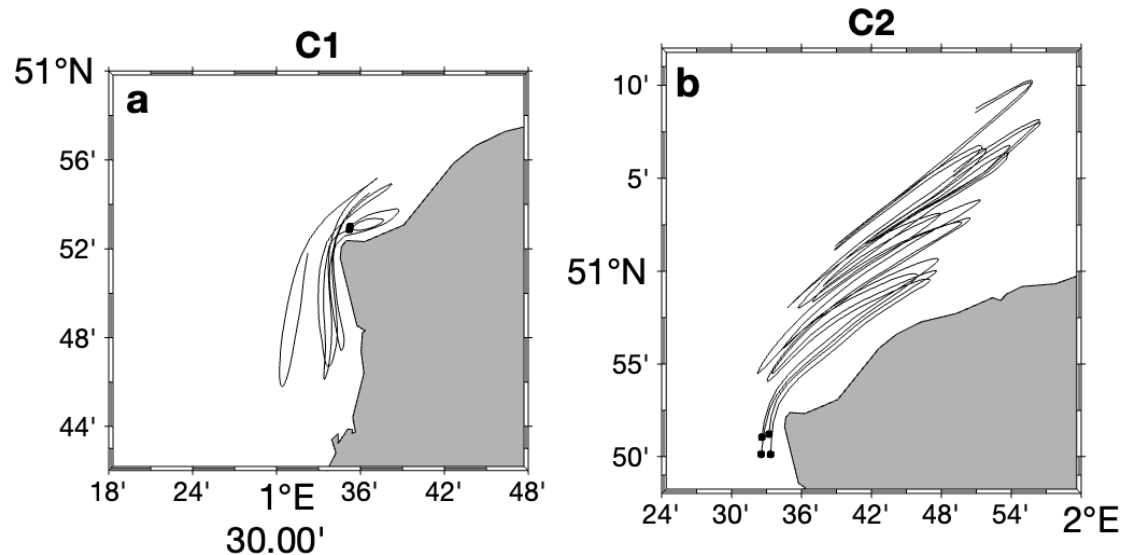


FIGURE 1.3 – Trajectoires des bouées lagrangiennes au cours des différentes campagnes (2 durant C1, 4 durant C2). La localisation du déploiement est symbolisée par un point noir. Les zooms ont été ajustés pour fournir la meilleure représentation des trajectoires.

TABLE 1.1 – Tableau récapitulatif des différentes campagnes menées en Manche orientale. Les dates et heures sont données en UTC.

Campagnes	Bouées	Lâcher	Récupération	Marée	Vent
C1	2	26/11/20 8h30	27/11/20 11h00	Moyenne (60)	Nord-est (4 m/s)
C2	4	10/05/21 9h15	12/05/21 07h30	Moyenne (78)	Sud-ouest (6 m/s)

Deux types de bouées ont été utilisés : les bouées de surface côtières Nomad fabriquées par SouthTek (<https://www.southteksl.com/>) (Fig. 1.4b) et des bouées dérivantes fabriquées au laboratoire (Fig. 1.4a). Ces dernières sont constituées d'une coque cylindrique en PVC de 0.6 m de long et de 0.1 m de diamètre, lestée dans sa partie inférieure. Une fine plaque carrée de 0.3 x 0.3 m a été installée dans la partie supérieure de la coque pour assurer une meilleure stabilité à la verticale et réduire le tangage. Les bouées dérivantes étaient équipées d'un émetteur-récepteur GPS/satellite Smartone du réseau satellitaire Global Star. Toutes les bouées dérivantes étaient équipées d'une ancre de 0.5 m de long positionnée dans la couche d'eau entre 0.8 et 1.3 m de profondeur, ce qui leur permettait de dériver avec les courants de surface sans être soumises à l'effet du vent. Les vitesses observées des courants de surface ont été estimées à partir des

trajectoires des bouées avec un pas de temps de 15 minutes, ce qui correspondait à la période nominale de positionnement des bouées par GPS.

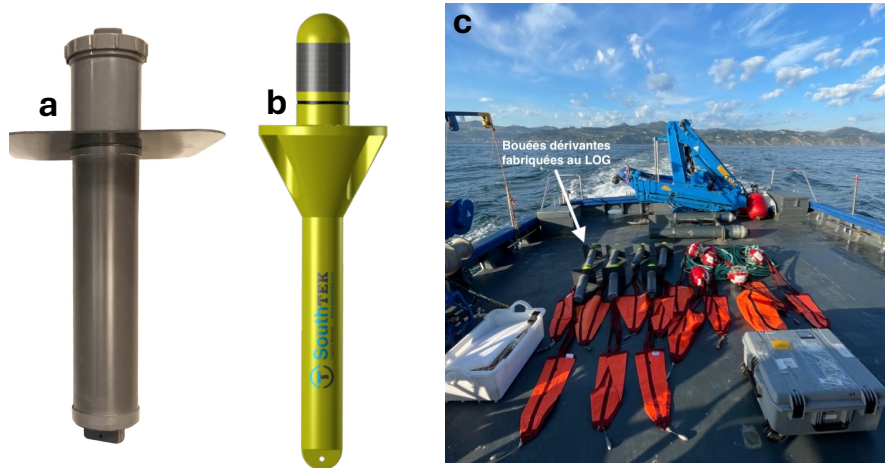


FIGURE 1.4 – Photo des bouées dérivantes fabriquées au LOG (a), des bouées Nomad fabriquées par SouthTek (b) et des bouées équipées d'une ancre flottante prêtes à être déployées (c).

### 1.3.2 Vitesses modélisées

Dans ces travaux de thèse, j'ai réalisé des simulations lagrangiennes d'advection de particules forcées par des courants barotropes 2D et par des niveaux d'eau modélisés dans le cadre du projet d'océanographie côtière opérationnelle MARC ("Modélisation et Analyse pour la Recherche Côtière") (PINEAU-GUILLOU, 2013; PREVIMER, 2014). Les données de prévision multi-échelles en temps réel et historiques pour les côtes françaises métropolitaines incluant la Manche Est sont disponibles gratuitement sur le site web du projet (<https://marc.ifremer.fr>). Les champs de courant, de salinité, de température et d'élévation du niveau de la mer ont été estimés à l'aide du Modèle d'Application à l'échelle Régionale (MARS) (LAZURE & DUMAS, 2008).

Le modèle MARS a été développé pour simuler les flux dans diverses zones côtières, allant des échelles régionales à l'échelle côtière de petites baies ou d'estuaires, où la circulation est généralement influencée par une combinaison de processus (LAZURE & DUMAS, 2008). La modélisation hydrodynamique est basée sur les équations de Navier-Stokes qui résolvent uniquement les composantes moyennes de l'écoulement en appliquant l'opérateur moyenne d'ensemble ou de Reynolds aux équations. C'est donc un modèle RANS (Reynolds Averaged Navier-Stokes). La turbulence est représentée comme l'écart à la moyenne et ses effets sont paramétrisés à l'aide de relations le plus souvent empiriques. Les modèles RANS sont largement utilisés en raison de leur efficacité, avec un temps de calcul relativement abordable. Cependant, ils ne captent pas les détails fins des écoulements turbulents. Des informations détaillées concernant les équations du modèle, le couplage des modes barotrope et barocline, la discrétisation du modèle, les méthodes de résolution, la stabilité de calcul selon le critère CFL (tableau 1, LAZURE et DUMAS, 2008) et les coûts sont méticuleusement décrites dans Lazure et Dumas (2008). Le modèle tient compte des conditions cinématiques aux limites de la surface libre et du fond, en fonction des termes de friction (LAZURE & DUMAS, 2008). La paramétrisation de la turbulence utilisée dans le modèle

suit l'approche décrite dans GASPAR et al. (1990).

Pour simuler précisément la dynamique des ondes de tempête, le modèle s'étend suffisamment au nord et à l'ouest pour capter le développement des dépressions, qui génèrent des ondes se propageant dans la Manche à plus grande distance dans l'Océan Atlantique ou dans la Mer du Nord (IDIER et al., 2012). Le modèle numérique utilise des configurations imbriquées avec des résolutions progressives : (i) 2 km couvrant l'Atlantique nord-est (niveau 0), (ii) 700 m à l'échelle régionale, englobant la Manche (niveau 1), et (iii) 250 m pour la Manche est (niveau 2). Cette technique d'imbrication permet de saisir avec précision les interactions entre les processus à grande et à petite échelle. Cette technique permet le transfert de tous les champs résolus des niveaux de résolution inférieurs vers les limites ouvertes des niveaux de résolution supérieurs. La résolution temporelle des sorties est de 1 h pour les niveaux 0 et 1, et de 15 min pour le niveau 2.

Tous les détails pertinents concernant le choix de la bathymétrie, les forçages météorologiques et de marée, les améliorations dans la paramétrisation du modèle (comme le frottement de surface et le coefficient de traînée), et la méthodologie employée pour le calcul de l'onde de tempête sont documentés dans le travail de PINEAU-GUILLOU (2013). Les données bathymétriques pour la configuration de niveau 2 proviennent du Service Hydrographique et Océanographique Français (SHOM). Les conditions limites de marée ont été dérivées du modèle global de marée FES2004 (LYARD et al., 2006). Les zones intertidales ont été simulées à l'aide d'un schéma de mouillage et de séchage. Le coefficient de traînée utilisé pour la paramétrisation de l'effet du vent est basé sur le coefficient variable de Charnock du modèle de vagues WWIII (ARDHUIN et al., 2011). Pour maintenir la stabilité du CFL, le pas de temps de la modélisation a été fixé à 30 s pour le modèle de niveau 2.

De plus, les sorties du modèle numérique ont fait l'objet d'un processus d'évaluation (PINEAU-GUILLOU, 2013 ; PREVIMER, 2014) impliquant les observations disponibles, des indicateurs de confiance et des descripteurs d'états du système. Concernant les niveaux d'eau, le modèle a été validé en février 2010 sur 19 marégraphes du réseau permanent RONIM (Réseau Français d'Observation du Niveau de la Mer). Pour la marée, les erreurs quadratiques moyennes sont divisées par deux entre le niveau 0 et le niveau 2 : elles sont en moyenne de 22 cm pour le niveau 0, 21 cm pour le niveau 1, et 11 cm pour le niveau 2. Des résultats similaires sont observés pour les niveaux d'eau (marée et déferlement) avec des erreurs quadratiques moyennes de 26 cm pour le niveau 0, 24 cm pour le niveau 1, et 16 cm pour le niveau 2. Ces améliorations sont attribuées à l'augmentation de la résolution spatiale du niveau 0 au niveau 2 (de 2 km à 250 m), ainsi qu'à l'incorporation du modèle de marée SHOM cstFRANCE (avec 115 composantes harmoniques) aux limites du modèle (PINEAU-GUILLOU, 2013). Par conséquent, il est établi que la comparaison des sorties du modèle pour la configuration de niveau 2 (avec une résolution spatiale horizontale de 250 m et une résolution temporelle de sortie de 15 min), englobant la Manche orientale avec les données in situ disponibles, est suffisamment précise pour être utilisée dans cette étude pour une modélisation plus poussée de l'advection des particules dérivantes pour estimer la dispersion turbulente.

Représentés à l'origine sur une grille C d'Arakawa, les courants de surface ont été interpolés sur la grille A d'Arakawa (ARAKAWA & LAMB, 1977) pour une analyse plus poussée et l'optimisation des vitesses du modèle.

Cette étude est basée sur deux séries de simulations MARS d'une durée d'un an couvrant



la période de janvier à décembre 2020 contenant C1, et de janvier à décembre 2021 contenant C2. Ces longues périodes de simulation sont utilisées pour la sélection des membres d'ensemble composant la matrice de covariance utilisée lors de la méthode d'optimisation (voir chapitre 2.2). Les simulations couvrent la zone suivante : 1.2°E-2°E; 50.6°N-51.2°N (Fig. 1.5).

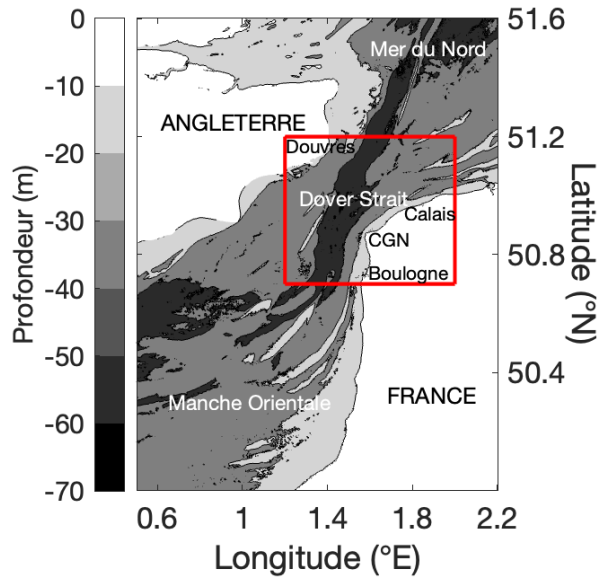


FIGURE 1.5 – Domaine de modélisation. La bathymétrie est représentée en gris. Le rectangle rouge délimite la région où les mesures lagrangiennes ont été effectuées.

La figure 1.6 présente les vitesses des courants de surface durant la campagne C1 au flot (a) et au jusant (b), illustrant les vitesses importantes observées et modélisées dans la zone d'étude.

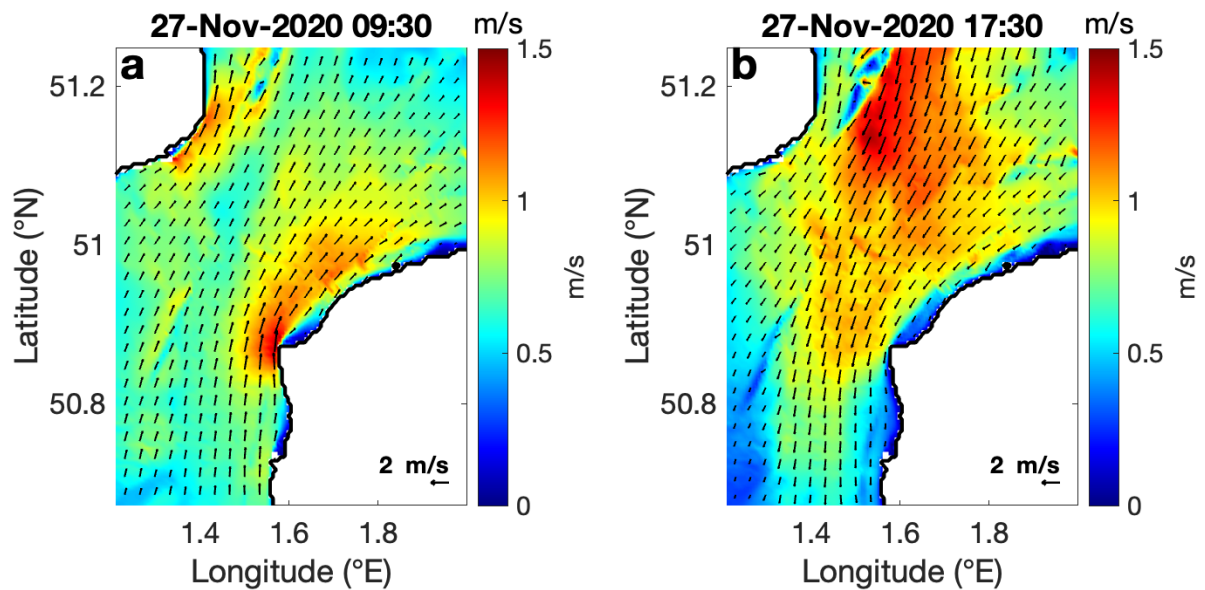


FIGURE 1.6 – Carte des vitesses des courants de surface issues du modèle MARS durant C1 pendant le flot (a) et le jusant (b).

### 1.3.3 Données de vent

Les données météorologiques (vent, température, humidité et pression atmosphérique) sont utilisées pour forcer le modèle MARS. Ces données sont fournies par le modèle atmosphérique opérationnel Arpege (Action de Recherche Petite Echelle Grande Echelle) de Météo-France avec une résolution spatiale de 5 km et une résolution temporelle de 1 h.

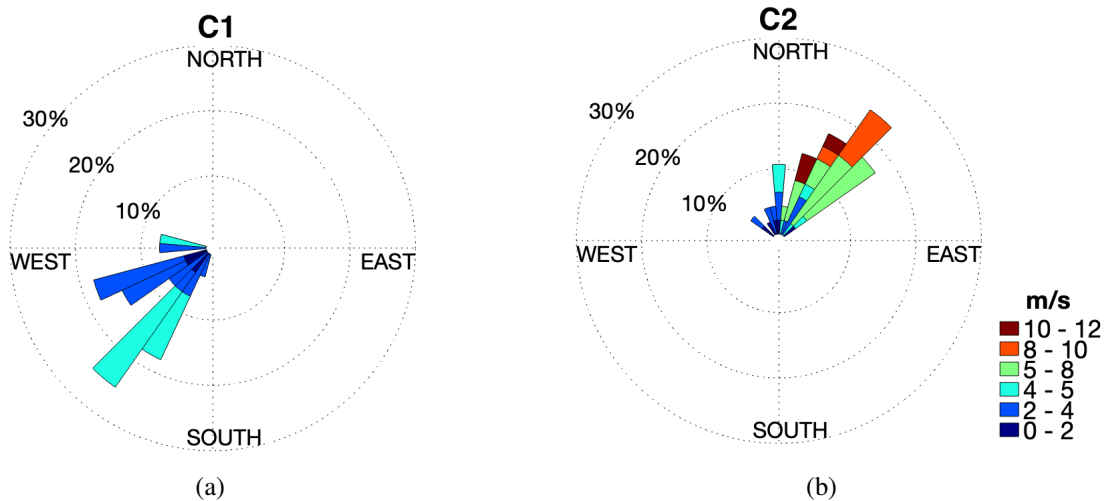


FIGURE 1.7 – Roses des vents provenant du modèle Arpege durant chacune des campagnes menées en Manche orientale.

## 1.4 Golfe de Gascogne

### 1.4.1 Contexte marégraphique et courantologie

La seconde région d'intérêt de cette étude couvre une partie du golfe de Gascogne (Fig. 1.8). La dynamique du golfe de Gascogne est influencée par la marée, les vents, les apports d'eau douce, la bathymétrie, ainsi que la circulation à grande échelle dans l'Atlantique, donnant lieu à une forte variabilité saisonnière de la circulation générale.

Le golfe de Gascogne est marqué par la séparation du domaine en plateau continental, plaine abyssale et à leur limite, talus continental. La bathymétrie détaillée de la région est donnée en Figure 1.8. Le talus continental est traversé par de nombreux canyons et est particulièrement raide sur la majeure partie de la région. Ce talus est très rapproché de la côte nord de l'Espagne où il suit la direction zonale. Ainsi l'isobathe 1000 m au large de Capbreton (face à la côte landaise) n'est qu'à 3 km de la côte. Le plateau continental présente quant à lui une pente faible et s'étend vers le nord du domaine. Il est relativement éloigné des côtes françaises tandis que ce dernier est étroit au niveau des côtes espagnoles (entre 20 et 50 km de large). La plaine abyssale se situe au centre du domaine, à près de 4000 m de profondeur (elle peut dépasser les 5000 m de profondeur en certains endroits).

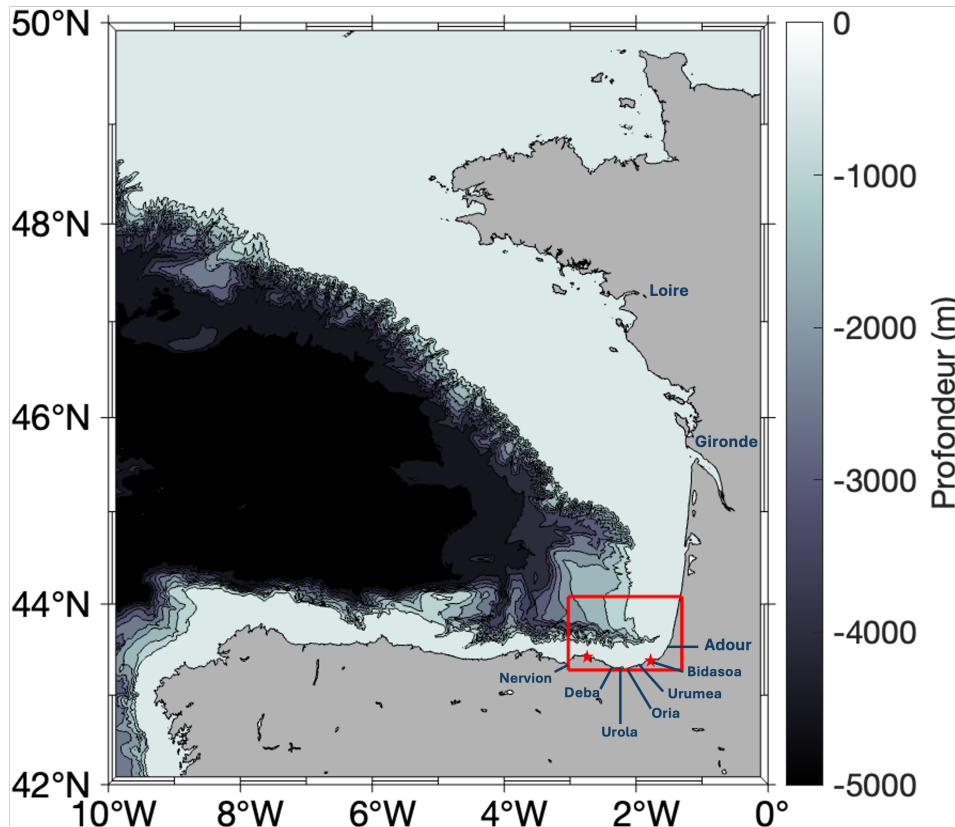


FIGURE 1.8 – Carte des profondeurs du golfe de Gascogne. Les données bathymétriques proviennent de la grille 2023 GEBCO (General Bathymetric Chart of the Oceans) ([https://www.gebco.net/data\\_and\\_products/gridded\\_bathymetry\\_data/](https://www.gebco.net/data_and_products/gridded_bathymetry_data/)). Les noms des principales rivières sont indiqués en bleu et les localisations des deux antennes radar sont données par des étoiles rouges. La région d'étude, le sud-est du golfe de Gascogne, est indiquée par un rectangle rouge.

Les cartes de circulation moyenne saisonnière dans le golfe de Gascogne sont présentées en Figure 1.9. Dans cette région, la marée provoque des courants alternatifs dont la période est de 12h25 (VANDERMEIRSCH et al., 2012). Ces courants sont particulièrement faibles lorsque les profondeurs sont importantes (moins de 10 cm/s) et gagnent en intensité sur le plateau continental. Ils présentent également une composante non périodique, connue sous le nom de courant résiduel, qui reste généralement faible en moyenne dans le golfe de Gascogne. Toutefois, à une échelle locale, ces courants peuvent prendre de l'importance, voire devenir dominants. Dans le sud-est du golfe de Gascogne spécifiquement, les courants de marée sont faibles avec des valeurs de vitesse inférieures à 10 cm/s pour M2, qui est la composante la plus énergétique (PAIRAUD et al., 2008 ; RUBIO et al., 2013).

Les courants persistants à l'échelle de quelques jours sont principalement induits par le vent et sont très variables selon la profondeur. Ces courants possèdent une variabilité saisonnière directement liée aux situations météorologiques. La circulation induite par le vent est également susceptible de générer des mouvements verticaux près des côtes, appelés upwelling et downwelling.

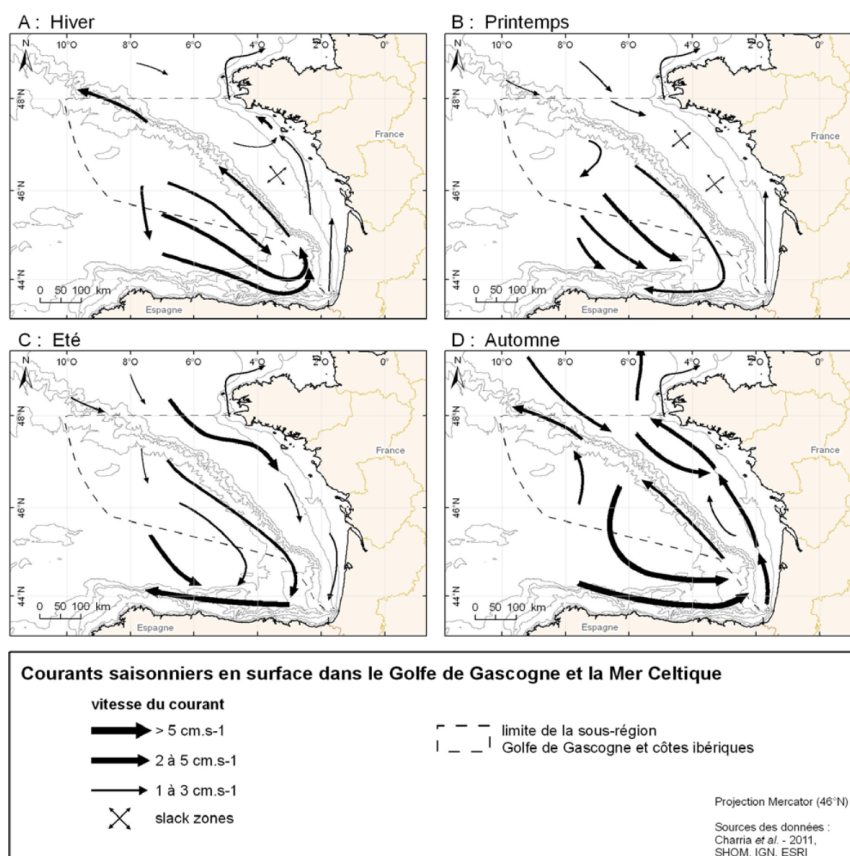


FIGURE 1.9 – Circulation moyenne saisonnière en surface dans le golfe de Gascogne (Sources : CHARRIA et al., 2013, IGN, ESRI, SHOM, OSPAR, 2010).

La circulation globale dans le golfe de Gascogne présente un schéma complexe (Fig. 1.9). Le long de la pente, le Courant Polaire Ibérique, une composante régionale du système de courants de la frontière orientale de l'Atlantique Nord, s'écoule vers l'est, au large de la côte espagnole, et vers le nord, au large de la côte française. En automne et en hiver, il présente un flux relativement intense vers le nord (20-50 cm/s à la surface), transportant des eaux chaudes du coin sud-est du golfe et se traduisant par une langue chaude qui s'étendant du Pays Basque à la Bretagne. Au printemps et en été, le Courant Polaire Ibérique s'affaiblit et devient plus variable, se déplaçant principalement vers l'est (CHARRIA et al., 2013; RUBIO et al., 2019; SOLABARRIETA et al., 2014; VANDERMEIRSCH et al., 2012).

## 1.4.2 Contexte météorologique

### *Le régime des vents*

Les courants induits par le vent sont observés sur une large gamme d'échelles temporelles dans le golfe de Gascogne (FONTÁN et al., 2013; FONTÁN et al., 2009). Sur le plateau continental, la circulation de l'eau est principalement déterminée par le vent et est fortement influencée par l'orientation de la ligne de côte.

En automne et en hiver, les vents sont principalement du sud-ouest et génèrent une dérive vers le nord et vers l'est sur le plateau. Au printemps, le régime des vents passe au nord-est,

entraînant des courants de haute mer vers l'ouest-sud-ouest, le long de la côte espagnole. La situation estivale est similaire à celle du printemps, mais la faiblesse des vents et la plus grande variabilité de la direction de la dérive générale rendent les courants plus incertains (GONZÁLEZ et al., 2004; SOLABARRIETA et al., 2015).

### *Le régime des houles*

Au nord du golfe de Gascogne, les hauteurs moyennes de houles les plus élevées ont lieu l'hiver, en raison du passage des perturbations. Les hauteurs augmentent vers le large pour atteindre 3 m. L'été, les hauteurs significatives de la houle restent le plus souvent inférieures à 1 m près des côtes, et inférieures à 2 m au large. La houle d'ouest domine sur le nord du domaine, puis la direction ouest nord-ouest s'impose en hiver comme en été en descendant vers le sud de la région (MAISONDIEU, 2017; VANDERMEIRSCH et al., 2012).

Dans la partie sud du golfe de Gascogne, les hauteurs moyennes les plus élevées ont lieu elles aussi durant l'hiver. Elles augmentent vers le large pour atteindre des valeurs égales ou supérieures à 3 m. L'été, près des côtes, les hauteurs de houle restent en général inférieures à 1 m. Au large, elles restent inférieures à 2 m. La houle de nord-ouest domine en hiver comme en été. Les hauteurs significatives de houle importantes sont peu fréquentes mais significatives. (MAISONDIEU, 2017; VANDERMEIRSCH et al., 2012)

### *Les apports fluviaux*

La Loire et la Gironde sont les deux principaux fleuves de la côte atlantique française et peuvent induire des courants de densité importants (d'environ 10 cm/s). Sur la côte espagnole, au Pays Basque, les rivières ont un caractère torrentiel, avec des décalages très courts entre les précipitations et le débit de la rivière qui en résulte. Les plus hauts débits se produisent principalement au printemps et à l'automne, tandis que les faibles débits fluviaux se produisent à la fin de l'été.

Les principaux cours d'eau qui drainent le Pays Basque espagnol sont le Nervión, l'Oria et la Bidassoa, avec des débits annuels moyens de 20 à 30 m<sup>3</sup>.s<sup>-1</sup> (FERRER et al., 2009). Cependant, l'influence la plus évidente de l'eau douce sur le plateau continental basque provient de l'Adour (France), dont le débit annuel moyen est supérieur à 300 m<sup>3</sup>.s<sup>-1</sup> (BRUNET & ASTIN, 1999; PUILLAT et al., 2004); il s'agit de la principale source d'eau continentale dans ce secteur du golfe de Gascogne (VALENCIA et al., 2004).

## **1.5 Données disponibles dans le golfe de Gascogne**

### **1.5.1 Vitesses des courants mesurées par les bouées et ADCP**

Deux campagnes ont été menées dans les eaux côtières du sud-est du golfe de Gascogne : l'étude BOBLIT0.1 au printemps (ci-après dénommée B1) et l'étude BOBLIT0.2 à l'automne (ci-après dénommée B2). Ces deux études ont été menées dans la même zone géographique, où des structures de convergences de déchets marins visibles à l'oeil nu avaient été signalées.

Au cours de l'étude B1, 23 bouées dérivantes de surface (13 avec ancre flottante, 10 sans ancre flottante) ont été déployées dans trois clusters différents à l'intérieur de structures de

convergence visibles où une agrégation de mousse et de macro-déchets a été observée qualitativement. Trois bouées dérivantes avec ancre flottante ont été déployées sur un transect longitudinal à  $-2.4^{\circ}\text{E}$ , dix autres bouées (5 avec ancre flottante, 5 sans ancre flottante) ont été déployées dans un cluster centré sur  $43.4^{\circ}\text{N}$ ,  $-2.25^{\circ}\text{E}$  et les dix dernières (5 avec ancre flottante, 5 sans ancre flottante) ont été déployées dans un autre cluster centré sur  $43.35^{\circ}\text{N}$ ,  $-2.3^{\circ}\text{E}$ . Dans le cluster le plus à l'ouest, la distance moyenne de séparation entre les bouées dérivantes était de 10 km. Dans celui du milieu, la distance moyenne était de 1.7 km et dans celui le plus à l'est, elle était de 0.6 km (Fig. 1.10). Les caractéristiques des différentes campagnes de mesures sont présentées dans le tableau 1.2.

Au cours de l'étude B2, aucune zone de convergence n'était visible. Ainsi, 9 bouées dérivantes avec ancre flottante ont été déployées dans une zone rectangulaire (15 km de long et 12 km de large), avec une séparation initiale de 3 km entre les paires de bouées (Fig. 1.10).

En plus des bouées dérivantes, les vitesses de la couche de surface (1 m de profondeur) ont été mesurées à l'aide d'un ADCP (Acoustic Doppler Current Profiler, RDI 600 kHz) remorqué sur un transect, réalisé au cours de la campagne B2. La trace ADCP a duré au total quatre heures avec une fréquence d'échantillonnage de 1 min et a couvert une zone de 9 km x 6 km, dans la partie sud du domaine ( $43,3^{\circ}\text{N}$ ,  $2,3^{\circ}\text{E}$ , Fig. 1.10 - trajectoire verte).

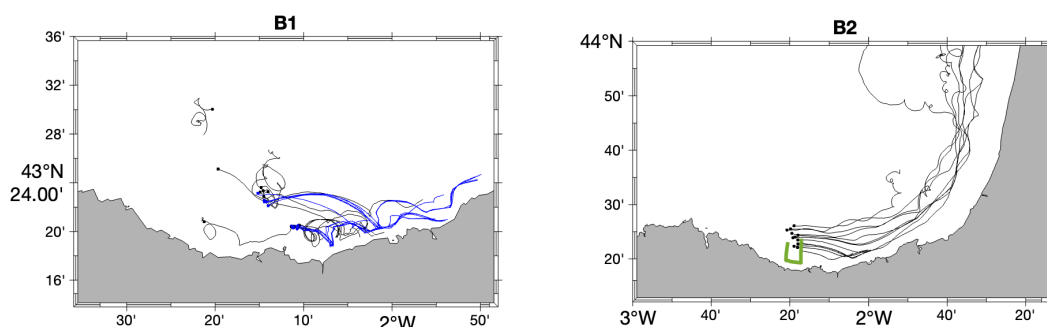


FIGURE 1.10 – Trajectoires des bouées lagrangiennes au cours des différentes campagnes (23 durant B1 et 13 durant B2). La localisation du déploiement est symbolisée par un point. Les zooms ont été ajustés pour fournir la meilleure représentation des trajectoires. Durant B1, les bouées avec ancre flottante sont représentées en noir tandis que les bouées sans ancre flottante sont représentées en bleu. Le transect ADCP réalisé pendant B2 est représenté par la trajectoire verte.

TABLE 1.2 – Tableau récapitulatif des différentes campagnes menées dans le sud-est du golfe de Gascogne. Les dates et heures sont données en UTC.

Campagnes	Bouées	Lâcher	Récupération	Marée	Vent
B1	23	26/04/22 11h45	28/04/22 07h00	Moyenne (55)	Nord-est (2 m/s)
B2	9	24/10/22 14h00	02/11/22 14h40	Forte (95)	Nord-ouest (4 m/s)

Les bouées sans ancre flottante sont fabriquées à l'AZTI avec un récipient en plastique de 9 cm de haut et de 9.5 cm de diamètre, pesant environ 200 grammes. La description détaillée de ces bouées est présentée dans l'article de RUIZ et al. (2022).

Concernant les bouées avec ancre flottante, trois types de bouées ont été utilisées : les bouées de surface côtières Nomad, de forme cylindrique, fabriquées par SouthTek (<https://www.southteksl.com/>), les bouées de surface ODI, anciennement fabriquées par Albatros Marine Technologies, et un ensemble de bouées dérivantes de forme cylindrique fabriquées au Laboratoire d'Océanologies et Géosciences (LOG), utilisées également en Manche orientale, dont les caractéristiques sont présentées au chapitre 1.3.1.

### 1.5.2 Vitesses radar HF

Les champs horaires de courant de vitesse de surface utilisés dans cette étude proviennent du radar HF d'EuskOOS, composé de deux stations CODAR Seasonde situées le long de la côte du Pays Basque espagnol dans le sud-est du golfe de Gascogne (Fig. 1.8). Le réseau radar HF d'EuskOOS (<https://doi.org/10.57762/T4WH-DQ48>) fait partie de JERICO (<https://www.jerico-ri.eu/>) et est exploité conformément aux normes et recommandations de JERICO (MANTOVANI et al., 2020; RUBIO et al., 2018; SOLABARRIETA et al., 2016). Les deux stations radar individuelles, fonctionnant à une fréquence de 4.86 MHz, permettent de mesurer la vitesse dans un rayon de 200 km avec une résolution de 5 km le long du faisceau et une résolution angulaire de 5°. Les données de vitesse sont obtenues à un pas de temps d'une heure et distribuées dans le cadre des produits Copernicus Global Ocean - courants océaniques de surface in situ en temps réel et en temps différé (<http://doi.org/10.48670/moi-00041> & <http://doi.org/10.17882/86236>).

En utilisant les vitesses radiales mesurées, deux produits de courants de surface ont été générés, couvrant la zone de -3°E à -1.3°E et de 43.3°N à 44°N. D'une part, la méthode gap-filling de l'analyse modale à frontière ouverte (OMA, KAPLAN et LEKIEN, 2007) a été appliquée aux vitesses radiales pour obtenir les courants OMA horaires avec une résolution spatiale de 5 km pour les périodes avril-mai 2022 et octobre-novembre 2022. Seules les valeurs OMA avec des erreurs de dilution géométrique de la précision (GDOP) inférieures au seuil de 8 cm/s (SOLABARRIETA et al., 2014) ont été prises en compte pour cette étude.

D'autre part, la méthode d'interpolation variationnelle bi-dimensionnelle 2dVar (YAREMCHUK & SENTCHEV, 2009) a été appliquée aux vitesses radiales pour obtenir des champs de courant horaires avec une résolution spatiale de 2.5 km. 2dVar est une méthode d'interpolation non locale qui permet de reconstruire les vecteurs vitesse des courants de surface en chaque point d'une grille régulière en utilisant une combinaison de toutes les vitesses mesurées. Des estimations de l'erreur relative d'interpolation ont été calculées à chaque pas de temps. Les champs 2dVar de vitesse ont été générés pour une période étendue, allant de janvier 2020 à décembre 2022, car ils ont également été utilisés comme données historiques pour la méthode d'optimisation (présentée au chapitre 2.2).

### 1.5.3 Vitesses modélisées par IBI

Les simulations numériques utilisées dans cette étude proviennent du produit Iberia-Biscay-Ireland (IBI) disponibles sur le catalogue du Copernicus Marine Environment Service (CMEMS) (<https://doi.org/10.48670/moi-00027>). IBI est un produit basé sur l'application d'un modèle

NEMO-v3.6 à une résolution horizontale de  $1/36^\circ$ , Mercator-Ocean étant chargé du développement du code du modèle. Ce produit couvre les eaux du nord-ouest de l'Europe et fournit des données océanographiques en trois dimensions à des fréquences quotidiennes et mensuelles. Il propose également des moyennes horaires et des valeurs instantanées toutes les 15 min pour certaines variables de surface. Les données incluent les moyennes quotidiennes et mensuelles des grandeurs suivantes : température, salinité, composantes zonale et méridionale de la vitesse, profondeur de la couche de mélange, température du fond marin et hauteur de la surface de la mer.

Le modèle est basé sur l'hypothèse de l'équilibre hydrostatique et de l'approximation de Boussinesq, et utilise une surface libre explicite non linéaire (SHCHEPETKIN & MCWILLIAMS, 2005) pour simuler correctement les ondes de gravité externes rapides telles que les mouvements de marée. Le mélange vertical est paramétré selon un modèle  $k-\varepsilon$  mis en œuvre sous la forme générique proposée par UMLAUF et BURCHARD (2003). La bathymétrie du modèle IBI est dérivée du jeu de données GEBCO 08 (BECKER et al., 2009) d'une résolution de 30 secondes d'arc.

Les données de la frontière ouverte latérale (température, salinité, vitesses et niveau de la mer) sont interpolées à partir des sorties quotidiennes du système de résolution des tourbillons Copernicus Marine Global à  $1/12^\circ$ . Elles sont complétées par 11 nouvelles harmoniques de marée (M2, S2, N2, K1, O1, Q1, M4, K2, P1, Mf, Mm), construits à partir de la solution FES2014 (Carrere et al., 2015).

Le produit IBI est forcé toutes les heures avec des champs atmosphériques (vent à 10 m, pression de surface, température à 2 m, humidité spécifique à 2 m, taux de précipitations, flux radiatifs d'ondes courtes et d'ondes longues) fournis par le ECMWF (European Centre for Medium-Range Weather Forecasts).

Les apports d'eau douce des rivières dans la zone IBI sont mis en œuvre en tant que condition limite frontière ouverte pour 33 rivières, sources ponctuelles correspondant aux principales rivières présentes dans la zone. Les données de débit sont basées sur des observations quotidiennes (recueillies dans le cadre du projet PREVIMER pour 9 rivières) et sur la climatologie (en utilisant des données climatologiques mensuelles provenant du GRDC - <http://www.bafg.de/GRDC/> - et des données de la "Banque Hydro" française - <http://www.hydro.eaufrance.fr/> ).

Le produit IBI est assimilé par le schéma d'assimilation du filtre de Kalman d'ordre réduit utilisant des données altimétriques (anomalies du niveau de la mer le long de la trajectoire), des profils verticaux de température et de salinité in situ (base de données CORA 4.1) et la température de surface de la mer par satellite haute-résolution (ODYSSEA).

Dans cette étude, les vitesses zonales et méridionales des courants de surface sont extraites des sorties du modèle dans la région, d'intérêt définie ci-dessous (Fig. 1.11), identique à la zone couverte par les radars HF, et durant les deux campagnes de mesures.



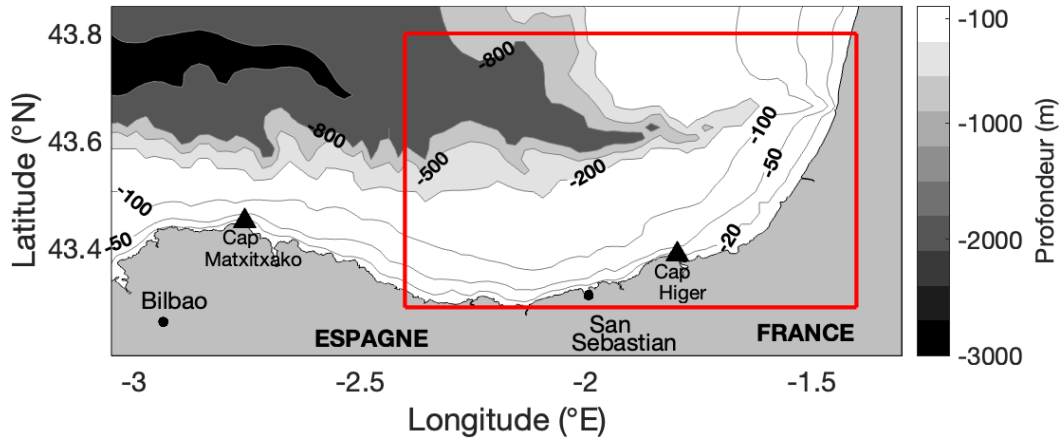


FIGURE 1.11 – Détail bathymétrique du domaine de modélisation. La bathymétrie est représentée en gris. Le rectangle rouge délimite la région où les mesures lagrangiennes ont été effectuées.

### 1.5.4 Données de vent

Les données de vent proviennent du modèle de recherche et de prévision météorologique Weather Research Forecast (WRF) fourni par l'agence météorologique de Galice. La description du modèle est détaillée dans SKAMAROCK et al. (2019). Dans cette étude, les composantes zonales et méridiennes horaires de la vitesse du vent à une hauteur de 10 m ont été utilisées, présentant une résolution spatiale de 4 km, couvrant la zone géographique prédéfinie (RUIZ et al., 2022).

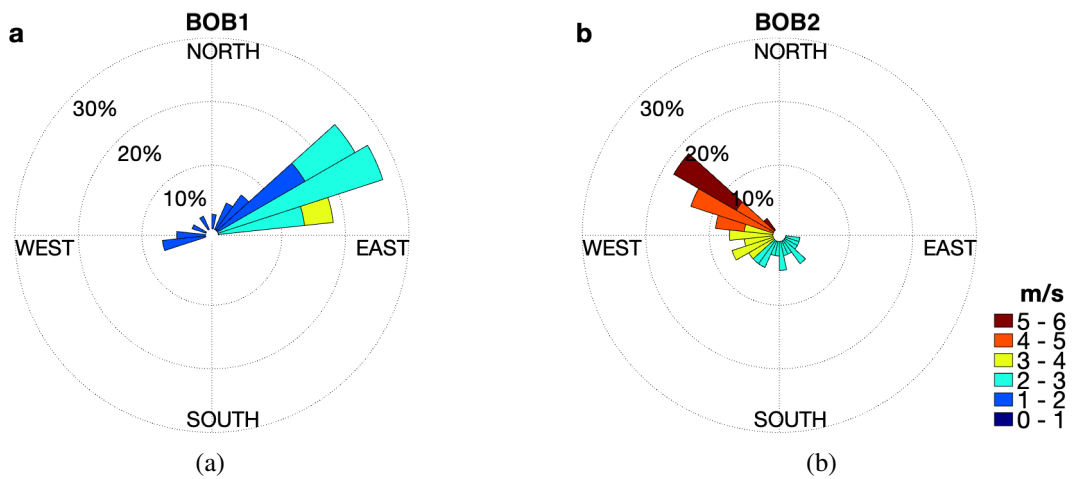


FIGURE 1.12 – Roses des vents provenant du modèle WRF durant chacune des campagnes menées dans le sud-est du golfe de Gascogne.

### 1.5.5 Données satellite

Les estimations de concentration de Chl-a de surface obtenues par télédétection proviennent du catalogue CMEMS (<https://doi.org/10.48670/moi-00286>). Ce produit provient du spectromètre OLCI (Ocean and Land Color Imager) de Sentinel-3.

La concentration de Chl-a utilisée est OC4Me, correspondant à la concentration de pigments basée sur l'algorithme de couleur de l'océan développé pour MERIS, et suivant l'approche de O'REILLY et al. (1998), avec une résolution spatiale de 300 m et une résolution temporelle d'un jour. Le produit de concentration de Chl-a utilisé est pollué par les nuages. Bien qu'il existe divers algorithmes de gap-filling (par exemple, STOCK et al. (2020)), ces méthodes nécessitent généralement des séries temporelles beaucoup plus longues d'images de Chl-a.

## Chapitre 2

# Méthodes

### 2.1 Préambule

Comme énoncé dans le premier chapitre, la Manche orientale et le sud-est du golfe de Gascogne sont deux régions bien différentes nécessitant des stratégies distinctes concernant l'acquisition de données, mais aussi le traitement et l'analyse de ces données.

Concernant les différents calculs d'erreur et de performance, les méthodes utilisées sont similaires. En revanche, la méthode d'optimisation nécessite d'adopter une démarche quelque peu différente car la variabilité des courants marins dans les deux régions d'étude est significativement distincte. La dynamique en Manche orientale est globalement dictée par la marée tandis que dans le sud-est du golfe de Gascogne, de nombreux autres phénomènes entrent en considération (vents, bathymétrie, oscillation inertielle...). Concernant les diagnostics lagrangiens, les disparités entre les régions d'études impliquent aussi d'utiliser différentes métriques.

Ce chapitre présente les méthodes utilisées dans cette thèse et les singularités des deux régions d'études.

### 2.2 Interpolation Optimale des mesures de bouées

#### 2.2.1 Théorie

Pour contraindre les champs de courants modélisés ou observés par radar HF, les observations, c'est-à-dire les vitesses des bouées dérivantes de surface, sont interpolées de manière optimale dans l'espace et dans le temps. Cela permet d'obtenir une estimation plus précise de l'état de la dynamique de l'océan en effectuant un ajustement par moindres carrés pondérés d'un champ de modèle d'arrière-plan aux observations. Les observations sont obtenues à des points répartis de manière irrégulière et sont considérées comme imparfaites, chaque observation étant sujette à une incertitude ( $\approx 5$  cm/s). Il est supposé que l'erreur d'observation n'est pas corrélée avec l'erreur des champs de courants modélisés ou obtenus par radar HF.

Dans la méthode d'IO, la correction d'un champ de vitesse d'arrière-plan  $\mathbf{u}(x, t)$ , fourni sur une grille régulière, est effectuée en utilisant une combinaison linéaire des différences pondérées entre les vitesses d'arrière-plan  $\mathbf{u}$  et les vitesses observées  $\mathbf{u}_i^*$  au point  $i$  (BRETHERTON et al., 1976; SENTCHEV et YAREMCHUK, 2015; THIÉBAUX et PEDDER, 1987; MOLCARD et al., 2003).

Les poids choisis pour minimiser la différence quadratique moyenne entre les deux vitesses sont une combinaison des covariances des champs de courants d'arrière-plan et des observations. Avec les matrices de covariance espace-temps du champ de courant d'arrière-plan  $\mathbf{B} = \langle \mathbf{u}(x, t) \mathbf{u}(x', t') \rangle$  et les observations  $\mathbf{R}_{ij} = \langle \mathbf{u}_i^* \mathbf{u}_j^* \rangle$ , les vitesses optimisées  $\mathbf{u}_{opt}$  sont calculées comme suit :

$$\mathbf{u}_{opt} = \mathbf{u} + \sum_{ij} \mathbf{B} \mathbf{H}_j^T (\mathbf{H}_i \mathbf{B} \mathbf{H}_j^T + \mathbf{R}_{ij})^{-1} (\mathbf{H}_i \mathbf{u} - \mathbf{u}_i^*), \quad (2.1)$$

où  $\mathbf{H}_i$  correspond à un opérateur linéaire projetant les valeurs de vitesse aux noeuds de la grille sur le  $i^{\text{ème}}$  point d'observation.

## 2.2.2 Composition des matrices de covariances

Une hypothèse importante sous-jacente à la méthode d'IO est que le champ d'arrière-plan est une bonne approximation de la vérité. Le calcul de  $\mathbf{B}$  et de  $\mathbf{u}$  est donc crucial. Les champs de vitesse d'arrière-plan sont fournis, par un modèle en Manche orientale ou issus de radars HF dans le sud-est du golfe de Gascogne, au pas de temps d'observation.

### 2.2.2.1 Manche orientale

En Manche orientale, la matrice de covariance est composée à partir des champs de courants de surface issus du modèle MARS. Un nombre variable de trajectoires, c'est à dire d'états dynamiques fournis par le modèle, de la même durée que la campagne considérée (26 h pour C1 et 46 h pour C2), appelées membres d'ensemble, est utilisé. Un test de sensibilité au nombre de membres d'ensemble utilisé pour composer la matrice de covariance est présenté au chapitre 3.

Trois approches ont été utilisées pour sélectionner les membres d'ensemble. La première, la plus simple, consiste à les extraire du modèle les jours environnant la campagne de mesure, en respectant les conditions climatiques (conditions de vent et marée similaires à celles observées durant les jours de campagne). Durant C1, 7 membres d'ensemble ont été sélectionnés au total en utilisant cette approche.

Cependant, pour un état océanique évolutif, un grand nombre de membres d'ensembles peut être nécessaire pour représenter l'évolution du champ d'écoulement avec une signification statistique. Par conséquent, dans la deuxième approche, les membres d'ensemble ont été extraits de la simulation du modèle d'une durée d'un an (janvier à décembre 2020) contenant C1, en recherchant des conditions de vent et de marée similaires à celles observées au cours de la période d'observation. En acceptant une variation de la vitesse et de la direction moyennes du vent dans une fourchette de  $\pm 2$  m/s et  $\pm 45^\circ$  respectivement, pour un stade de marée donné, un total de 31 membres d'ensemble a été sélectionné.

Dans un troisième temps, un critère plus restrictif de la plage de variation du vent, par exemple  $\pm 1$  m/s et  $\pm 45^\circ$ , a permis d'obtenir 11 membres d'ensemble. La figure 2.1 montre la chronologie des membres d'ensemble sélectionnés pour l'IO des observations de vitesse pendant C1. Chaque membre d'un ensemble représente une trajectoire de modèle de 26 heures.

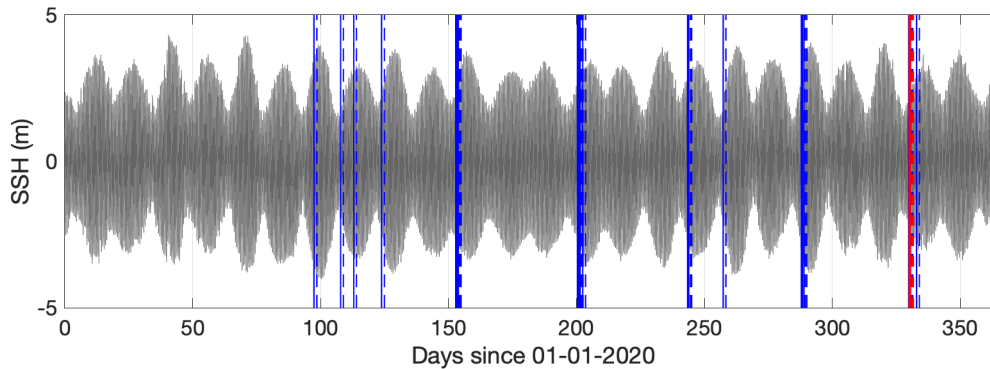


FIGURE 2.1 – Un total de 31 membres d’ensemble (en bleu) extraits de la simulation du modèle de circulation en Manche orientale. La campagne C1 est indiquée en rouge. Le début de chaque membre d’ensemble est représenté par un trait plein tandis que la fin est représentée par un trait en pointillé. La hauteur de la surface de la mer (SSH) provenant de la simulation du modèle est indiquée en gris foncé.

En Manche orientale, le nombre de membres d’ensemble nécessaire pour composer la matrice de covariance modèle est supposément faible compte tenu du fait que la région est largement dominée par les courants de marée, impliquant une corrélation importante entre les différents membres d’ensemble.

### 2.2.2.2 Sud-est du golfe Gascogne

Le sud-est du golfe de Gascogne étant une région où la dynamique est dirigée par de nombreux phénomènes, la sélection des membres d’ensemble a été faite en utilisant l’algorithme de clustering non linéaire K-Means (HASTIE et al., 2009 ; SOLABARRIETA et al., 2015) en utilisant les champs de courants issus des radars HF de janvier à décembre 2022 (méthode présentée au chapitre 2.6.1).

Les principaux objectifs de cette méthode sont doubles : (i) identifier et catégoriser les groupes de courants de surface dominants présents durant les deux périodes d’observations, et (ii) extraire les ensembles avec des groupes dominants similaires dans les données de la série temporelle étendue, qui sont ensuite utilisés pour créer la matrice de covariance requise pour l’IO, représentant 90% de la variabilité.

L’algorithme K-Means utilise un nombre fixe de cluster, qui a été fixé à 12 après un test de sensibilité présenté au chapitre 4.

Simultanément au test de sensibilité pour le nombre de clusters dans la méthode K-Means, un test de sensibilité a également été effectué pour le nombre de membres d’ensemble composant la matrice de covariance de l’IO. Pour construire cette matrice de covariance, les groupes dominants durant chaque période d’observation ont été identifiés. Ensuite, les membres d’ensemble présentant les mêmes groupes prédominants dans un ordre de grandeur similaire ont été extraits pour former les membres d’ensemble et estimer la matrice de covariance. Le nombre de membres d’ensemble optimal obtenu est de 80.

Il est important de noter que chaque membre d'ensemble sélectionné débute au même stade de marée que les données d'observation. Pour éviter toute influence des marées, seuls des membres d'ensemble se synchronisant avec les oscillations des marées ont été sélectionnés, même si les marées représentent une part relativement faible de l'énergie cinétique totale dans la zone d'étude (environ 10 à 15 %, comme indiqué par SOLABARRIETA et al., 2014).

## 2.3 Calcul d'erreur

Afin d'évaluer la fiabilité des différents champs de courants de surface utilisés dans chacune des régions d'étude ainsi que la performance de la méthode d'optimisation, différentes quantités sont calculées.

### 2.3.1 Erreur relative et quadratique moyenne

L'erreur des champs de courants modélisés ou observés par radar HF, ainsi que la qualité du schéma d'IO, est quantifiée en estimant la différence relative moyenne entre les observations  $\mathbf{u}^*$  et les champs de vitesses  $\mathbf{u}$ .

Dans un premier temps, l'erreur relative est estimée à chaque pas de temps, au point  $i$ , comme suit :

$$\varepsilon = \sqrt{\frac{\sum_i (\mathbf{H}_i \mathbf{u} - \mathbf{u}_i^*)^2}{\sum_i (\mathbf{u}_i^*)^2}} \quad (2.2)$$

Dans un second temps, l'erreur quadratique moyenne est estimée à chaque pas de temps comme suit :

$$RMSE = \sqrt{\sum_i (\mathbf{H}_i \mathbf{u} - \mathbf{u}_i^*)^2} \quad (2.3)$$

Ici,  $\mathbf{u}_i^*$  correspond aux vitesses observées par les bouées dérivantes au point  $i$  et  $\mathbf{H}_i$  correspond à un opérateur linéaire projetant les valeurs de vitesse aux noeuds de la grille, sur le  $i^{\text{ème}}$  point d'observation. La matrice de vitesse  $\mathbf{u}$  peut quant à elle correspondre à la fois à des vitesses modélisées, observées par radar HF, et même optimisées.

### 2.3.2 Distance de séparation et distance parcourue

Pour estimer la précision des différents champs de courants de surface utilisés, le logiciel de transport lagrangien OpenDrift (<https://opendrift.github.io/>, voir DAGESTAD et al., 2018 ; JONES et al., 2016) est utilisé pour simuler la dérive de particules virtuelles dans le champ de courant considéré. Des particules virtuelles sont lâchées à la position initiale et au temps initial de chaque bouée dérivante, et advectées en utilisant le schéma de Runge-Kutta d'ordre 4 et l'interpolation bilinéaire, nous renseignant sur leur position  $x(t)$  et  $y(t)$  en fonction du temps. Dans le logiciel OpenDrift, le paramètre de diffusion horizontale a été fixé à 0.

La distance de séparation  $d$ , estimée à un pas de temps de 15 min pour les données modélisées et 1 h pour les données radar HF, est une mesure couramment utilisée démontrant la qualité des trajectoires de dérive simulées avec différents champs de courant. Cette dernière est estimée

à chaque pas de temps comme la distance en kilomètre entre la particule virtuelle et la bouée dérivante considérée.

La distance parcourue est aussi couramment utilisée pour évaluer la fiabilité d'un champ de courants. Elle correspond à la distance parcourue par une particule virtuelle et est comparée à la distance parcourue par une bouée réelle.

### 2.3.3 Erreur lagrangienne

Une autre méthode utilisée dans cette thèse pour estimer les erreurs des différents champs de courants est le calcul de l'erreur lagrangienne.

De la même manière que pour le calcul de la distance de séparation et de la distance parcourue, le logiciel OpenDrift est utilisé pour simuler l'advection de particules virtuelles dans un champ de courant. Ces dernières sont lâchées toutes les heures à l'emplacement des bouées dérivantes réelles.

La précision des trajectoires virtuelles est évaluée à l'aide de l'erreur lagrangienne  $L_\epsilon$  (LIU & WEISBERG, 2011 ; RUIZ et al., 2022). L'estimation de  $L_\epsilon$  implique le calcul de la distance de séparation entre l'emplacement de la bouée réelle et celui de la particule virtuelle se déplaçant dans le champ de vitesse.

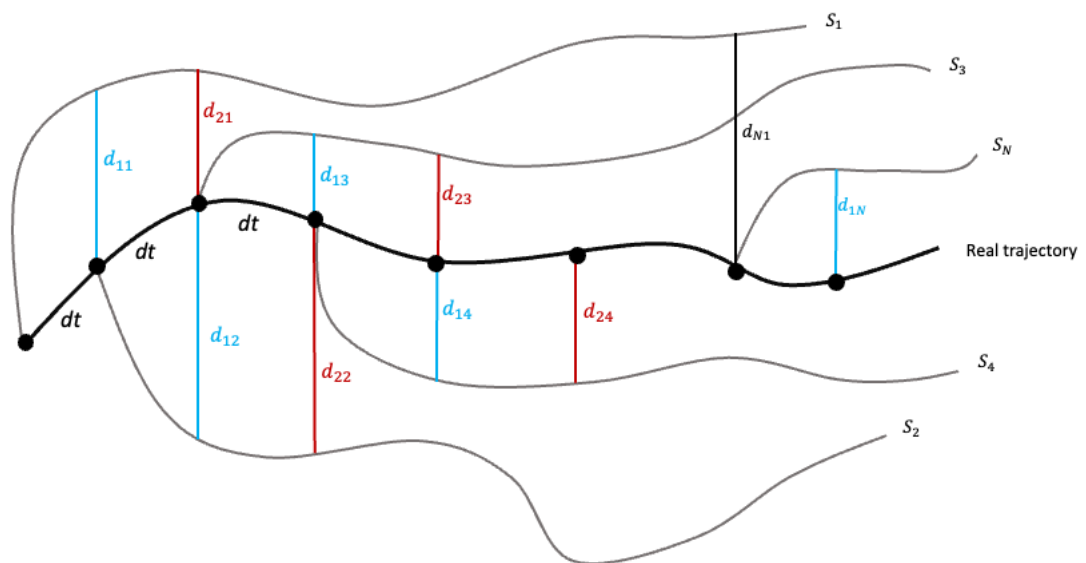


FIGURE 2.2 – Schéma du calcul de l'erreur lagrangienne.

En considérant  $d_{tk}$ , la distance de séparation entre la trajectoire réelle et les  $k$  trajectoires simulées au pas de temps  $t$ , l'erreur lagrangienne est calculée à chaque pas de temps et moyennée sur toutes les trajectoires simulées correspondant à chaque trajectoire de dérive réelle, suivant

l'expression :

$$L_\epsilon(t) = \left\langle \sum_{t=1}^N \sum_{k=1}^{N-(t+1)} \frac{d_{tk}}{N-(t+1)} \right\rangle, \quad (2.4)$$

avec  $N$  correspondant au nombre maximum de pas de temps de déplacement du drifter, qui est aussi égal au nombre de trajectoires simulées (notées  $S_1, S_2, \dots, S_N$  sur la Figure 2.2).

Pour faciliter les comparaisons, l'erreur lagrangienne est divisée par la distance moyenne parcourue des vraies bouées ce qui donne lieu à l'indice sans dimension  $L$ .

## 2.4 Correction de la vitesse induite par le vent

Après divers calculs d'erreur et l'optimisation, il est possible qu'une trajectoire simulée présente encore une distance de séparation significative par rapport à la trajectoire observée. Ce problème soulève des préoccupations quant à la précision de l'estimation de l'effet du vent, notamment dans le modèle MARS, en Manche orientale. La reproduction précise du couplage océan-atmosphère, en particulier dans les régions côtières, reste un défi important. Le vent induit des courants d'Ekman de surface. Cette vitesse induite par le vent peut atteindre 1 à 3 % de la vitesse du vent à une hauteur de 10 m (JENKINS, 1987; WEBER, 1983).

Il est possible d'améliorer la représentation de la vitesse du courant due vent dans le modèle en incorporant des mesures de vitesse provenant de bouées dérivantes. En supposant que la vitesse d'écoulement peut être décomposée en deux composantes - vitesse interpolée de manière optimale ( $\mathbf{u}_{opt}$ ) et une correction supplémentaire,  $\mathbf{c}\bar{\mathbf{U}}_{10}$ , où  $\bar{\mathbf{U}}_{10}$  représente le vecteur vitesse du vent à 10 m de hauteur, moyenné sur la zone d'étude et la période de campagne. Ici,  $\mathbf{c} = \text{diag}(c_x, c_y)$  est une matrice diagonale 2 x 2, et ses éléments diagonaux sont estimés en minimisant la fonction de coût :

$$J(\mathbf{c}) = \left[ \mathbf{X}_k^* - (\mathbf{X}_0 + \sum_k \mathbf{u}_{opt,k} \Delta t + \mathbf{c}\bar{\mathbf{U}}_{10} \Delta t) \right], \quad (2.5)$$

où  $\mathbf{X}^*$  correspond aux coordonnées de la dérive à  $\Delta t = 15min$ ,  $\mathbf{X}_0$  est la coordonnée initiale, et la sommation est effectuée sur une trajectoire de dérive. L'expression entre parenthèses (.) représente une trajectoire de dérive virtuelle après correction de l'effet du vent. Les coefficients ( $c_x, c_y$ ) ont été estimés individuellement pour chaque trajectoire de bouée, puis moyennés. Cette hypothèse repose sur le fait que le vent, en particulier sa direction, est stationnaire. Le champ de courant de surface obtenu après correction de l'effet du vent est appelé  $\mathbf{u}_{cor}$  dans la suite de l'étude.

## 2.5 Diagnostics lagrangiens

Pour traiter de la dispersion turbulente en zone côtière, il est courant de se placer dans le cadre lagrangien et d'utiliser différentes grandeurs permettant de caractériser l'écoulement. Ces grandeurs sont calculées à partir de trajectoires de bouées réelles ou de particules fictives advectées dans différents champs de courants de surface.



### 2.5.1 Dispersion absolue d'un cluster

La dispersion absolue d'un cluster  $A^2$  est définie comme la variance de l'étalement des particules par rapport à la coordonnée moyenne des particules dans un cluster, le barycentre. En deux dimensions, la dispersion peut être estimée de deux manières : le long des axes  $x$  et  $y$  (BERTI et al., 2011 ; ENRILE et al., 2019), ou dans la direction de l'étirement maximal (EMERY & THOMSON, 2004).

Le long des axes  $x$  et  $y$ , elle est calculée en utilisant les coordonnées spatiales de bouées réelles ou particules fictives et s'exprime comme suit (BERTI et al., 2011 ; ENRILE et al., 2019) :

$$a_{ij}^2(t) = \frac{1}{M} \sum_{m=1}^M \{ [x_i^m(t) - \overline{x_i^m(t)}] [x_j^m(t) - \overline{x_j^m(t)}] \} \quad A^2(t) = a_{xx}^2(t) + a_{yy}^2(t), \quad (2.6)$$

avec  $a_{ij}^2(t)$  correspondant à la variance le long des coordonnées spatiales  $i$  et  $j$ ,  $a_{xx}^2$  et  $a_{yy}^2$  correspondant à la dispersion absolue le long de  $x$  et  $y$  respectivement,  $M$  étant le nombre de bouées/particules,  $x^m(t)$  étant la position de la  $m^{\text{ième}}$  bouée/particule au temps  $t$  et  $\overline{x^m(t)}$  étant la coordonnée du barycentre.

L'estimation de la dispersion absolue d'un cluster permet d'identifier deux régimes fondamentaux de dispersion dans l'écoulement turbulent. À court terme, c'est-à-dire pendant les premières heures du déploiement de la bouée dérivante, la croissance de la dispersion absolue suit généralement un régime balistique correspondant à  $A(t)^2 \sim t^2$  (BERTI et al., 2011 ; BOUZAIENE et al., 2021 ; DAVIS, 1983 ; ENRILE et al., 2019 ; POULAIN & NIILER, 1989 ; TAYLOR, 1922). Ce régime se produit lorsque les vitesses des paires de bouées dérivantes sont fortement corrélées, sous l'influence de courants forts et persistants tels qu'un jet côtier. À des échelles de temps plus grandes, lorsque la mémoire de la condition initiale est perdue, et la dispersion absolue peut croître comme  $A(t)^2 \sim t$  (BERTI et al., 2011 ; BOUZAIENE et al., 2021 ; DAVIS, 1983 ; ENRILE et al., 2019 ; POULAIN & NIILER, 1989 ; TAYLOR, 1922).

En fonction de la géographie de la zone côtière étudiée et pour mieux prendre en compte la direction dominante de l'écoulement, la variance est parfois calculée dans la direction de l'étalement maximal des particules et dans la direction perpendiculaire, fournissant deux quantités  $A_1^2$  et  $A_2^2$ . Elles représentent l'axe majeur et l'axe mineur du tenseur de déformation et sont estimées en appliquant l'Analyse en Composantes Principales (ACP) à la distribution des particules à chaque pas de temps (EMERY & THOMSON, 2004). Comme la direction du flux de marée dans les bassins peu profonds est généralement contrainte par la topographie locale et l'orientation du trait de côte, l'orientation de l'ellipse (axe majeur) donne la direction dominante du flux. L'orientation de l'ellipse  $\theta$  et les variances  $A_1^2$  et  $A_2^2$  ont été calculées comme suit (EMERY & THOMSON, 2004) :

$$\theta = \frac{1}{2} \tan^{-1} \left[ \frac{2\overline{x'y'}}{\overline{x'^2} - \overline{y'^2}} \right] \quad (2.7)$$

$$\begin{bmatrix} A_1^2 \\ A_2^2 \end{bmatrix} = \frac{1}{2} \left\{ (\overline{x'^2} + \overline{y'^2})^2 \pm [(\overline{x'^2} - \overline{y'^2})^2 + 4(\overline{x'y'})^2] \right\}, \quad (2.8)$$

avec  $\overline{x'^2}$  et  $\overline{y'^2}$  représentant les variances des coordonnées des particules le long des axes  $x$  vers l'est et  $y$  vers le nord, respectivement.

### 2.5.2 Fonction de densité de probabilité

Pour estimer les zones de passage intense des particules virtuelles dans la région étudiée, la densité de probabilité de passage dans chaque cellule de la grille a été calculée. Cette métrique révèle les zones les plus probables de passage des particules et se calcule comme suit :

$$N(i, j, t) = \sum_{i=1}^{N_i-1} \sum_{j=1}^{N_j-1} \sum_{k=1}^{N_k} \text{Ind}[x(i, j), x(i+1, j)](\text{lon}(t, k)) \cdot \text{Ind}[y(i, j), x(i, j+1)](\text{lat}(t, k)), \quad (2.9)$$

où  $t$  correspond au pas de temps et  $N_t$  au nombre maximal de pas de temps,  $i$  (resp.  $j$ ) correspond à la taille de la grille en longitude (resp. latitude).  $x$  (resp.  $y$ ) correspond aux coordonnées de longitude (resp. latitude) de la grille régulière (du modèle ou des observations par radar HF),  $\text{lon}$  et  $\text{lat}$  correspondent respectivement aux longitudes et latitudes des particules virtuelles.  $\text{Ind}[a, b](x)$  est la fonction indicatrice de l'intervalle  $[a, b]$ , qui vaut 1 si  $x$  est dans l'intervalle et 0 sinon.

### 2.5.3 Exposants de Lyapunov de taille finie

En outre, l'étude des processus de transport turbulent implique la caractérisation des structures cohérentes dans le champ d'écoulement. Le développement de structures turbulentes cohérentes dans l'écoulement (telles que les jets et les tourbillons) est déterminé par la variabilité spatiale et temporelle intrinsèque du champ de courants de surface considéré, ainsi que par les forçages externes tels que le vent ou l'apport d'eau douce qui interagissent avec le champ d'écoulement. Ces structures dynamiques, même si elles présentent une décroissance lente par rapport à la variabilité du champ d'écoulement, dictent une grande partie des caractéristiques de transport (HAZA et al., 2008).

L'estimation des FSLE pour la caractérisation du champ d'écoulement permet d'identifier ces structures de transport, fournissant des informations sur les fluctuations de vitesse et le mélange dans l'écoulement (AURELL et al., 1997; D'OVIDIO et al., 2004; HERNÁNDEZ-CARRASCO et al., 2011b; LACASCE, 2008).

Cette méthode permet d'identifier les Structures Cohérentes Lagrangiennes (LCS) qui organisent le transport dans un champ d'écoulement, en localisant les régions de l'écoulement qui évoluent à des échelles de temps différentes de la variabilité temporelle du champ environnant. Habituellement, ces LCS caractéristiques semblent être entourées par les lignes de grandes valeurs des FSLE, également appelées lignes de crête. Elles identifient les régions d'étirement maximal correspondant aux structures océaniques telles que les fronts, les limites des tourbillons, etc. Comme les LCS ne peuvent pas être traversées par les trajectoires des particules, elles agissent comme une barrière au transport, régissant ainsi le mouvement des particules du fluide environnant autour d'elles. Dans cette étude, le calcul des FSLE se fait à rebours dans le temps car dans ces conditions, les LCS associées ont une interprétation physique directe : elles correspondent à des structures hyperboliques attractives.

Les valeurs des FSLE, données par  $\lambda$ , sont estimées comme l'inverse du temps  $\tau(x)$  nécessaire pour que deux particules de fluide se séparent d'une distance initiale  $\delta_0$  à une distance finale  $\delta_f$ , et s'expriment à la position  $x$  et au temps  $t$  comme suit :

$$\lambda(\mathbf{x}, t, \delta_0, \delta_f) = \frac{1}{\tau(x)} \ln \frac{\delta_f}{\delta_0}. \quad (2.10)$$

Dans cette étude, les valeurs de  $\lambda$  sont calculées à l'aide de l'algorithme décrit dans HERNÁNDEZ-CARRASCO et al. (2011b) avec les paramètres  $\delta_0$  et  $\delta_f$  fixés en tenant compte de la région d'étude.

### 2.5.4 Divergence lagrangienne dans un domaine fini

En complément de la FSLE, une autre quantité lagrangienne utilisée dans cette étude est la FDL D. De la même manière que les FSLE, les lignes de gradient de FDL D calculées à rebours coïncident avec les LCS révélées par les FSLE. Ainsi, utiliser ces deux métriques en complément l'une de l'autre permet de cartographier les structures cohérentes océaniques.

En supposant que le champ de vitesse est divergent, la divergence horizontale ( $\Delta_H \cdot \mathbf{v}$ ) s'accumule le long d'une trajectoire  $s(x_0, y_0, t_0)$  dans le domaine fini. Les valeurs des FDL D sont calculées en intégrant et en moyennant la divergence horizontale dans le temps comme suit (HERNÁNDEZ-CARRASCO, ORFILA et al., 2018) :

$$FDL D(x_0, y_0, t_0, t_f) = \frac{1}{t_f - t_0} \int_{t_0}^{t_f} \Delta_H \cdot \mathbf{v}(x(t), y(t), t) dt, \quad (2.11)$$

avec  $t_f - t_0$  l'intervalle de temps de l'intégration de la trajectoire et  $x_0$  (resp.  $y_0$ ) la longitude (resp. latitude) de la trajectoire considérée.

## 2.6 Méthodes de clustering

Les méthodes de clustering constituent une alternative aux méthode d'analyse des séries temporelles plus fréquemment utilisées, avec pour principal avantage de traiter les relations non linéaires (SOLABARRIETA et al., 2015). Elles permettent, à partir de séries temporelles consécutives, d'identifier modes de dynamique similaire pour la méthode K-Means ou encore d'identifier des régions de dynamique similaire pour la méthode Self-Organizing Maps (SOM).

### 2.6.1 K-Means

L'algorithme K-Means est utilisé dans cette étude dans le cadre de la méthode d'optimisation (chapitre 2.2). Cet algorithme divise l'espace de données, ici les champs de courants de surface, en un certain nombre de groupes. Chaque groupe est défini par un champ de courant de surface moyen dans l'espace et est formé par les données dont le champ de courant de surface est le plus similaire avec celui du groupe (HASTIE et al., 2009). Les 12 groupes extraits dans le sud-est du golfe de Gascogne sont présentés en figure 2.3.

Une base de données est composée de vecteurs à  $N$  dimensions  $\mathbf{X} = x_1, x_2, \dots, x_N$ , où  $N$  est le nombre total de pas de temps. Le nombre de groupes sélectionnés pour obtenir la classification des configurations spatiales à l'aide de l'algorithme K-Means est généralement fixé de manière subjective, les critères de sélection du nombre approprié de groupes n'étant pas encore connus (GUANCHE et al., 2014).

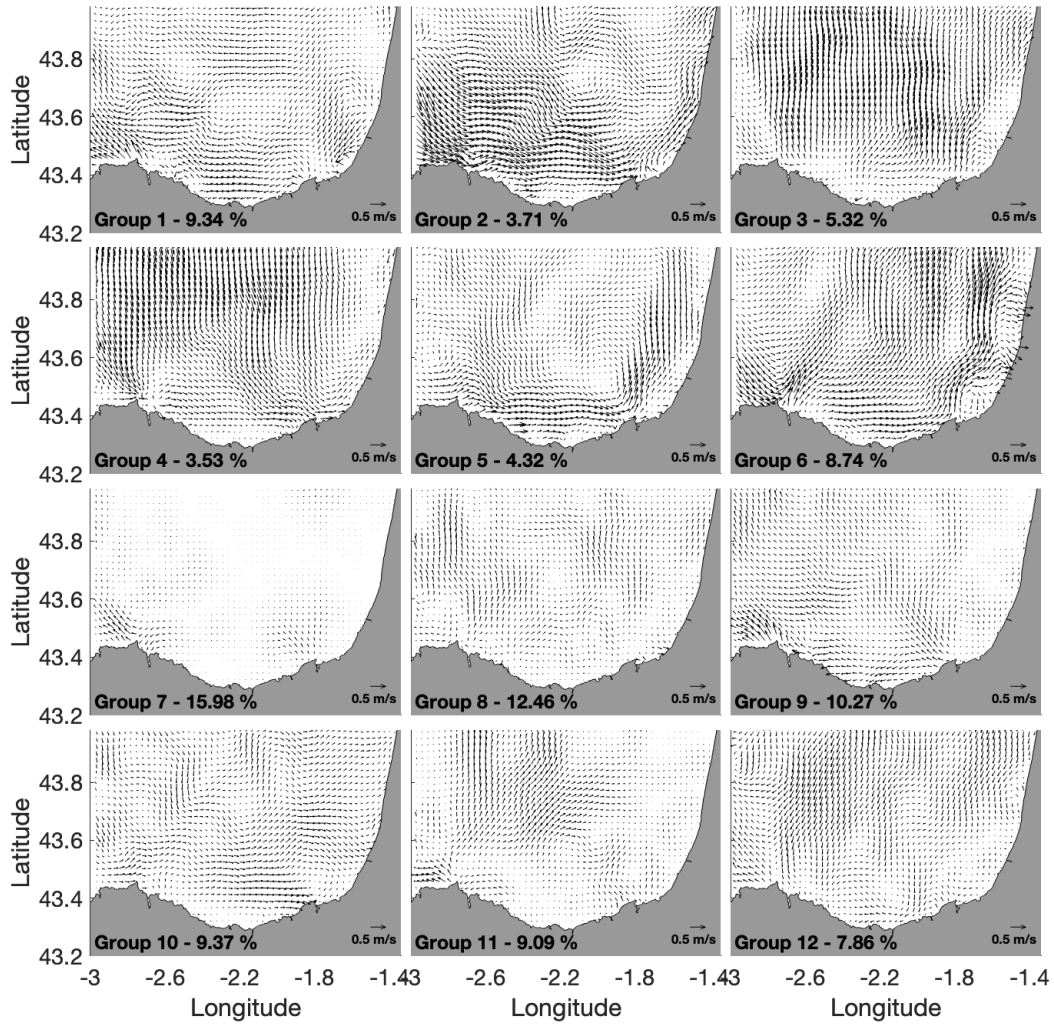


FIGURE 2.3 – Cartes 4 x 3 de l’algorithme K-Means appliqué aux courants de surface HFR de janvier 2020 à décembre 2022. Le pourcentage de variabilité expliqué par chaque groupe est indiqué dans le titre. 12 groupes ont été utilisés.

### **2.6.2 Self-Organizing Maps**

La méthode de clustering SOM est utilisée dans cette étude pour distinguer des régions présentant des régimes dynamiques différents dans le but de comprendre la connectivité dans la région d'étude. De la même manière que l'algorithme K-Means, cette méthode permet d'extraire les variabilités similaires entre différentes régions hydrodynamiques en se basant sur de longues séries temporelles de champs de courants de surface.

Les cartes SOM sont basées sur un réseau neuronal d'apprentissage concurrentiel non supervisé qui est utilisé pour extraire des modèles dans de grands ensembles de données sans nécessiter d'hypothèses sur la distribution des données (KOHONEN, 1982). En tant que réseaux neuronaux artificiels non supervisés, la phase d'apprentissage est un processus au cours duquel le réseau apprend à classer les échantillons en reconnaissant différents modèles.

Le processus d'apprentissage SOM consiste à introduire des champs de vitesse dans un réseau neuronal, qui subit des modifications par le biais d'une procédure itérative. Chaque neurone est caractérisé par un vecteur de poids, correspondant aux dimensions de l'échantillon de données d'entrée. À chaque itération, le neurone dont le vecteur de poids est le plus proche du vecteur de données d'entrée, mesuré par la distance euclidienne minimale, est ajusté avec ses neurones voisins selon une relation de voisinage spécifiée définie par une fonction mathématique. À la fin du processus d'apprentissage, l'algorithme SOM approxime la fonction de densité de probabilité des données d'entrée, en attribuant à chaque neurone un modèle de référence.

Dans cette étude, la technique SOM est appliquée dans le domaine temporel afin de classer les régions en fonction de la variabilité spatiale des courants de surface. L'ensemble des données d'entrée consiste en une série temporelle de trois ans de vitesses de courant de surface, normalisée avant de commencer le processus d'apprentissage. L'algorithme SOM fournit ensuite une carte des différentes sous-régions caractérisées par une variabilité temporelle particulière. Le nombre de neurones a été fixé à six,  $2 \times 3$ , pour distinguer six régions avec des comportements dynamiques distincts.

## Chapitre 3

# Dispersion turbulente en Manche orientale

### 3.1 Introduction

#### *Motivations de l'article*

La dynamique côtière de la Manche orientale est principalement due au forçage de la marée. Cependant, la complexité de la bathymétrie et les vents dominants jouent également un rôle crucial. La combinaison de ces facteurs génère la variabilité des courants et divers phénomènes de circulation locale, influençant fortement la dispersion et le transport de matière particulaire. Ces processus structurent notamment les écosystèmes marins de la région.

Dans les études précédentes, la circulation à fine échelle en Manche orientale n'a pas été complètement caractérisée. En effet, les caractéristiques de la circulation océanique à sous-mésoéchelle sont difficiles à mesurer avec les systèmes d'observation existants et, lorsqu'elles le sont, leur résolution correspond rarement à celle obtenue par la modélisation. De plus, en raison de la complexité de l'environnement océanique côtier, les modèles numériques de circulation à haute résolution ont des difficultés à simuler avec précision les flux très turbulents et les processus de dispersion, en particulier dans les bassins de marée très énergétiques tels que la Manche orientale. Par ailleurs, peu d'observations à haute résolution spatio-temporelle et de simulations avec une bonne précision sont disponibles dans cette région.

Dans ce contexte, le lâcher de bouées dérivantes offre une réelle opportunité pour réaliser une analyse de la dynamique des eaux côtières et leur évolution spatio-temporelle, dans le but de caractériser les processus contrôlant le transport et la dispersion de matière particulaire.

#### *Méthodes d'analyse*

La Manche orientale dispose de données du modèle numérique de circulation MARS, et en particulier des données de courants barotropes 2D et des niveaux d'eau modélisés dans le cadre du projet d'océanographie côtière opérationnelle MARC.

Les vitesses des courants de surface issues des mesures des bouées de surface ont été utilisées pour optimiser le modèle de circulation océanique côtière à haute résolution MARS, comme dé-

taillé dans le chapitre 2.2. En utilisant l'IO, les vitesses de bouées sont interpolées dans le temps et l'espace afin de contraindre le modèle de circulation MARS. L'erreur de la reconstruction du champ de vitesse a été évaluée à chaque pas de temps. Une attention particulière a été portée à la sélection des membres d'ensemble nécessaires à l'estimation de la matrice de covariance des vitesses modélisées. Cette matrice est utilisée dans la fusion de données et peut influencer le résultat de l'optimisation.

En utilisant le logiciel de transport lagrangien OpenDrift, des particules virtuelles ont été déployées et advectées par les courants de surface résultants de la méthode d'optimisation. Ces dernières ont dans un premier temps été déployées aux positions initiales des bouées réelles afin d'évaluer la capacité du modèle optimisé à reconstruire des trajectoires réalistes. Ainsi, les trajectoires réelles ont été comparées aux trajectoires simulées obtenues en advectant les particules virtuelles dans les champs de courants de surface MARS initial et optimisé.

Le calcul de la distance de séparation entre ces différentes trajectoires a permis de mettre en évidence la difficulté du modèle à reproduire correctement l'effet du vent. Ainsi, une méthode de correction du courant induit par le vent a été appliquée, permettant d'obtenir un champ de courant de surface corrigé.

Dans un second temps, les calculs précédents ayant établi la fiabilité du champ de courant de surface corrigé, deux clusters de particules virtuelles ont été déployés et advectés à l'aide du logiciel OpenDrift. Ces deux clusters ont été déployés au nord et au sud du Cap Gris-Nez, une zone très turbulente de la région d'étude, afin de calculer la dispersion, qui reflète la turbulence en Manche orientale. Cette approche permet de mieux comprendre les mécanismes de dispersion des particules et la dynamique turbulente dans cette région d'étude.

#### *Résultats*

Les résultats de cette étude sont multiples. D'une part, la fusion des sorties du modèle avec les mesures de vitesse des bouées dérivantes permet une amélioration significative de la capacité du modèle à simuler la dérive des traceurs passifs en Manche orientale. En effet, cette fusion de données hétérogènes a permis de diminuer de moitié l'erreur entre les vitesses modélisées et les vitesses observées. Cette nette amélioration a aussi été observée lors du calcul des distances de séparation entre les trajectoires réelles indépendantes et les trajectoires simulées avec les champs de courants de surface initiaux et optimisés.

Dans un second temps, cette étude a permis de mettre en évidence le fait que la paramétrisation du vent dans les modèles de circulation océanique côtière est cruciale, en particulier pour des applications telles que la recherche et le sauvetage en mer. Effectivement, la méthode de correction de la vitesse induite par le vent a permis de réduire d'avantage les erreurs et distances de séparations entre les trajectoires réelles et simulées.

Ensuite, la dispersion des particules, estimée à partir de vitesses plus réalistes, obtenues la fusion des données lagrangiennes et du modèle MARS et après la correction du courant induit par le vent, s'est avérée 20% plus élevée au nord du Cap Gris Nez et 13% plus faible au sud du Cap Gris Nez, par rapport au run du modèle initial. La variabilité spatiale de la dispersion a été identifiée. Cette dernière est supposément liée aux caractéristiques à petite échelle de la circulation locale générées par l'interaction du flux de marée avec le Cap Gris Nez et la topographie

irrégulière, à l'origine d'un tourbillon de marée au nord du Cap Gris Nez apparaissant durant le flot.

Les implications de ces résultats sont importantes, car ils peuvent être utilisés pour améliorer les outils d'aide à la décision existants ou pour concevoir de nouveaux outils de surveillance du transport et de la dispersion dans l'environnement océanique côtier avec des applications telles que la pollution, la recherche et le sauvetage en mer ou la surveillance des écosystèmes marins.

Différents aspects de l'étude, brièvement présentés ci-dessus, ont été développés plus en détail dans l'article de recherche : « *Fusion of Lagrangian drifter data and numerical model outputs for improved assessment of turbulent dispersion* » publié dans la revue Ocean Science. doi : <https://doi.org/10.5194/egusphere-2024-176>



**3.2 Article : *Fusion of Lagrangian drifter data and numerical model outputs for improved assessment of turbulent dispersion***



# Fusion of Lagrangian drifter data and numerical model outputs for improved assessment of turbulent dispersion

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**Abstract.** Transport and dispersion processes in the ocean are crucial, as they determine the lifetime and fate of biological and chemical quantities drifting with ocean currents. Due to the complexity of the coastal ocean environment, numerical circulation models have difficulties to accurately simulate highly turbulent flows and dispersion processes, especially in highly energetic tidal basins such as the eastern English Channel. A method of improving the results of coastal circulation modeling and tracer dispersion in the Dover Strait is proposed. Surface current velocities derived from Lagrangian drifter measurements in November 2020 and May 2021 were optimally interpolated in time and space to constrain a high-resolution coastal circulation MARS model, with careful attention given to selecting ensemble members composing the model covariance matrix. The space–time velocity covariances derived from model simulations were utilized by the optimal interpolation algorithm to determine the most likely evolution of the velocity field under constraints provided by Lagrangian observations and their error statistics. The accuracy of the velocity field reconstruction was evaluated at each time step. The results of the fusion of model outputs with surface drifter velocity measurements show a significant improvement (by  $\sim 50\%$ ) of the model capability to simulate the drift of passive tracers in the Dover Strait. Optimized velocity fields were used to quantify the absolute dispersion in the study area. The implications of these results are important, as they can be used to improve existing decision-making support tool or design new tools for monitoring the transport and dispersion in a coastal ocean environment.

## 1 Introduction

Numerous studies have emphasized the significance of sub-mesoscale  $O$  (1–10 km) variability of ocean circulation, which appears highly energetic and ageostrophic. Such sub-mesoscale motions have a notable impact on energy cascade and energy dissipation in the ocean (Ferrari and Wunsch, 2009), as well as on horizontal transport of suspended matter (Aleskerova et al., 2019) or budgets of physical and biological quantities (Uchida et al., 2020). Keerthi et al. (2022) demonstrated that the annual changes in phytoplankton biomass in the Gulf of Mexico are driven by small-scale physical processes (eddies, atmospheric storms, etc.), which control growth and spatial distribution of phytoplankton, and are influenced by the exchange of energy and matter between the atmosphere and the ocean. Both models and observations indicate that the dispersal rate in the presence of sub-mesoscale turbulence can easily exceed the mesoscale dispersal rate in the geostrophic current by an order of magnitude (Haza et al., 2008; Poje et al., 2014).

Incomplete knowledge of forcings in combination with the complexity of coastal environments, which includes a complex shoreline, river mouths, beaches, submarine banks, etc., presents a real challenge for numerical modeling. As a result, circulation models have difficulties in simulating a highly turbulent coastal flow at sub-mesoscale. Hence, it is important to develop techniques that can improve the model skill to reconstruct the water circulation and dispersion processes in coastal environments in a simple and efficient way.

The current study employs a method of optimal interpolation (OI) of Lagrangian observations using a high-resolution

regional circulation model as a background. Lagrangian observations of current velocities are used to correct the model trajectories in an optimal way. Pioneered by Gandin (1963) and applied in atmospheric modeling, the OI has been widely used in different fields of geosciences for mapping the sea surface temperature (Bretherton et al., 1976), modeled current velocity optimization (Molcard et al., 2003; Sentchev and Yaremchuk, 2015), and topography optimization (Wu et al., 2021). Compared to other approaches to optimizing ocean circulation such as variational methods (e.g., Kalnay, 2002; Sentchev and Yaremchuk, 1999; Wikle, 2005), the OI has several advantages. Firstly, the method is straightforward to implement and ensures a reasonable balance between the computational complexity and statistical consistency of the model–data misfits. Secondly, the accuracy of the reconstructed velocity field can be inexpensively evaluated at every time step of the model.

The use of OI of observations leads to a significant improvement of the current velocity fields and velocity gradients, which are often inadequately represented in the models due to their low resolution or intrinsic limitations. Therefore, the turbulent dispersion also appears to be affected by these limitations. Many studies have focused on the investigation of processes that influence the dispersion in the ocean, such as tidal motions (Meyerjürgens et al., 2020), waves (Weichman and Glazman, 2000), and the variability of ocean circulation (Haza et al., 2008; LaCasce and Ohlmann, 2003; Lumpkin and Elipot, 2010). The present study aims to quantify the effect of current velocity optimization on the dispersion rate of passive tracers in a tide-dominated region: the Dover Strait, in the eastern English Channel (EEC).

The paper is organized as follows: Sect. 2 provides a general presentation of the study area and the data used. Section 3 provides a detailed description of the methods utilized in this study. The results of OI of Lagrangian measurements and characterization of dispersion processes are presented in Sect. 4. Furthermore, a technique for correcting the wind-driven velocity component of surface currents is proposed in Sect. 4. Discussion and conclusions are presented in Sects. 5 and 6, respectively.

## 2 Study site and the data

### 2.1 Study site and hydrodynamic conditions

The study was carried out in the Dover Strait, a shallow-water region of the northwestern European continental shelf connecting the English Channel to the North Sea (Fig. 1a). The region is characterized by highly irregular bathymetry, with depth not exceeding 65 m and the presence of many sandbanks, roughly oriented in the dominant current direction, with depth only of a few meters at low tide.

Tidal motions of semi-diurnal periods dominate the local circulation. The tidal range in Boulogne, located on the east-

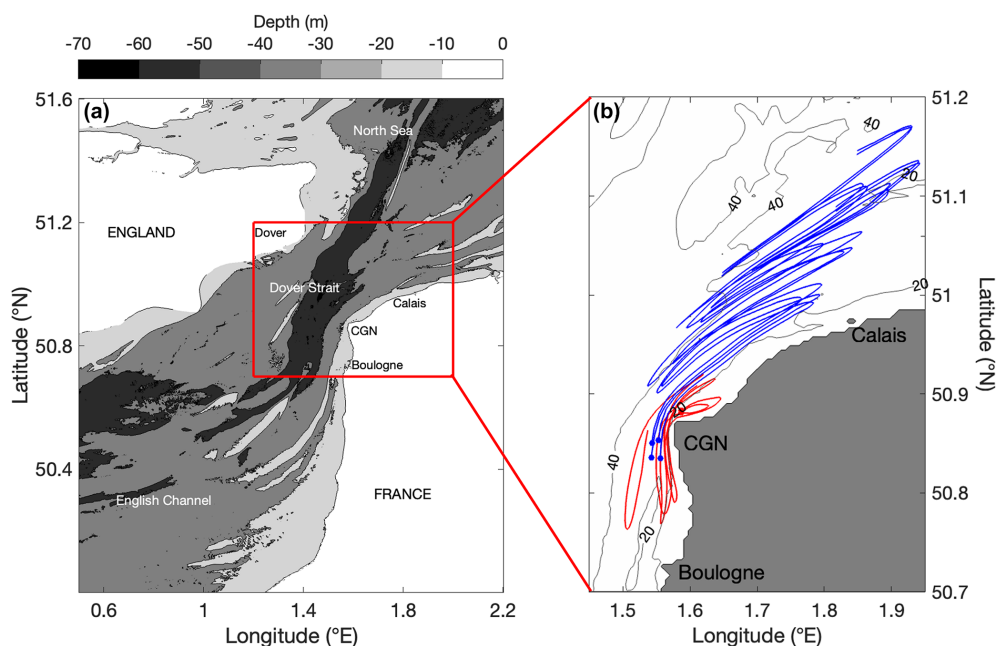
ern coast of France (Fig. 1a), is close to 9 m, and the current speed can reach  $2 \text{ m s}^{-1}$  during spring tide. The tidal stream loses around 20 % of its intensity during ebb tide, and the sea surface height and velocity are characterized by a pronounced asymmetry. A large-scale circulation in the North Atlantic Ocean generates sea level difference, driving a weak residual flow from the Atlantic Ocean towards the North Sea. This is another remarkable feature of the local hydrodynamics. The order of magnitude of the tidal residual currents in the Dover Strait ranges between 5 and  $10 \text{ cm s}^{-1}$  (Lazure and Desmare, 2012). The spatial variability of residual currents is caused by topographic features of the English Channel that constrain tidal wave propagation (e.g., Sentchev and Yaremchuk, 2007).

The wind significantly affects the local circulation. Southwestern winds can increase the average eastward flow, while northwestern and northeastern winds can reduce the tidal flow opposing the wind and even reverse it (e.g., Lazure and Desmare, 2012). The freshwater input from rivers located on the French coast (the Seine, the Somme, and the Authie rivers) has only a little influence on the water circulation in the study area.

### 2.2 Current velocity measurements

A total of six Lagrangian surface drifters were deployed in the Dover Strait during two periods of time, under relatively calm to moderate winds (mean wind speed less than  $8 \text{ m s}^{-1}$ ) and waves not exceeding 1 m height. During the first survey, referred to hereafter as S1, two surface drifters were released north of the Cap Gris-Nez (CGN) (Fig. 1b, red trajectories) for a 26 h period, from 26 November 2020 at 08:30 UTC to 27 November 2020 at 11:00 UTC. At the release, the two drifters were separated by 250 m. The survey was performed under mean tide conditions and northeastern wind of  $4 \text{ m s}^{-1}$ , on average. The two drifters of S1 will be referred to hereafter as S1-1 and S1-2. During the second survey, referred to hereafter as S2, four surface drifters were deployed west of the CGN (Fig. 1b, blue trajectories) for a 46 h period, from 10 May 2021 at 09:15 UTC to 12 May 2021 at 07:30 UTC. The drifters formed a rectangle of size of 1.3 km by 2 km. The survey was performed under spring tide conditions and stronger southwestern winds of  $6 \text{ m s}^{-1}$ , on average, with gusts up to  $12 \text{ m s}^{-1}$ . The four drifters of S2 will be referred to hereafter as S2-1, S2-2, S2-3, and S2-4.

Two types of buoys were used: coastal Nomad surface buoys manufactured by SouthTek (<https://www.southteksl.com/>, last access: 15 January 2024) and drifters manufactured at the lab. The coastal Nomad buoy is a cylinder 0.72 m long and 0.22 m wide with a cone positioned at the surface to increase buoyancy. The laboratory-made drifter comprises a cylindrical PVC hull 0.6 m long and 0.1 m in diameter weighted in its lower part. A thin square plate of  $0.3 \times 0.3 \text{ m}$  in size was installed in the upper part of the hull to assure better stability in the vertical and reduce the pitch. The drifters



**Figure 1.** (a) Modeling domain. Bathymetry shown by grey shading. Red square delimits the region where the Lagrangian measurements were performed. (b) Trajectories of Lagrangian drifters released during two field surveys on 26 November 2020 (red) and on 10 May 2021 (blue). Geographic names used in the text are also shown.

were equipped with a SmartOne GPS/satellite transceiver of Globalstar satellite network. All drifters were equipped with a 0.5 m long floating anchor positioned in the water layer at 1 m depth, allowing them to drift with surface currents. For this study, we assumed that SouthTek drifters and laboratory-made drifters behaved similarly at sea and that the discrepancies implied by their differences were negligible.

Observed surface current velocities were estimated from the drifter trajectories with a time step of 15 min, which was a nominal period of drifter positioning via GPS. Spatial coordinates (longitude, latitude) were obtained and then derived and divided by the time step to estimate the velocity at each time step.

During S1, the mean drifter velocity was  $0.8 \text{ m s}^{-1}$ . The maximum speed of  $1.6 \text{ m s}^{-1}$  was reached during peak flood flow and observed north of the CGN. The minimum speed of  $0.1 \text{ m s}^{-1}$  was reached 2 h after peak ebb flow and observed south of the CGN. During S2, the mean and maximum drifter velocities were found to be 1 and  $2.1 \text{ m s}^{-1}$  respectively (Fig. 1a).

### 2.3 Current velocity from numerical model

The water dynamics in the EEC are largely dominated by tides. The baroclinic effects on the vertical are negligible due to the enhanced mixing affecting the entire water column (e.g., Breton and Salomon, 1995). Moreover, the study area is located fairly far away from the major source of buoyancy – the Seine River, whose discharge was low during the

measurement period. The use of a 2D model was therefore justified. The variation of salinity in the horizontal plan is taken into account in the 2D model.

Lagrangian particle advection simulations performed in this study were forced by a 2D water current (barotropic) and water level forecasts modeled in the framework of the operational coastal oceanography project Modeling and Analysis for Coastal Research (Pineau-Guillou, 2013; Dumas et al., 2014). The forecasting multi-scale real-time and historical data for the French metropolitan coastlines including the eastern English Channel are freely available on their project website (<https://marc.ifremer.fr>, last access: 12 October 2023). Current, salinity, temperature, and sea level elevation fields were forecasted using the Model for Application at Regional Scale (MARS) (Lazure and Dumas, 2008).

The MARS model was developed to simulate flows across various coastal areas, ranging from regional scales to the in-shore scale of small bays or estuaries, where circulation is typically influenced by a combination of processes (Lazure and Dumas, 2008). The modeled processes involve simplifications of the incompressible Reynolds-averaged Navier–Stokes (RANS) equations based on classic Boussinesq and hydrostatic assumptions. Comprehensive information regarding model equations, the coupling of barotropic and baroclinic modes, model discretization, solving methods, computational stability according to the Courant–Friedrichs–Lewy criterion (Table 1; Lazure and Dumas, 2008), and costs is meticulously outlined in Lazure and Dumas (2008). The model accounts for kinematic free-surface and bottom

boundary conditions, contingent upon friction terms (Lazure and Dumas, 2008). The turbulence closure employed in the model follows the approach described in Gaspar et al. (1990).

To accurately simulate storm surge dynamics, the model extends sufficiently to the north and west to capture the development of depressions, which generate surges propagating in the English Channel at a greater distance in the Atlantic Ocean or in the North Sea (Idier et al., 2012). The numerical model utilizes nested configurations with progressive resolutions: (i) 2 km covering the northeastern Atlantic (level 0); (ii) 700 m at the regional scale, encompassing the English Channel (level 1); and (iii) 250 m for the eastern English Channel (level 2). This nesting technique enables the accurate capture of interactions between large-scale and small-scale processes. This technique enables the transfer of all resolved fields from lower-resolution levels to the open boundaries of higher-resolution levels. The temporal resolution of the outputs is 1 h for levels 0 and 1 and 15 min for level 2.

All relevant details concerning the choice of bathymetry, meteorological and tidal forcings, improvements in model parameterization (such as surface friction and drag coefficient), and the methodology employed for storm-surge computation are meticulously documented in Pineau-Guillou (2013). The bathymetry data for the level-2 configuration were sourced from the French Hydrographic and Oceanographic Service (SHOM). Tidal boundary conditions were derived from the global tidal model FES2004 (Lyard et al., 2006). Intertidal areas were simulated using a wetting and drying scheme. The drag coefficient utilized for wind effect parameterization is based on the variable Charnock coefficient from the WaveWatch III model (Ardhuin et al., 2011). To maintain CFL stability, the modeling time step was set to 30 s for the level-2 model.

Moreover, the numerical model outputs underwent an evaluation process (Pineau-Guillou, 2013; Dumas et al., 2014) involving available observations, confidence indicators, and descriptors of system states. Regarding the water levels, the model was validated in February 2010 at 19 tide gauges of the permanent network RONIM (French Sea Level Observation Network). For the tide, the root mean square errors are halved between level 0 and level 2: they average 22 cm for level 0, 21 cm for level 1, and 11 cm for level 2. Similar results are observed for the water levels (tide and surge) with root mean square errors averaging 26 cm for level 0, 24 cm for level 1, and 16 cm for level 2. These improvements are attributed to the enhancement of spatial resolution from level 0 to level 2 (from 2 km to 250 m), as well as the incorporation of the SHOM CST-France tidal model (with 115 harmonic components) at the boundaries of the model (Pineau-Guillou, 2013).

Hence, we posit that the comparison of model outputs for the level-2 configuration (with 250 m horizontal spatial resolution and 15 min output temporal resolution), encompassing the EEC area with available in situ data, is sufficiently ac-

curate to be utilized in our study for further drifting particle advection modeling. Originally represented on an Arakawa C-grid, surface currents were interpolated on the Arakawa A-grid (Arakawa and Lamb, 1977) for further analysis and optimization of model velocities. The model used in the analysis will be referred to hereafter as M2D.

## 2.4 Wind data

Meteorological data (wind, temperature, humidity, and atmospheric pressure) are used as the forcing of M2D. The data were provided by the Arpege (Action de Recherche Petite Echelle Grande Echelle) operational atmospheric model of Météo-France with 5 km spatial and 1 h time resolution.

The meteorological data from the model were compared to in situ measurements collected at meteorological station in Boulogne and Calais during the year of model simulations. The time- and space-averaged difference between the observed and modeled wind speed was found to be 1.7 and 0.9 m s<sup>-1</sup> for the surveying periods, giving confidence in the model wind data.

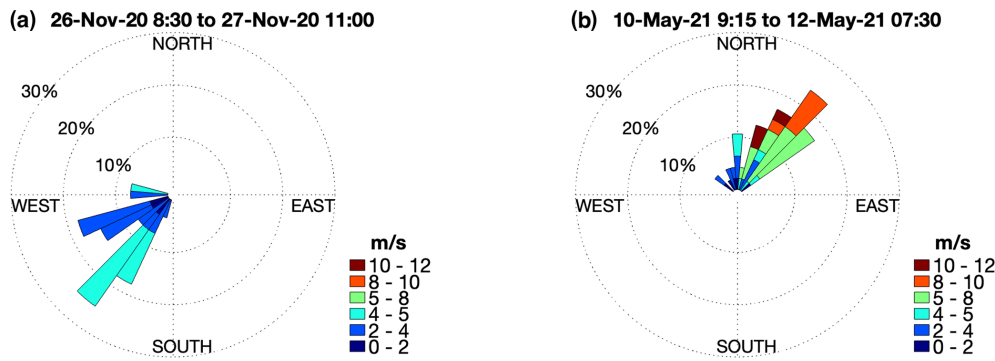
Figure 2 shows the wind rose for each survey from the Arpege model. Two dominant wind regimes were observed during the surveyed days. During S1, the wind direction was towards the southwest, and the speed did not exceed 5 m s<sup>-1</sup> with the mean value of 4 m s<sup>-1</sup>. During S2, the wind had an opposite direction, and the speed varied within the range of 4–9 m s<sup>-1</sup>, with the mean speed 6 m s<sup>-1</sup> and the maximum speed 11 m s<sup>-1</sup>.

## 3 Methodology

### 3.1 Optimal interpolation of velocity measurements

One of the methods used to constrain the numerical model outputs by observations is the optimal interpolation of observations. It provides an estimate of the state of the ocean dynamics by performing a weighted least squares fit of a background model field to observations. In general, observations are available at irregularly distributed points and are assumed to be imperfect, i.e., each observation being affected by an uncertainty (observation error). It is assumed that the observation error is not correlated with the model error.

In the OI method, a correction of a background velocity field  $\mathbf{u}_m(\mathbf{x}, t)$ , provided by a numerical model on a regular grid, is done using a linear combination of the weighted differences between the background model velocity  $\mathbf{u}_m$  and the observed velocities  $\mathbf{u}_i^*$  at point  $i$  (Bretherton et al., 1976; Sentchev and Yaremchuk, 2015; Thiébaux and Pedder, 1987). Weights chosen for minimization of the mean square difference between the observed and background velocities are a combination of model and observation covariances. With the space–time covariance matrices of the model  $\mathbf{B} = \langle \mathbf{u}_m(\mathbf{x}, t) \mathbf{u}_m(\mathbf{x}', t') \rangle$  and observations  $\mathbf{R}_{ij} = \langle \mathbf{u}_i^* \mathbf{u}_j^* \rangle$ , the op-



**Figure 2.** Wind roses for two survey periods: S1 (a) and S2 (b) from the Arpege atmospheric model (hourly data) spatially averaged over the study region.

timized velocities  $\mathbf{u}_{OI}$  are computed as follows:

$$\mathbf{u}_{OI} = \mathbf{u}_m + \sum_{ij} \mathbf{B}\mathbf{H}_j^T (\mathbf{H}_i\mathbf{B}\mathbf{H}_j^T + \mathbf{R}_{ij})^{-1} (\mathbf{H}_i\mathbf{u}_m - \mathbf{u}_i^*). \quad (1)$$

Here,  $\mathbf{H}_i$  corresponds to a linear operator projecting gridded velocity values from the apexes of the model grid cell onto the  $i$ th observation point. The covariance of  $\mathbf{u}_m$  is calculated between space–time coordinates  $(\mathbf{x}, t)$  and  $(\mathbf{x}', t')$ , where  $\mathbf{x}$  and  $t$  correspond to the velocity and time of observations, and  $\mathbf{x}'$  and  $t'$  are the corresponding velocity and time in the different ensemble members composing the covariance matrix.

The quality of the interpolation scheme is quantified by estimating the mean relative difference between the observations  $\mathbf{u}^*$  and optimized model velocities  $\mathbf{u}_{OI}$  as follows:

$$\varepsilon_{OI} = \sqrt{\frac{\sum_i (\mathbf{H}_i\mathbf{u}_{OI} - \mathbf{u}_i^*)^2}{\sum_i (\mathbf{u}_i^*)^2}}. \quad (2)$$

The relative error of the initial model,  $\varepsilon_m$ , is quantified in the same way.

An important assumption underlying the OI method is that the background field, also called the “first guess”, is a good approximation of the truth. Thus, the computation of  $\mathbf{B}$  and  $\mathbf{u}_m$  is crucial. The background velocity fields were provided by the model at the observation time step.

For estimation of model covariances, a variable number of model trajectories (26 h long for S1 and 46 h long for S2), referred to as ensemble members, were used. The sensitivity test of OI to the number of ensemble members used was performed, and the results are presented in Sect. 4.1. Three approaches were used in selecting the ensemble members. The first and easiest way is extracting them from the model on the days surrounding the survey and by respecting the conditions (wind and tidal conditions) observed during the surveying days. A total of seven ensemble members were selected using this approach.

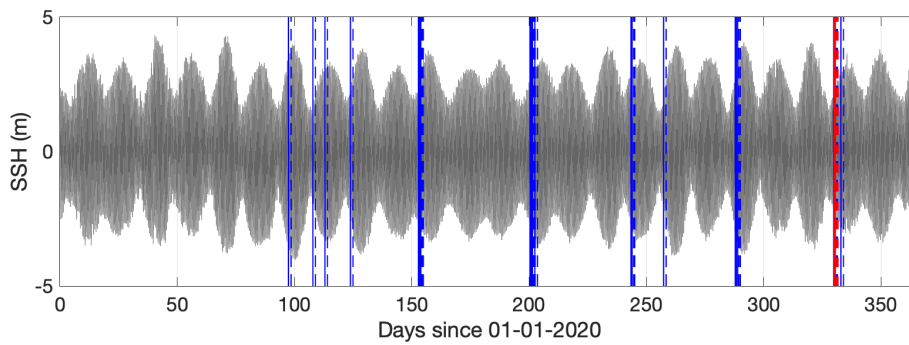
However, for an evolving ocean state, a large number of ensemble members might be required to represent the flow field evolution with statistical significance. Therefore, in the second step, ensemble members were extracted from the 1-year-long model simulation (January to December 2020) containing S1, by searching for wind and tidal conditions similar to those observed during the surveying period. By accepting a variation of the average wind speed and direction within the range  $\pm 2 \text{ m s}^{-1}$  and  $\pm 45^\circ$  respectively, for a given tidal stage, a total of 31 ensemble members were selected. In a third step, a more restrictive criterion of the range of variation of the wind, for instance,  $\pm 1 \text{ m s}^{-1}$  and  $\pm 45^\circ$ , allowed us to obtain 11 ensemble members. Figure 3 shows a chronology of ensemble members selected for OI of velocity observations during S1. Each ensemble member represents a 26 h long model run.

### 3.2 Lagrangian trajectory reconstruction

In addition to the relative error  $\varepsilon$ , the quality of the interpolation scheme can be assessed by estimating the separation distance  $d$  between the real trajectories of drifting buoys and the trajectories provided by the model. The latter were reconstructed using the OceanParcels Lagrangian framework (<https://oceanparcels.org/>, last access: 29 October 2023). Virtual particles were seeded at the time and location of the real drifters at the release. Then, they advected during a given period of time using the horizontal forward Euler method without diffusion, giving their time-dependent position  $x(t)$  and  $y(t)$ . The separation distance  $d$ , estimated at a 15 min time step and averaged over drifters, is a commonly used metric that shows how good the drifter trajectories are reconstructed by the initial model (the separation distance  $d_m$ ) or the optimized model (the separation distance  $d_{OI}$ ).

### 3.3 Correction of the wind-induced current velocities

The ocean–atmosphere coupling is difficult to reproduce correctly, especially in coastal regions. In fact, the wind generates surface Ekman currents directed  $45^\circ$  to the right of the



**Figure 3.** A total of 31 ensemble members (in blue) extracted from the model simulation of circulation in the Dover Strait using the realistic forcing in 2020. S1 period is shown in red. Sea surface height (SSH) from the model run is given in dark grey. Solid lines mark the beginning of the ensemble member, whereas dashed lines mark the end.

wind at the sea surface. It is assumed that this wind-driven velocity can attain 1 %–3 % of the wind speed at 10 m height (Jenkins, 1987; Weber, 1983).

Imperfect representation of the wind-driven velocity in the model can be improved by using velocity measurements by surface drifters. Let us assume that the flow velocity can be decomposed in two parts: the optimally interpolated velocity  $\mathbf{u}_{OI}$  and an additional correction,  $c\bar{U}_{10}$ , where  $\bar{U}_{10}$  is the wind velocity vector at 10 m height, averaged over the study area and the survey period, and  $c = \text{diag}(c_x, c_y)$  is a diagonal  $2 \times 2$  matrix whose diagonal elements are estimated by minimizing the cost function:

$$J(c) = \left[ \mathbf{X}^* - \left( \mathbf{X}_0 + \sum_k (\mathbf{u}_{OI})_k \Delta t + c\bar{U}_{10} \Delta t \right) \right]^2 \rightarrow \min_c \quad (3)$$

Here  $\mathbf{X}^*$  is a sequence of drifter coordinates at  $\Delta t = 15$  min time stepping,  $\mathbf{X}_0$  is the coordinate at the release, and summation is performed over a drifter trajectory. The expression in parentheses represents a virtual drifter trajectory after correction for the wind effect. The coefficients  $(c_x, c_y)$  were estimated for each drifter trajectory and then averaged. This implies an assumption of a stationary wind (mainly wind direction) that was supported by observations at meteorological stations during the surveying periods S1 and S2. Correcting the wind-induced velocity enables better reconstruction of the optimized velocity fields denoted hereinafter by  $\mathbf{u}_{cor}$ . The relative error of the model after performing the wind-induced velocity correction is computed using Eq. (2), after replacing  $\mathbf{u}_{OI}$  by  $\mathbf{u}_{cor}$ . The separation distance between the observed trajectories and trajectories reconstructed from the model after performing wind-induced velocity correction is referred to as  $d_{cor}$ .

### 3.4 Absolute dispersion

The absolute dispersion  $A^2$  is defined as the variance of particle spreading with respect to the mean coordinate of particles in a cluster (the barycenter). In two-dimensions, the dispersion is generally estimated along the  $x$  and  $y$  axes (Berti et al., 2011; Enrile et al., 2019). But in this study, to better account for the dominant flow direction, the variance is computed in the direction of the maximum spreading of particles and in the perpendicular direction, providing the two quantities  $A_1^2$  and  $A_2^2$ . They represent the major and minor axes of the deformation tensor and are estimated by applying the principal component analysis (PCA) to particle distribution at each time step (Emery and Thomson, 2004). As the tidal flow direction in shallow-water basins is generally constrained by local topography and coastline orientation, the ellipse orientation (major axis) gives the dominant flow direction. The ellipse orientation  $\theta$  and the variances  $A_1^2$  and  $A_2^2$  were computed as follows (Emery and Thomson, 2004):

$$\theta = \frac{1}{2} \tan^{-1} \left[ \frac{2\overline{x'y'}}{\overline{x'^2} - \overline{y'^2}} \right] \quad (4)$$

$$\begin{aligned} \left[ \begin{array}{c} A_1^2 \\ A_2^2 \end{array} \right] &= \frac{1}{2} \left\{ \left( \overline{x'^2} + \overline{y'^2} \right) \right. \\ &\quad \left. \pm \left[ \left( \overline{x'^2} - \overline{y'^2} \right)^2 + 4\left( \overline{x'y'} \right)^2 \right]^{\frac{1}{2}} \right\}. \end{aligned} \quad (5)$$

Here,  $\overline{x'^2}$  and  $\overline{y'^2}$  stand for variances of particle coordinates along the eastward  $x$  and northward  $y$  axes, respectively.

## 4 Results

### 4.1 Model velocities after optimal interpolation of the Lagrangian observations

Figure 4 shows the results of Lagrangian drifter velocity interpolation for S2. The largest number of ensemble members, 36, was used in interpolation. The discrepancy between

**Table 1.** Relative error  $\varepsilon$  (columns 2–4) and mean (time–space-averaged) separation distance (columns 5–7) between the observed and reconstructed drifter trajectories, using the initial model, optimized model, and the model after performing wind-induced velocity correction. Errors obtained with 31 ensemble members for S1 and 36 for S2 are shown in normal font, and those obtained with 7 ensemble members are given in italics.

	Relative error			Mean separation distance (in km)		
	$\varepsilon_m$	$\varepsilon_{OI}$	$\varepsilon_{cor}$	$d_m$	$d_{OI}$	$d_{cor}$
S1	0.27	0.16 (0.17)	0.16	1.5	1.4	1.3
S2	0.22	0.10 (0.11)	0.10	5.7	3.0	2.1

the initial and optimally interpolated velocities during peak flood and ebb flow (Fig. 4, color shading) varies in space with lower values found south of the CGN, for both flood and ebb flow, and the largest value ( $\sim 0.5 \text{ m s}^{-1}$ ) found in the northern part of the strait, close to the UK coast. In the French sector of the strait, the discrepancy attains  $0.2 \text{ m s}^{-1}$  over the sandbanks. However, the surface current direction is reproduced fairly well by M2D. The mean (time- and space-averaged) error of flood and ebb tide velocity is  $0.17$  and  $0.25 \text{ m s}^{-1}$ , respectively, while for the current direction, the corresponding errors are  $2$  and  $2.5^\circ$ . In general, larger discrepancies are found over sandbanks, indicating difficulties in modeling the tidal stream over complex and rapidly changing bathymetry.

The relative error of velocity,  $\varepsilon$ , appears significantly different in the initial model and after OI (Table 1). In the initial model, the error ( $\varepsilon_m$ ) is found to be fairly large:  $0.27$  for S1 and  $0.22$  for S2. Blending the model with Lagrangian observations allowed a decrease in the relative error by  $40\%$  for S1 and by more than  $50\%$  for S2 (Table 1, columns 2 and 3). The larger error obtained for S1 could be due to the location of drifter trajectories too close to the shore, during  $10 \text{ h}$  after the release. The model performance in reconstructing the drifter trajectories is probably limited in this very shallow-water region, in the vicinity of the CGN.

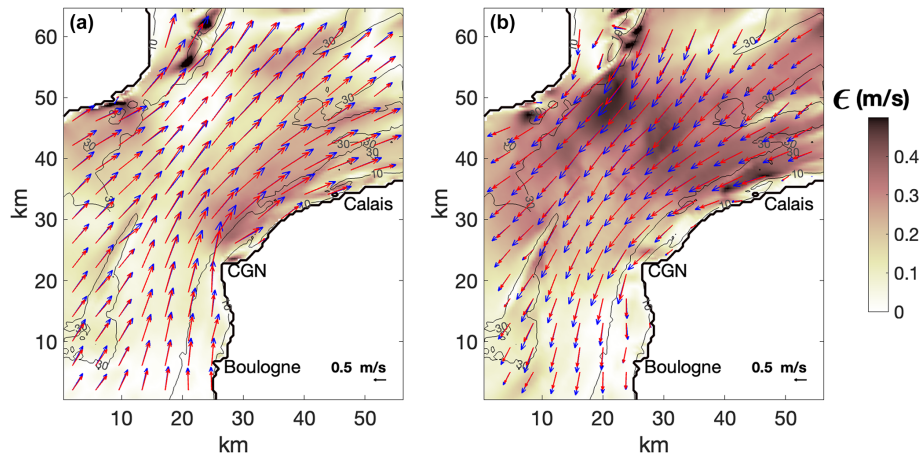
It is interesting to quantify the sensibility of OI to the number of ensemble members used. While the number is limited to seven (the smallest number identified for both surveys), the results of velocity interpolation do not change much, by less than  $10\%$  (Table 1, column 3, values in italics). These outcomes prove that in a basin dominated by the combination of tide and wind, and in the EEC in particular, the accuracy of OI is not much sensitive to the number of ensemble members used in calculating the velocity covariances. Note that wind forcing is taken into account when selecting ensemble members. With tidal range varying between  $3$  and  $9 \text{ m}$  in the EEC, the correlations of current velocities are high. This may explain why increasing the number of ensemble members results in only a slight decrease in the interpolation error.

Figure 5 shows the evolution of differences between the velocity provided by the initial and optimized model in drifter locations during S1 (Fig. 5a, b) and S2 (Fig. 5c, d). During S1, larger discrepancies between both the observed–modeled velocities and the observed–optimized velocities ( $\sim 0.3 \text{ m s}^{-1}$ ) occur at peak flood,  $13 \text{ h}$  after the deployment. However, the optimization enables the reduction of the mean discrepancy from  $0.070$  to  $0.066 \text{ m s}^{-1}$  for  $u_x$  and from  $0.10$  to  $0.066 \text{ m s}^{-1}$  for  $u_y$ . During S2, larger discrepancies are attained during both peak flood (time  $26, 40 \text{ h}$ ) and peak ebb flow (time  $8, 22, 33, 46 \text{ h}$ ). The optimization process enables the reduction of the mean discrepancy from  $0.09 \text{ m s}^{-1}$  down to  $0.07 \text{ m s}^{-1}$  for  $u_x$  and from  $0.07$  down to  $0.06 \text{ m s}^{-1}$  for  $u_y$ . These results highlight the difficulty models have to reproduce accurately the exact timing between peak flood and ebb flow.

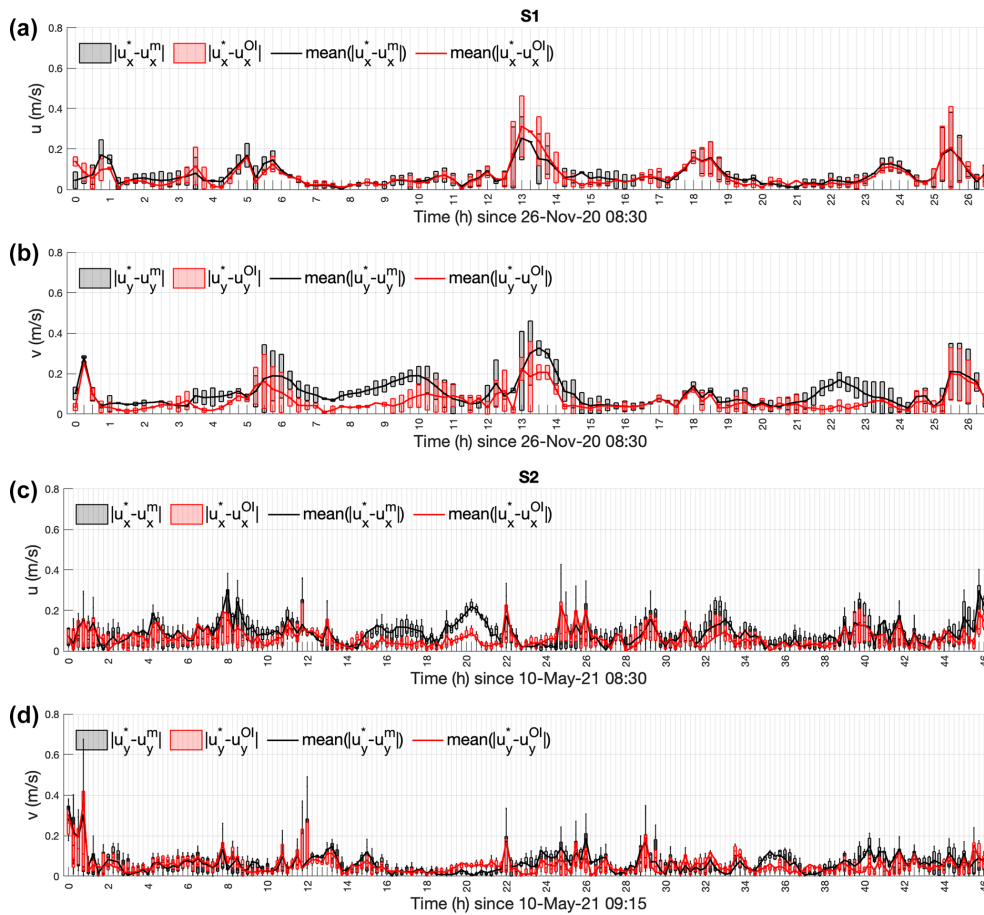
Figure 6 shows the observed drifter trajectory and that provided by the initial and optimized model during S1 and S2. The corresponding separation distance, time- and space-averaged, is given in Table 1 (columns 5–7). During S1, the model seems to underestimate the northward flow component, especially in the southern part of the domain (Fig. 6a, b). The mean initial separation distance  $d_m$  is  $1.5 \text{ km}$ . It decreases by  $0.1 \text{ km}$  after OI ( $d_{OI} = 1.4 \text{ km}$ ). During S2, the model, both initial and optimized, underestimates the northward flow component (Fig. 6c, d, e, and f). The time evolution of the trajectories is well reproduced by the model, especially for drifter of drifter S2-1, but appears shifted by  $4 \text{ km}$  compared to the observed trajectory. This gives a large mean separation distance  $d_m = 5.7 \text{ km}$ . Blending the model with observations enables the reduction of the mean separation distance by  $7\%$  for S1 and by  $48\%$  for S2 (Table 1, columns 4–5). However, the difference between the real and virtual drifter trajectories remains significant, especially for S2 (Fig. 6c, d, e, and f).

Another way to evaluate the performance of OI is to perform a “leave-one-out validation” experiment. In this experiment, one drifter trajectory is removed from the data set, and the optimization is done using the remaining drifters. This validation method provides a much less biased measure of relative error compared to that used in cross-validation, because the model is repeatedly fit to a data set that contains  $n - 1$  drifter trajectories. At the end of the experience, the relative error was reduced by  $22\%$  for S1 and by  $36\%$  for S2. This proves that the proposed OI technique is capable of efficiently correcting the few drifter trajectories of the model velocity field. These values of error reduction appear similar to those given in Table 1 and demonstrate the efficiency of the OI. This means that with only one or few drifters, it is possible to improve the model velocities in an optimal way.

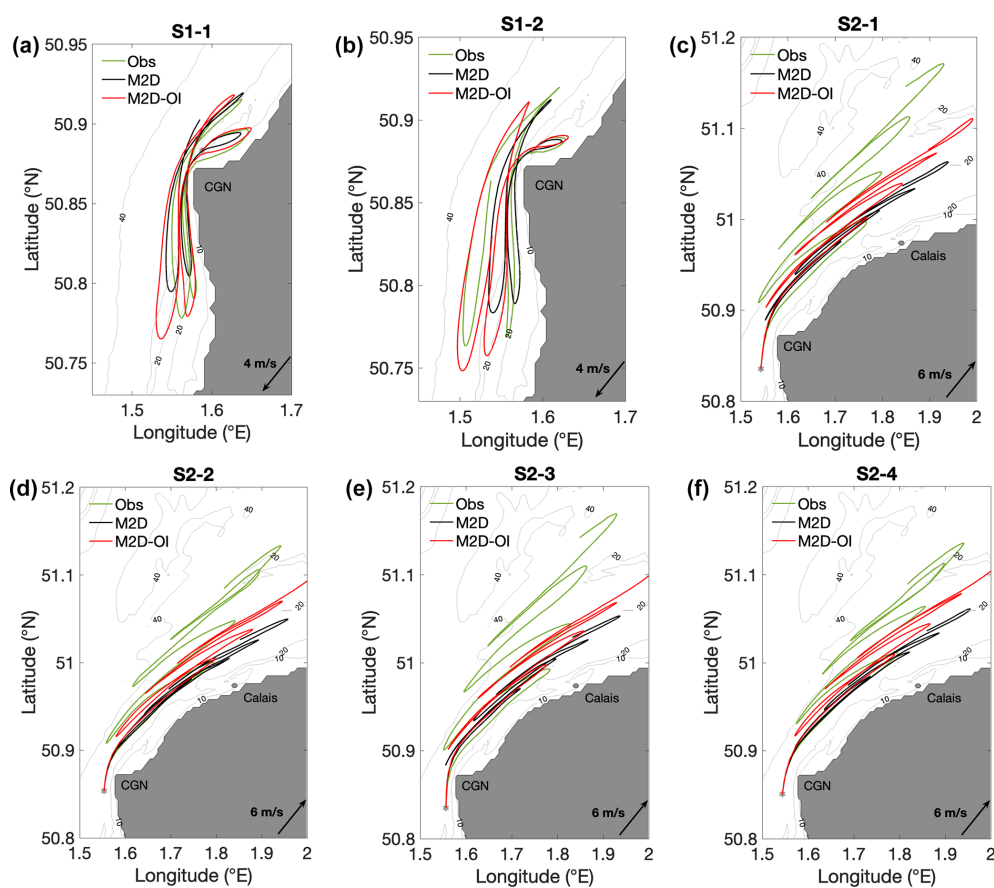




**Figure 4.** Current velocities during peak flood flow (a) and peak ebb flow (b) of survey S2. Red and blue vectors show the initial and optimized model velocities, respectively. The absolute difference between the initial and optimized model velocity ( $|u_{OI} - u_m|$ ) is shown by color shading.



**Figure 5.** Boxplots of absolute zonal (a, c) and meridional (b, d) velocity difference between observations and initial model (black) and observations and optimized model (red) at S1 (a, b) and S2 (c, d) drifter locations. For S1 (a, b), where only two drifters are considered, the top and bottom edges of each box are the maximum and minimum values, and the line within the box indicates the median, equal to the mean. For S2 (c, d), the top and bottom edges of each box are the upper (0.75) and lower (0.25) quartiles, respectively. The line within the box indicates the median. The whiskers above and below each box are the nonoutlier maximum and minimum.



**Figure 6.** Observed trajectory of drifter S1 (a, b) and S2 (c, d, e, f) (green lines). Corresponding trajectories provided by the initial model M2D are shown in black and those resulting from OI in red. Mean wind speed and direction are denoted by a black arrow.

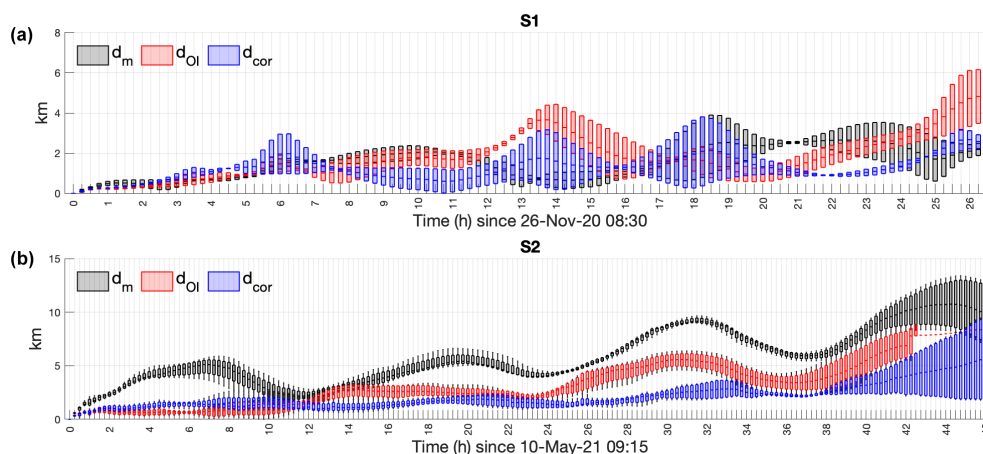
#### 4.2 Wind-induced velocity correction

The fact that larger separation distance between the observed and reconstructed trajectories was obtained during S2 (under strong wind conditions) indicates that the effect of wind on surface currents is poorly reproduced in the model. To further reduce the discrepancy between the observed and modeled trajectory, the least squares method is used to estimate a correction to wind-induced velocity. Figure 7 shows the evolution of the separation distance between the observed and reconstructed drifter trajectories using the initial, optimized, and velocity field after correcting the wind-induced current. The correction term  $c\bar{U}_{10}$  (Eq. 3) was calculated for both zonal and meridional wind components and given as percentage of the wind speed, corresponding to  $-0.4\% \bar{U}_{10}$  and  $2.8\% \bar{V}_{10}$  during S1 and  $-0.4\% \bar{U}_{10}$  and  $1.7\% \bar{V}_{10}$  during S2.

Figure 7a demonstrates that, under certain conditions, the wind-induced velocity correction is not effective. For example, at hour 7 and 14 (Fig. 7a), the separation distance attains its largest value ( $d_{\text{cor}} \sim 3$  km). This could be due to the location of the buoy too close to the shore and to the CGN cliffs (50 m high) where the sea surface and the buoys are less ex-

posed to the effect of northwestern winds. However, the correction of the wind-induced velocity enables much better trajectory reconstruction with an averaged separation distance of  $d_{\text{cor}} = 1.3$  km (Fig. 8a, b; Table 1, column 7). During S2, the separation distance  $d_{\text{cor}}$  is slightly larger ( $\sim 2$  km) than  $d_{\text{OI}}$  during the first 11 h of drift (Fig. 7b), when the wind speed decreased from  $10.5$  to  $2 \text{ m s}^{-1}$ . On the contrary, during the second part of the survey, when the wind increased again to  $\sim 8 \text{ m s}^{-1}$ , the correction provides much better results, with  $d_{\text{cor}}$ , averaged over all the drifters of S2, always inferior to  $d_{\text{m}}$  and  $d_{\text{OI}}$  (Fig. 7b).

On the whole, it is remarkable that the wind-induced velocity correction enables much better trajectory reconstruction during both S1 and S2. It allows a total reduction of  $d$  by 13 % during S1 and by 63 % during S2 (Table 1, columns 5–7). During S1, the optimization and correction processes produce realistic trajectories for both dinghies. During S1, the optimization and wind processes produce realistic trajectories, particularly for dinghies S2-1 and S2-3 (Fig. 8c, e). On the other hand, the corrected trajectories of S2-2 and S2-4 still lack precision compared to observed trajectories (Fig. 8d, f). At the time of deployment, S2-1 and S2-3 were to the south of the other pair, S2-2 and S2-4. It seems that



**Figure 7.** Boxplots of separation distance between the observed and reconstructed trajectories for drifters S1 (a) and drifters S2 (b) using the initial and optimized model and the model after performing the wind-induced velocity correction. For S1 (a), where only two drifters are considered, the top and bottom edges of each box are the maximum and minimum values, and the line within the box indicates the median, equal to the mean. For S2 (b), the top and bottom edges of each box are the upper (0.75) and lower (0.25) quartiles, respectively. The line within the box indicates the median. The whiskers above and below each box are the nonoutlier maximum and minimum. During S2 (b), three of the four modeled trajectories escaped the domain around 42 h, resulting in a simple curve without boxplot.

the impact of the wind is different between these two pairs of buoys, even though they are only 1 km apart. This means that wind correction is less effective for drifters S2-2 and S2-4.

Compared to the separation distance  $d$ , the relative error  $\varepsilon$  (Table 1, columns 2–4) appears equal for both the optimized model and model after performing wind-induced velocity correction. Because the relative error accumulates over time, a small error does not imply the best trajectory reconstruction, either in space or time. This underlines the usefulness of separation distance for evaluating the model velocity field in terms of Lagrangian tracking.

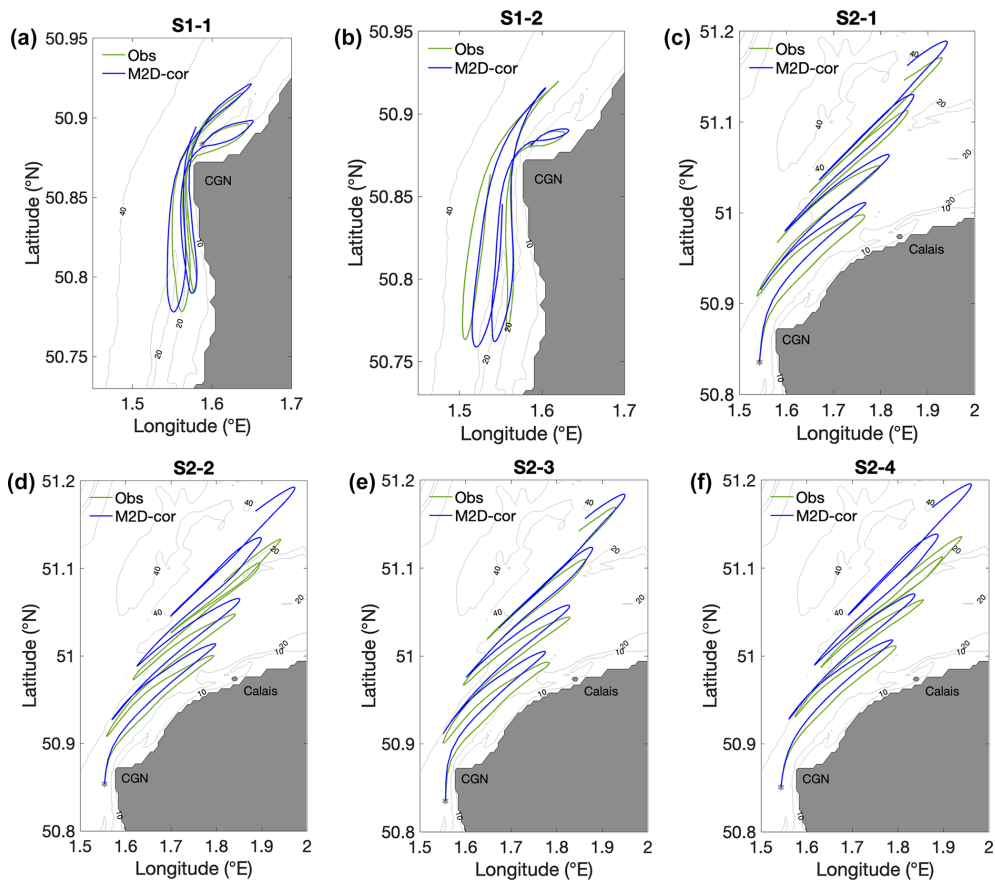
### 4.3 Absolute dispersion

After applying OI of velocity observations and correcting the wind-induced velocity, the resulting surface current fields are used with more confidence to assess dispersion processes, in particular by estimating the absolute dispersion. A total of 225 particles separated by 250 m were seeded within a rectangular shape area north of CGN. The center of mass of this cluster of particles, referred to hereafter as cluster-N, was located 1.7 km offshore. The second cluster, referred to as cluster-W, was located west of CGN with its center of mass separated from the shore by 2.1 km (Fig. 9a). Each cluster formed a rectangle of size 3.3 km by 3.5 km. The particles were advected during the 26 and 46 h time periods using the OceanParcels software. Three velocity fields were utilized: those provided by the initial model, the optimized model, and the model after applying the wind-induced velocity correction.

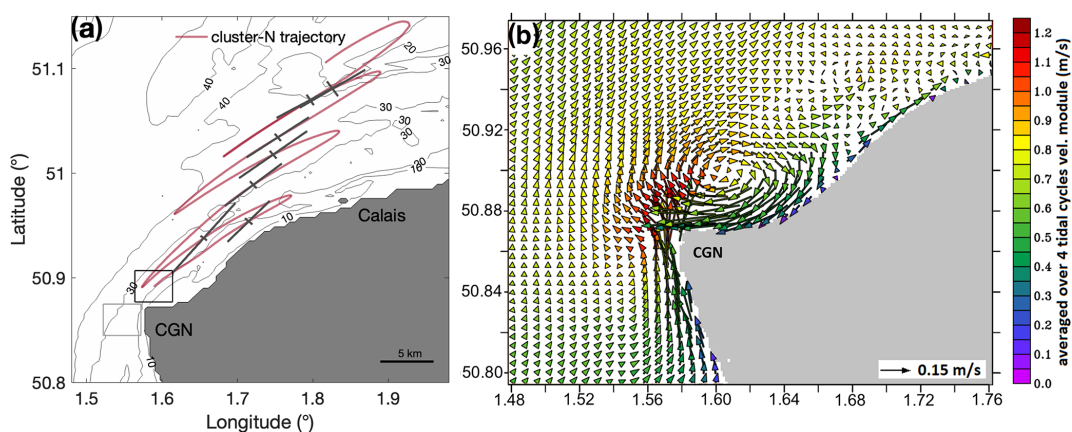
Absolute dispersion is used to quantify the rate of spreading. PCA allows the characterization of the dominant direction of spreading and the shape of a cluster of passively drift-

ing particles at different time intervals. Figure 9a shows the time evolution of spreading along the ellipse axes ( $A_1$  and  $A_2$ ) during S2 at a 6 h time step roughly corresponding to the time of high and low water in Boulogne. The spreading appears significantly larger in the alongshore direction. Similar results are obtained for particles in cluster-W (not shown). The effect of tidal currents on particles spreading consists in elongation of the cloud of particles in the dominant current direction.

The time evolution of spreading during both surveys is shown in Fig. 10. Similar results are obtained during S1 and S2 with spreading that is estimated to be 4 times larger for cluster-N than for cluster-W ( $A_1 = 5.8$  km for cluster-N and  $A_1 = 1.4$  km for cluster-W during S1). Particles in cluster-N experienced very large spreading shortly after the release under stronger wind conditions observed during S2:  $A_1 = 10$  km at time  $t = 8$  h (Fig. 10b). The spreading is found to be 30 % weaker ( $A_1 = 7$  km) under northeastern wind conditions (S1) with lower wind speed (Fig. 10a). The enhanced spreading of particles in cluster-N is due to large velocity shear induced by an anticyclonic tidal eddy generated during flooding tide (Fig. 9b). Particles in this cluster are effectively driven by the eddy, whose higher nearshore velocities induce stirring of particles westward along the shore. When the tidal eddy weakens and disappears,  $A_1$  slightly decreases, causing particle alignment in the main direction of the flow. At each peak flood tide, the stronger and heterogeneous tidal flow coming from the Dover Strait towards the North Sea (mean velocity of  $0.9 \text{ m s}^{-1}$  and spatial range of variation of  $1.6 \text{ m s}^{-1}$ ) causes shear dispersion and increases the spreading rate. However, at each peak ebb flow, the spreading along the major axis decreases. Particles seem



**Figure 8.** Observed drifter trajectory (green lines) and the trajectory reconstructed after applying the wind-induced velocity correction (blue) for S1 (a, b) and S2 (c, d, e, f).



**Figure 9.** (a) Evolution of the spreading in cluster-N (6 h spacing) during S2. The red trajectory represents the trajectory of the center of mass of cluster-N particles. Semi-axes of the ellipse are represented by black lines. The length of semi-axis of the ellipse approximating the particle dispersion accounts for the particle spreading ( $A_1$ ,  $A_2$ ). Results were obtained with optimized surface currents after performing wind-induced velocity correction. The area of particle release is shown by a black rectangle for cluster-N and a grey rectangle for cluster-W. (b) Residual velocity around the CGN obtained by averaging the model velocities over four tidal cycles.

to become more concentrated when impacted by the weaker and homogenous tidal flow (mean velocity of  $0.4 \text{ m s}^{-1}$  and spatial range of variation of  $1.2 \text{ m s}^{-1}$ ). In contrast, particles in cluster-W are impacted only by the tip of the tidally induced eddy, pushing them toward the northeast, away from the area of high nearshore velocity. Thus, particles in cluster-W experience a relatively weak spreading during both survey periods ( $A_1 < 2.4 \text{ km}$ ).

In comparison to the model after interpolating the velocity measurements and performing the wind-induced velocity correction, the initial model tends to underestimate the spreading along the major axis  $A_1$  by 20 %, whereas it tends to overestimate the spreading along the minor axis  $A_2$  by 13 % for both clusters and both surveys. It should be mentioned that the optimized model and the model with wind-induced velocity correction provide nearly identical results (a 1 % difference is found).

## 5 Discussion

The present study demonstrates how the Lagrangian observations can be used to improve the performance of the high-resolution model MARS. It is shown that the optimal interpolation of drifter data affects not only the model velocity fields but also the dispersion properties. Optimizing the model outputs and correcting for wind-induced velocity reduces the model–data misfit for velocity by 50 % and results in a significant (10 %–20 %) change of the dispersion rate caused by the correction of velocities.

Objective mapping methods, including OI, have been widely used in oceanographic studies. Sentchev and Yaremchuk (2015) and Thiébaud et al. (2019) applied the OI to constrain a high-resolution simulation of coastal currents by the MARS-2D model using towed acoustic Doppler current profiler (ADCP) measurements in the English Channel. They obtained a significant decrease of the model error (50 %), as the result of the velocity correction.

Kim et al. (2008) optimally interpolated the surface current velocities derived from high-frequency radar measurements along the west coast of the United States by using a predefined (exponential shape) isotropic spatial covariance function instead of a covariance matrix derived from ensemble model simulations. The method allowed one to obtain a continuous set of current vector maps by taking into account the measurement accuracy. A similar approach has been used for surface current mapping from satellite altimeter data at the global scale (e.g., Ma and Han, 2019; Wilkin et al., 2002).

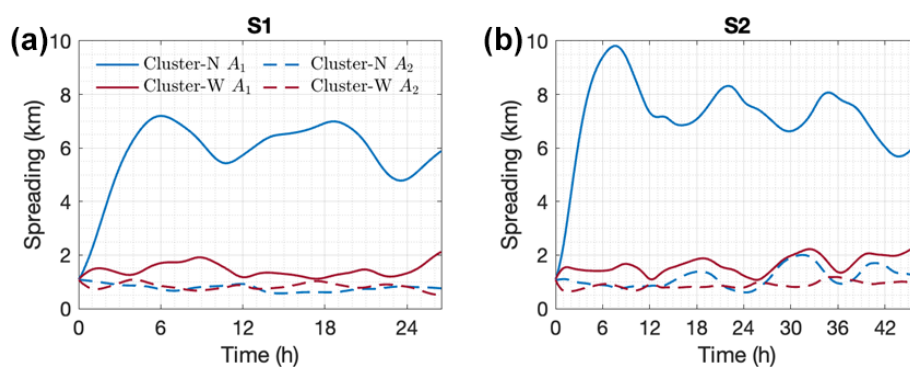
The efficiency of optimal interpolation of drifter observations has been assessed in detail by Molcard et al. (2003). Using a quasi-geostrophic model within an idealized domain, an interpolation scheme based on general Bayesian theory, and twin data experiments with virtual drifters, the authors quantified the performance of data assimilation. For an optimal choice of parameters (number of drifters, sampling pe-

riod, and uncertainties of observations and model outputs), the relative error between the observed and modeled quantities decreased from 58 % to 11 %. The final model–data discrepancy obtained in our study appears to be similar (Table 1, column 3). This increases confidence in the results of the proposed optimization technique.

In this study, we utilized fused data sources to assess fusion outcomes. Such an evaluation process may not objectively reflect the effectiveness of the fusion method and the characteristics of the real ocean current field. However, the “leave-one-out validation”, which uses one drifter trajectory as a control data set (repeatedly replaced) and other trajectories for optimization, provides a much less biased measure of error and enables trusting the fusion outcomes. An alternative to this technique, which requires more drifter trajectories, is the “cross-validation” method where high-resolution observational data could be partitioned into training and validation sets (Le Rest et al., 2013, 2014).

To further explore the performance of OI in application to drifter data in the tide-dominated basin, sensitivity of the model correction to the number of ensemble members was assessed. The results showed that in the EEC, the performance of OI was not significantly affected by the number of ensemble members. Increasing this number from 7 to 31 provided only a 10 % reduction in relative error. However, in regions with low tidal forcing (e.g., Mediterranean Sea) or with significant swell and freshwater inputs, selecting ensemble members could be more challenging. In such cases, alternative clustering methods like  $K$ -means clustering or SOMs (self-organizing maps) could be considered (Hernández-Carrasco et al., 2018; Nguyen-Duy et al., 2021; Solabarrieta et al., 2015).

A method of correction for the wind effect, often poorly represented in numerical models, especially during the periods of strong winds, appears simple, physically robust, and efficient. A comparison of the modeled and observed drifter trajectories revealed that wind-induced velocities are largely overestimated in M2D. As a result, a significant shoreward displacement of the modeled trajectory under strong southwestern winds was obtained (Fig. 6b). The mean separation distance between the observed and modeled trajectories attained 5 km (Table 1 column 5) and the maximum separation 13 km for drifter S2-3. In order to achieve better agreement, the wind-induced current velocity correction was done (Eq. 3) under the assumption of a stationary wind over the observation period. The wind time series from ARPEGE atmospheric model and observations at meteorological stations supported this choice. In principle, the method of correction can be easily adopted for situations with evolving wind. However, in other situations, for example, when the drifters were observed close to the shore (Fig. 6a), the correction method may be less efficient. In our case, the proposed correction method allowed the reduction of the separation distance between the observed and modeled trajectories by 63 %



**Figure 10.** Time evolution of spreading  $A_1$  (solid line) and  $A_2$  (dashed line) under environmental conditions observed during S1 (a) and S2 (b). Spreading of cluster-N is shown in blue and of cluster-W in dark red. Results are obtained from the model after performing a wind-induced velocity correction.

for S2, under strong winds, and by 13 % for S1, under weak winds (Table 1, column 7).

This highlights the importance of an accurate representation of the wind effect in high-resolution coastal circulation models. For example, the effect of the Stokes drift on passive tracers, drifting in the surface layer, should be accounted for. In fact, MARS is an Eulerian hydrodynamic model, not coupled with a wave model in the considered configuration. For this reason, wave–current interactions are neglected in the model. Moreover, the wave-induced current velocity (Stokes drift velocity), estimated as 1 % of the wind speed (Ardhuin et al., 2012, 2018), can considerably modify the transport pathways of passively advected particles. Dobler et al. (2019), van den Bremer and Breivik (2018), and Curcic et al. (2016) also highlighted the impact of the Stokes drift on the behavior of passive tracers, micro-plastics, and oil spills, especially under strong winds.

One of the practical applications of oceanographic studies is the assessment of turbulent dispersion of materials in the marine coastal environment. It attracts a growing interest because our seas and oceans are being degraded by human activities that harm marine life, undermine coastal communities, and inject harmful substances into the ocean (Landrigan et al., 2020). Marine turbulence is considered the main factor controlling the spreading of materials in seawater (van Sebille et al., 2020). The present study aims to evaluate the turbulent dispersion and demonstrate how the dispersion estimates can be improved in one of the busiest maritime straits. Optimal interpolation of drifter data was used to optimize the sea current velocities. It was found that the resulting change in the velocity field may lead to adjustment of the velocity gradients, which, in turn, increase the rate of dispersion. Consequently, the absolute dispersion in the model was found to be significantly larger after interpolation of the drifter data, which is not surprising given the results reported in other studies. Modeled velocities are generally lower and less variable than observed velocities (Kjellsson and Döös, 2012), especially under strong wind conditions (Curcic et al., 2016).

In addition, other studies highlighted that in tide-dominated regions, with large spatial variation of velocity, the coastal flow is characterized by strong shear dispersion (Van Dam et al., 1999; Zimmerman, 1986). In particular, an enhancement of the dispersion rate was found in the vicinity of headlands or under a significant bathymetric change (Geyer and Signell, 1992). Numerical studies in the English Channel have shown that passive particles released offshore experience lower dispersion compared to the particles released close to the shore where the bathymetry variation is large. Sentchev and Korotenko (2005) documented that under the joint effect of freshwater input and tides, a cluster of particles released in the nearshore coastal flow experienced large stretching along the shore. These results are in good agreement with the behavior of particles in cluster-N, affected by the near-shore coastal flow and tide-generated transient eddy.

## 6 Conclusions

In this study, we tested a computationally efficient method of combining numerical modeling with surface drifter observations to obtain a more reliable estimate of turbulent dispersion in the narrowest and most energetic part of the EEC: the Dover Strait.

Using optimal interpolation to combine the high-resolution MARS model outputs with two and four drifter trajectories allowed reconstruction of the surface velocity evolution with a 50 % reduction in the error between observed and modeled velocities. Additional correction of the wind-induced velocity component enabled the further reduction of the separation distance between observed and modeled trajectories (63 % reduction of separation distance under strong winds). Particle spreading, estimated from more realistic velocities, obtained after the OI and wind-induced current corrections, was found to be 20 % higher north of the CGN and 13 % lower south of the CGN, compared to the

initial model run. Spatial variability in dispersion was identified. It is assumed to be related to small-scale features of the local circulation generated by tidal flow interaction with the headland (CGN) and irregular topography.

The proposed methodology can be used to improve existing decision-making support tools or to design new tools for monitoring the transport and dispersion of materials in coastal ocean environments.

*Code availability.* The optimal interpolation code and the scripts to reproduce the figures of the article are available upon request to the corresponding author.

*Data availability.* MARC simulations from the MARS numerical ocean model are available on the project website at <https://marc.ifremer.fr> (MARC project, 2024).

Meteorological data from Météo-France (observations and model) are now available at <https://donneespubliques.meteofrance.fr/> (Meteo France, 2024).

*Author contributions.* AS designed the drifter experiments and carried them out. SB performed the data analysis and prepared the manuscript with contributions from AS and EA.

*Competing interests.* The contact author has declared that none of the authors has any competing interests.

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## Chapitre 4

# Structures de convergence des courants dans le golfe de Gascogne

### 4.1 Introduction

#### *Motivations de l'article*

Le golfe de Gascogne a été décrit dans de nombreuses études comme une zone d'accumulation de déchets marins flottants. Généralement, ces derniers s'accumulent dans des structures linéaires d'environ un kilomètre. Ces structures de convergence des courants côtiers sont souvent associées à des processus tels que les mouvements verticaux aux limites des panaches fluviaux, les fronts estuariens, les cellules de Langmuir ou les ondes internes. Comprendre la dynamique et la variabilité des structures de convergence des courants côtiers et leur rôle dans le transport des déchets marins est essentiel pour atténuer la pollution.

De nombreuses recherches ont été consacrées à l'étude du transport et de la distribution spatiale de matières particulaires telles que les polluants, les marqueurs isotopiques radioactifs, l'ichtyoplancton, les algues ou les déchets marins. La plupart de ces travaux se concentrent sur la présence de structures de convergence à grande échelle tels que les gyres subtropicaux. À mésoéchelle, les études se sont concentrées sur la convergence de matière particulaire dans des tourbillons dont la durée de vie varie de quelques semaines à quelques mois. Ces tourbillons ont la capacité de retenir la matière particulaire et leur dérive peut conduire au transport de matière particulaire sur des milliers de kilomètres. À une échelle inférieure, la sous-mésoéchelle, l'étude du transport et de la distribution spatiale de matière particulaire fait l'objet d'une attention croissante. Les structures de sous-mésoéchelle jouent un rôle crucial dans le transport et la dispersion de matière particulaire et se manifestent sous la forme de fronts, de filaments et de tourbillons qui constituent une voie rapide et efficace de transport.

Cependant, à ces échelles, l'étude des processus de transport est un défi en raison de la petite taille et de la nature éphémère des structures associées. Les modèles numériques de circulation et les techniques d'observations ont des difficultés à représenter avec précision les structures de convergence des courants à fine échelle par manque de résolution spatio-temporelle principalement.

Dans ce contexte, la présence de deux stations radar HF du réseau EuskOOS et le lâcher de bouées dérivantes dans la région d'étude offre une réelle opportunité pour l'analyse de la dynamique des eaux côtières et son évolution spatio-temporelle dans l'objectif d'identifier les

structures de convergence des courants et leur rôle dans le transport et la dispersion des déchets marins.

### *Méthodes d'analyse*

L'objectif de cette étude est d'améliorer l'efficacité de l'identification des structures de convergence des courants côtiers dans le sud-est du golfe de Gascogne et d'étudier leurs caractéristiques, par la fusion des données des mesures de vitesses de courants de surface radar HF et des bouées lagrangiennes, et d'évaluer les résultats obtenus par des données indépendantes (images satellites de Chl-a).

Le sud-est du golfe de Gascogne est pourvu de sorties de modèle opérationnel tridimensionnel et de données radar HF. Cependant, les modèles numériques de circulation, tout comme les observations radar, sont souvent limités par leur résolution spatio-temporelle, ce qui complique parfois la mesure précise des flux turbulents et des processus de dispersion. Ainsi, la fiabilité de ces différents champs de courant de surface pour représenter les processus de transport à méso et sous-mésoéchelle a été comparée. D'une part, les différents champs de courant de surface ont été comparés avec des données ADCP acquises durant des campagnes de mesures. D'autre part, l'erreur quadratique moyenne et l'erreur lagrangienne ont été calculées. Cette dernière est estimée en sommant et en moyennant les distances de séparation entre les trajectoires des bouées réelles et celles des particules virtuelles. Ces particules virtuelles, lâchées à l'emplacement des bouées réelles, sont advectées dans les différents champs de courants à chaque pas de temps à l'aide du logiciel de transport lagrangien OpenDrift.

Ensuite, le champ de courant de surface démontrant les erreurs quadratiques moyennes et lagrangiennes les plus faibles, le champ radar HF, a été optimisé en utilisant la technique d'IO. Les vitesses des courants de surface fournies par des bouées réelles sont utilisées pour contraindre les vitesses mesurées par radar HF et diminuer davantage l'erreur entre les vitesses mesurées par les bouées et observées par radar HF.

A partir de ces courants de surface optimisés et en se plaçant dans le référentiel lagrangien, la dispersion a été estimée. Cette métrique est généralement utilisée pour quantifier l'intensité de l'étalement des particules dans un cluster induit par le champ de courant de surface et permet d'identifier les régimes fondamentaux de dispersion dans le flux turbulent.

Dans un second temps et toujours dans le référentiel lagrangien, les FSLE ont été estimés. Les FSLE sont généralement utilisées pour analyser le transport et le mélange océanique. Les valeurs des FSLE les plus élevées se concentrent le long des lignes caractéristiques, identifiant les régions d'étirement maximal, qui correspondent aux structures océaniques telles que les fronts ou les limites des tourbillons. Comme ces lignes caractéristiques (lignes de crête) ne peuvent pas être traversées par les particules, elles agissent comme des barrières au transport, régissant ainsi le mouvement des particules de fluide environnantes. Ainsi, les lignes de crête des FSLE sont utilisés comme proxy des structures de convergence des courants de surface, susceptible d'accumuler de la matière particulaire et tout particulièrement les déchets marins flottants. La position des lignes de crête des FSLE a aussi été comparée à la distribution spatiale des hautes concentrations de Chl-a issues de la télédétection afin de valider les fronts océaniques identifiés. Ainsi, cette dernière étape nous permet de vérifier nos hypothèses en validant la pertinence de notre approche et de nos résultats.

### *Résultats*

Les résultats de cette étude sont multiples. D'une part, la fusion des données radar HF et des données de bouées lagrangiennes a permis de réduire l'erreur lagrangienne moyenne et l'erreur quadratique moyenne de 50% et 30% respectivement, et d'améliorer de manière significative la reconstruction de la vitesse des courants de surface.

La distribution spatiale des FSLE, obtenues à partir de l'analyse lagrangienne des vitesses optimisées, a été comparée aux champs de Chl-a obtenus par télédétection. Il a été démontré que les lignes de crête des FSLE contrôlent la distribution spatiale des eaux riches en Chl-a provenant des rivières, identifiées via satellite. Ces lignes de crête représentent fondamentalement la position des structures de convergence des courants côtier pouvant potentiellement affecter l'agrégation des déchets marins flottants.

L'analyse de la dispersion absolue a révélé une forte intensité de la turbulence en direction parallèle à la côte (cent fois supérieur par rapport à la direction côte-large), ce qui est également confirmé par la distribution spatiale des lignes de crête des FSLE. L'alignement entre les lignes de crête des FSLE et la distribution spatiale de forte concentration de Chl-a a été observé durant les campagnes de mesures, déterminant souvent les limites de l'expansion du panache fluvial de l'Adour dans la zone d'étude.

Ces résultats représentent une avancée significative dans l'identification des structures de convergence des courants à partir de la vitesse des courants de surface télédétektés et ont des applications directes pour l'étude du transport et de l'agrégation des déchets marins flottants. Cette connaissance est essentielle en océanographie physique car les déchets marins servent de traceurs pour comprendre la circulation océanique à sous-mésoéchelle. Inversement, notre compréhension des courants océaniques joue un rôle important dans la détermination des trajectoires des déchets marins flottants. Cela facilite la recherche de solutions pour lutter contre la pollution, ce qui est particulièrement important pour la gestion durable de l'environnement marin.

Différents aspects de l'étude, brièvement présentés ci-dessus, ont été développés plus en détail dans l'article de recherche : « *Coastal current convergence structures in the Bay of Biscay from optimized high-frequency radar and satellite data* » publié dans la revue *Science of the Total Environment*.

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En outre, un article de vulgarisation est paru dans la revue en ligne *The Conversation*, intitulé « *Une nouvelle méthode pour localiser les déchets en mer* »

<https://theconversation.com/une-nouvelle-methode-pour-localiser-les-dechets-en-mer-223896>.

**4.2 Article : *Coastal current convergence structures in the Bay of Biscay from optimized high-frequency radar and satellite data***



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# Coastal current convergence structures in the Bay of Biscay from optimized high-frequency radar and satellite data

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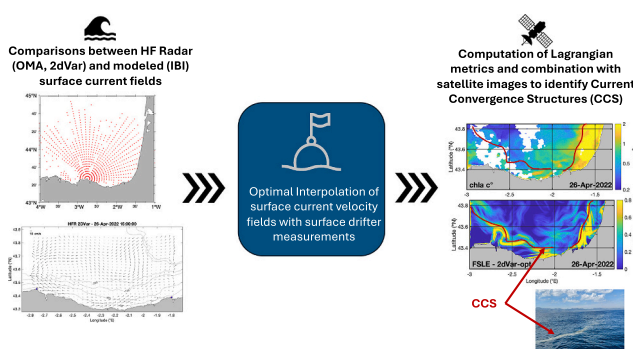
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## HIGHLIGHTS

- Southeastern BoB exhibits small-scale structures efficiently aggregating particles.
- OI merges drifter and HFR data, aiding CCS research and identification.
- FSLE are valuable to study CCS, but highly reliant on underlying Eulerian fields.
- Backward-in-time FSLE ridgelines delimit the spatial distribution of Chl-a.

## GRAPHICAL ABSTRACT



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## ABSTRACT

The southeastern Bay of Biscay has been described as a “dead end” for floating marine litter, often accumulating along small-scale linear streaks. Coastal Current Convergence Structures (CCS), often associated with vertical motions at river plume edges, estuarine fronts, or other physical processes, can be at the origin of the accumulation. Understanding the formation of CCS and their role in the transport of marine litter is essential to better quantify and to help mitigate marine litter pollution. The Lagrangian framework, used to estimate the absolute dispersion, and the finite-size Lyapunov exponents (FSLE), have proved very effective for identifying CCS in the current velocity field. However, the quality of CCS identification depends strongly on the Eulerian fields. Two surface current velocity data sets were used in the analysis: the remotely sensed velocities from the EuskOOS High-Frequency Radar (HFR) network and velocities from three-dimensional model outputs. They were complemented by drifting buoy velocity measurements. An optimization method, involving the fusion of drifting buoys and HFR velocities is proposed to better reconstruct the fine-scale structure of the current velocity field. Merging these two sources of velocity data reduced the mean Lagrangian error and the Root Mean Square Error (RMSE) by 50 % and 30 % respectively, significantly improving velocity reconstruction. FSLE ridgelines obtained from the Lagrangian analysis of optimized velocities were compared with remotely sensed concentrations of Chlorophyll-a. It was shown that ridgelines control the spatial distribution of phytoplankton. They fundamentally

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represent the CCS which can potentially affect marine litter aggregation. Analysis of the absolute dispersion revealed large stirring in the alongshore direction which was also confirmed by spatial distribution of FSLE ridgelines. The alignment between FSLE ridgelines and patterns of high Chlorophyll-a concentration was observed, often determining the limits of river plume expansion in the study area.

## 1. Introduction

Extensive research has been devoted to the study of the transport and spatial distribution of particulate material such as pollutants, radioactive isotope markers, algae or marine litter, at large scales (Budyansky et al., 2015; Chenillat et al., 2021). Most of these works focus on the presence of macro-scale CCS, such as subtropical gyres that exhibit remarkable persistence over time (Cózar et al., 2021, 2014; van Sebille et al., 2020, 2015) or the transport of particulate material and connectivity patterns at basin scales, in regions like the northern Iberian waters and the Mediterranean sea (Pereiro et al., 2019; Zambianchi et al., 2017). Some basins, dominated by large-scale open ocean processes and particularly by Ekman dynamics, have been identified as convergence regions for floating marine litter (van Sebille et al., 2020). Indeed, it was estimated that approximately half of the global floating marine litter is concentrated within these hotspots (Kershaw, 2016), emerging the concept of the 7th continent of plastic. However, global distribution and surface-transport mechanisms of floating marine litter are still a long-standing puzzle largely determined by the object's characteristics (like buoyancy and sizes of litter items) (Morales-Caselles et al., 2021).

At mesoscale (i.e. spatial scales from ten to hundreds of kilometers), previous studies focused on the convergence of passive particles in eddies with lifetimes ranging between few weeks to few months (Lebreton et al., 2018; van Sebille et al., 2020). Mesoscale eddies have the capacity to retain particulate material such as radioactive isotope markers (Budyansky et al., 2015) or marine wildlife (Berline et al., 2013; Johnson et al., 2005), and their drift can lead to the transport of passive particles over thousands of kilometers. At sub-mesoscale (i.e. spatial scales under ten kilometers), the study of transport processes is a challenge due to the small size and ephemeral nature of the associated structures, while these processes are known to play a crucial role in the transport and dispersion of particles (McWilliams, 2019, 2016; Poje et al., 2014). Sub-mesoscale fronts, filaments, and vortices provide a fast and efficient route of particulate material transport within the mixed layer (Suaria et al., 2022) and can be associated to several mechanisms such as tides, slope current and shelf-break interactions, upwelling/downwelling, river plumes, eddies or internal waves (Ayouche et al., 2020; Mann and Lazier, 2005). The intense convergence of horizontal flow induced by the sub-mesoscale dynamics can impact the distribution of buoyant material, thereby trapping or collecting floating material as shown in D'Asaro et al. (2018), Huntley et al. (2015) and Hernández-Carrasco et al. (2018a).

CCS have the capacity to concentrate various particulate material, including marine litter (Cózar et al., 2021), larvae, algae, and organic matter such as phytoplankton (Hernández-Carrasco et al., 2018a). The accumulation in CCS can enhance biological productivity, increasing the chance of litter ingestion or entanglement, in line with what is discussed in Gove et al. (2019). Our understanding of ocean currents can help in determining the pathways and fate of floating marine litter, facilitating the prediction of accumulation areas driven by CCS and guiding policy actions to mitigate the impacts of marine litter. In addition, improving our comprehension of processes inducing CCS and their impact on the transport of particulate material by ocean currents is essential in physical oceanography, as these particles can serve as pathways indicator for understanding ocean circulation at sub-mesoscale.

In the southeastern Bay of Biscay (hereinafter BoB), coastal waters are considered as accumulation zone for floating marine litter (Pereiro et al., 2019; Rodríguez-Díaz et al., 2020). Coastal currents are highly influenced by seasonal variations, with weaker currents during the

spring and summer months, leading to higher retention of particulate material (Declerck et al., 2019). In the observation-based study of Ruiz et al. (2020), CCS with large accumulation of floating material were documented. These CCS, referred to by the authors as "marine litter windrows" are identifiable to the naked eye as rectilinear lines parallel to the coast spanning over one kilometer in length. They are found to concentrate  $10^4$  more marine macrolitter items than the surrounding waters (Ruiz et al., 2020; Basurko et al., 2022).

For better understanding the processes underlying the formation of CCS, often visible in the BoB, it is necessary to improve the capability of detection of these phenomena from observational data. Surface fronts are commonly detected by satellites using infrared and microwave sensors (Prants, 2022). Identification can be made based on the calculation of gradients by detecting rapid changes in the physical properties of water, such as temperature, salinity or density (Belkin and O'Reilly, 2009; Chapman et al., 2020; Xing et al., 2023; Zainuddin et al., 2006). However, these techniques are limited by the spatial resolution of the satellite images and by the presence of cloud cover. In the Lagrangian framework, and when surface currents at the desired spatio-temporal resolution are available, surface fronts can be detected by computing the Lagrangian Fronts, maximum of Lagrangian indicators (Prants et al., 2014a, 2014b). Lagrangian indicators correspond to functions of trajectory of a single fluid particle such as finite-size Lyapunov exponents (FSLE, d'Ovidio et al., 2004; Della Penna et al., 2017; Hernández-Carrasco et al., 2011; LaCasce, 2008). In this study, we propose to adopt a Lagrangian approach based on the computation of absolute dispersion (Berti et al., 2011; Bouzaïene et al., 2020; Davis, 1983; Enrile et al., 2019) and FSLE using surface current velocity fields from observation and modelling.

Backward-in-time FSLE are used to investigate mixing, transport pathways, and to identify attracting Lagrangian Coherent Structures (aLCS) (d'Ovidio et al., 2004; Hernández-Carrasco et al., 2011). Ridgelines of backward-in-time FSLE correspond to aLCS (Hernández-Carrasco et al., 2018a) where surrounding Lagrangian trajectories converge. Hence, ridgelines of aLCS can be used as a proxy of Lagrangian convergence regions (Hernández-Carrasco et al., 2018a; Huntley et al., 2015), where larger concentration of marine litter is likely to occur. While FSLEs and aLCS seem appropriate for identifying regions with potential floating marine litter concentration, their capacity to accurately identify CCS locations and characteristics strongly relies on the underlying Eulerian fields used to estimate these Lagrangian quantities. Even more crucial is the application of these techniques in coastal areas, where dynamics are inherently complex. The majority of research has predominantly concentrated on the open ocean, with fewer studies reporting on the use of these methods in coastal regions. (Ghosh et al., 2021; Suaria et al., 2020).

The aim of this study is to improve the efficiency of the CCS identification in the southeastern BoB, by the fusion of data from HFR surface velocity measurements and Lagrangian drifters, assessed against Chlorophyll-a (Chl-a) satellite images. Using the best representation of CCS from optimized HFR data, and Lagrangian diagnostics directly computed from the in-situ Lagrangian drifters, the characteristics of CCS in the area are described for two different periods of time under contrasting hydrodynamic conditions.

## 2. Study site and hydrodynamic conditions

The coastal circulation in the southeastern BoB has a complex pattern. Along the slope, the Iberian Poleward Current, a regional

component of the North Atlantic Eastern Boundary Current System, flows eastward, along the Spanish coast, and northward, along the French coast. It exhibits a relatively intense northward flow (0.4–0.7 m/s at the surface) during late autumn and winter, while the flow weakens and becomes more variable, moving mostly eastward in summer (Charria et al., 2013; Rubio et al., 2019; Solabarrieta et al., 2014).

On the shelf, the circulation is primarily driven by seasonal variations in winds (González et al., 2004; Solabarrieta et al., 2015). Autumn and winter are affected by prevailing southwesterly winds that reinforce the northward flow, while spring and summer bring northeasterly winds, resulting in a weaker and highly variable eastward flow. Due to the narrow shelf and relatively low river discharges, wind-driven currents dominate tidal or density-driven currents over the shelf (González et al., 2004; OSPAR Commission, 2000). The Adour River is the primary source of runoff (Ferrer et al., 2009), supplemented by the Gironde River, located 170 km north of the study area, and much weaker river runoff sources along the Spanish coast. Persistent mesoscale eddies are recurrently triggered by the interaction of the Iberian Poleward Current with the abrupt bathymetry, a phenomenon well-documented in previous studies (e.g. Pingree and Le Cann, 1992; Rubio et al., 2018). As a result, transport and retention conditions at the surface exhibit inter-annual, seasonal, and daily variability, influenced by general circulation, wind-driven currents, mesoscale structures, and fronts (Rubio et al., 2020). Shorter time-scale variability is caused by tidal and inertial motions (Rubio et al., 2011).

The region faces several pressing concerns, which extend to the broader BoB area. These include pollution from hazardous substances near urban and industrial areas such as Bilbao and San Sebastian, the rising incidence of toxic algal blooms causing eutrophication, biodiversity loss, and the proliferation of invasive species (Borja et al., 2019). The southeastern BoB has been identified by global and regional models as an accumulation zone for marine litter (Pereiro et al., 2018; Rodríguez-Díaz et al., 2020). Recent studies on both macrolitter (Ruiz et al., 2020; Basurko et al., 2022), microlitter at the water surface, and biota (Davila et al., 2021) have shed a light on litter quantities and behaviors within the BoB. These recent studies emphasize the hypothesis of the Bay being an accumulation zone and they also highlight its large seasonal variability (Declerck et al., 2019; Pereiro et al., 2019; Ruiz et al., 2022).

The existence of the EuskOOS HFR network in the study area, in combination with numerical model outputs and satellite data, provides extensive historical and near real-time information on the regional ocean-meteorological conditions. It also offers opportunities for the development of Lagrangian studies at meso and upper sub-mesoscale levels, particularly in areas where CCS act as barriers to transport of particulate material.

### 3. Data and methods

#### 3.1. Current velocity from surface drifters and ADCP

Two in-situ surveys were conducted in coastal waters of the southeastern BoB: the BOBLIT0.1 survey in spring (hereinafter referred to as S1) and the BOBLIT0.2 survey in autumn (hereinafter referred to as S2). Both surveys were conducted in the same geographical area, where occurrence of coastal CCS had been reported. During S1, from April 26th 11:45 UTC to 28th April 07:00 UTC, 13 surface drifters were deployed in three different clusters (Fig. 1, red dots), within visible CCS where qualitative aggregation of foam and macrolitter items was observed (Fig. 2). Three drifters were deployed in a longitudinal transect at  $-2.4^{\circ}\text{E}$  in cluster 1 (Fig. 1b), five other drifters were deployed in cluster 2 (Fig. 1b) centered at  $43.4^{\circ}\text{N}$ ,  $-2.25^{\circ}\text{E}$  and the last five were deployed in cluster 3 centered at  $43.35^{\circ}\text{N}$ ,  $-2.3^{\circ}\text{E}$ . The mean separation distance between drifters was 10, 1.7 and 0.6 km in clusters 1, 2 and 3 respectively.

During S2, conducted from October 24th 14:00 UTC to October 28th 03:00 UTC, no visible CCS were present. Thus, nine Lagrangian drifters were deployed in a rectangular area (15 km long and 12 km wide), with an initial separation of 3 km between drifter pairs (Fig. 1b). In addition to surface drifters, the surface layer velocities (1 m below the sea surface) were measured using a towed Acoustic Doppler Current Profiler (ADCP, RDI 600 kHz) transect survey, conducted during S2. The ADCP track lasted a total of four hours with a sampling frequency of 1 min and covered an area of  $9\text{ km} \times 6\text{ km}$ , in the southern part of the domain ( $43.3^{\circ}\text{N}$ ,  $2.3^{\circ}\text{E}$ , Fig. 1b – green trajectory).

For consistency, the subsequent analysis covers 40 h period of time starting at the beginning of each survey.

Three types of drifters were used: the coastal Nomad surface buoys, cylinder-shaped manufactured by SouthTek (<https://www.southteksl.com/>), ODI surface buoys formerly manufactured by Albatros Marine Technologies, and a set of cylinder-shaped home-made drifters (whose design was similar to the coastal Nomad drifters). All drifters were equipped with an anchor of 0.5 m long positioned in the water column between 0.8 and 1.3 m depth, allowing them to drift with surface currents. Observed surface current velocities were estimated from drifter trajectories with a timestep of 15 min, nominal period of drifter positioning via GPS. Potential differences in behavior between different drifter designs were neglected in the Lagrangian computation of absolute dispersion and Lagrangian errors, taking into account the short-term window considered (40 h). For the optimization of the Eulerian fields with the Optimal Interpolation (OI) method, the uncertainty associated to the different drifter designs was parametrized using an observational error equal to 0.05 m/s (see Section 3.6).

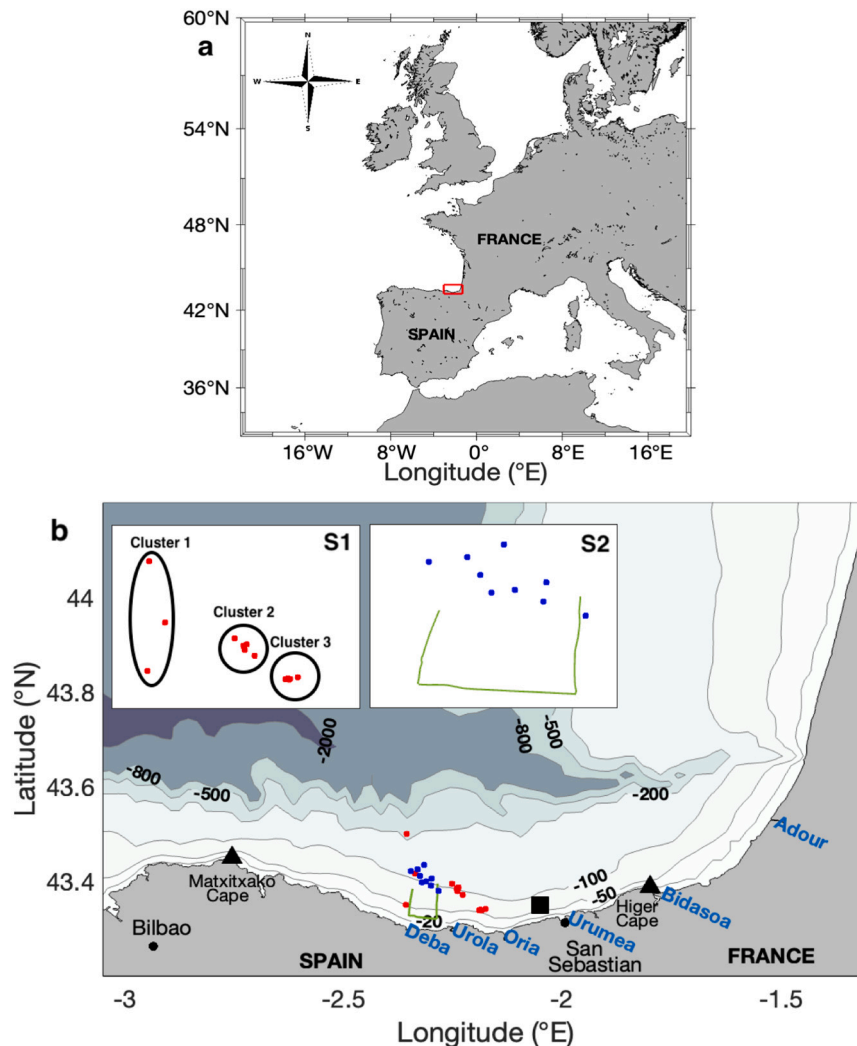
#### 3.2. Current velocities from HFR

Hourly surface velocity current fields were obtained from the EuskOOS HFR network, consisting of two CODAR Seasonde stations located along the Spanish Basque Country coast in the southeastern BoB (Fig. 1). EuskOOS HFR network (<https://doi.org/10.57762/T4WH-DQ48>) is a part of JERICO-RI (<https://www.jerico-ri.eu/>) and it is operated following JERICO-RI standards and recommendations (Mantovani et al., 2020; Rubio et al., 2018; Solabarrieta et al., 2016). The two individual radar stations, operating at a frequency of 4.86 MHz, enable velocity measurements within the range of up to 200 km with 5 km along-beam resolution, and  $5^{\circ}$  angular resolution. Velocity data are obtained at 1 h time step and distributed as part of Copernicus Global Ocean- in-situ near real- and delayed-time surface ocean currents products (<http://doi.org/10.48670/moi-00041&> <http://doi.org/10.17882/86236>).

Using the measured radial velocities, two products of total surface current velocities fields were generated, covering the area from  $-3^{\circ}\text{E}$  to  $-1.3^{\circ}\text{E}$  and from  $43.3^{\circ}\text{N}$  to  $44^{\circ}\text{N}$ . On one hand, the open-boundary modal analysis (OMA, Kaplan and Lekien, 2007) gap-filling method was applied to the HFR radial velocities (Table 1) to obtain hourly OMA current fields with 5 km spatial resolution, during the periods of April–May 2022 and October–November 2022. Only OMA values with the geometric dilution of precision (GDOP) errors below the threshold of 8 cm/s (Solabarrieta et al., 2014) were considered for this study.

On the other hand, the two-dimensional variational interpolation method (2dVar, Yaremchuk and Sentchev, 2009) was applied to HFR radial fields (Table 1) to obtain hourly 2dVar current fields with 2.5 km spatial resolution. 2dVar is a non-local and kinematically constrained interpolation method which generates surface current velocity maps by using a combination of all measured velocities, within a given grid, reconstructing the velocity vector in one location. Estimations of the relative error of interpolation were computed at each time step. 2dVar fields were computed for an extended period from January 2020 to December 2022, since they were also used as historical input for the optimization method. To ensure that the method effectively captured





**Fig. 1.** (a) Study area in the southeastern BoB, represented by the red rectangle. (b) Map and details of the study area. Bathymetry is shown by blue shading. Black triangles show the two HFR stations located at Matxitxako and Higer Capes. The location of major rivers is given in blue. Geographic names used in the text are also shown. Red and blue dots show the release location of drifters during S1 and S2 surveys, respectively, and are presented in a zoom in the upper part of the figure. The ADCP transect conducted during S2 is shown by the green trajectory.

the information derived from the available radial velocities, a careful consideration was given generating the grid for performing the 2dVar interpolation. On average, there were 870 observations collected every hour, with the mean distance of about 5 km between observations within the study area. For this study, we opted for a 2dVar grid consisting of 1500 interpolation points, establishing a 2:1 ratio between the number of observations and the number of interpolated points, in coherence with the 2:1 ratio between the average distance between observations and the grid resolution. The 2dVar method was applied with a noise level set to 0.06 m/s.

### 3.3. Current velocities from IBI model

Model surface current velocities were obtained from the Iberia-Biscay-Ireland (IBI) product (Table 1) provided by the Copernicus Marine Environment Service (CMEMS, doi:10.48670/moi-00027). IBI is a product that covers the northwestern European waters, based on an eddy-resolving three-dimensional NEMO configuration, with a time-space resolution of 15 min and 1/36°. The model is forced with 3-hourly atmospheric fields provided by ECMWF. Lateral open boundary conditions (temperature, salinity, velocities, and sea level) are interpolated from the daily outputs from the MyOcean Global eddy resolving system.

Tidal boundary conditions were obtained from the global tide models FES2004 (Lyard et al., 2006) and TPX07.1 (Egbert and Erofeeva, 2002). Freshwater river discharge inputs are implemented as a lateral open boundary condition for 33 rivers with flow rates based on observations and climatology data. The bottom topography is taken from the GEBCO08 dataset (Maraldi et al., 2013). IBI does not include data assimilation (Mason et al., 2019). The velocity data were extracted from the model outputs within the region of interest defined above.

### 3.4. Wind and satellite data

Wind data were obtained from the Weather Research and Forecasting model (WRF) provided by the meteorological agency of Galicia. The description of the model is detailed in (Skamarock et al., 2019). In this study, hourly zonal and meridional components of wind velocity at a height of 10 m were used, featuring a spatial resolution of 12 km, covering the predefined geographic area.

Remotely sensed surface Chl-a estimates were obtained from the CMEMS catalogue (doi:10.48670/moi-00286). This product derives from the Sentinel-3 Ocean and Land Color Imager (OLCI) spectrometer. The Chl-a concentration used is OC4Me, corresponding to the pigment concentration based on the ocean color algorithm developed for MERIS,



Fig. 2. Photos of one of the coastal CCS observed during S1 survey, where drifters were deployed (easternmost cluster). Aggregation of foam and macrolitter was observed, although not quantified.

and following the approach of O'Reilly et al. (1998), with a spatial resolution of 300 m and a temporal resolution of one day. The Chl-a concentration product used is cloud polluted. Although various algorithms exist to fill the data gaps below clouds (e.g., Stock et al., 2020), these methods generally require much longer time series of Chl-a images, which results in smoother fields, and are not the best suited to study CCS during the short-time periods of S1 and S2 surveys.

### 3.5. Estimation of Lagrangian trajectories and Lagrangian error

Virtual trajectories were computed using the open-source Lagrangian framework OpenDrift (<https://opendrift.github.io/>, see Dagestad et al., 2018; Jones et al., 2016) and the different surface current fields (Table 1). Virtual drifters were seeded at the time and location of the real drifters every hour, and advected using the 4th order Runge-Kutta scheme and bilinear interpolation. The horizontal diffusion parameter was set to 0. The accuracy of virtual trajectories was assessed using the Lagrangian error  $\epsilon_L$  (Liu and Weisberg, 2011; Ruiz et al., 2022). The estimation of  $\epsilon_L$  involves the computation of the separation distance between the location of the real drifter and that of virtual drifter moving in the velocity field.

Table 1

Characteristics of the surface current fields used in the study for computation of Lagrangian error, absolute dispersion and FSLE.

Short name	Spatial resolution (km)	Temporal resolution (h)	Description
OMA	5	1	HFR computed with OMA method from radial hourly surface current velocity field, provided by CMEMS.
2dVar	2.5	1	HFR computed with 2dVar method from radial hourly surface current velocity field, provided by CMEMS.
2dVar-opt	2.5	1	2dVar fields optimized with measurements of surface current velocity by surface drifters through OI (described Section 3.6)
IBI	3	0.25	Model data provided by CMEMS.

Considering  $d_{tk}$ , the separation distance between the real and the  $k$ -th simulated trajectory at time step  $t$ , the Lagrangian error is computed at each time step and averaged over all the simulated trajectories corresponds to each real drifter trajectory, following the expression:

$$\epsilon_L(t) = \left\langle \sum_{t=1}^N \sum_{k=1}^{N-(t+1)} \frac{d_{tk}}{N-(t+1)} \right\rangle \quad (1)$$

with  $N$  corresponding to the maximum number of time steps of drifter displacement, which is also equal to the number of simulated trajectories. The Lagrangian error calculation schema is shown in Appendix A.

To facilitate comparisons, the Lagrangian error was divided by the mean drift distance of the real drifters (travel distance averaged over all the drifters), giving rise to a dimensionless index  $L$ .

The Lagrangian error calculation also included the estimation of the travel distance and separation distance (distance between the real drifter trajectory and the virtual drifter trajectory) of each available surface current field at each time step, considering only the virtual particle deployed at the initial time.

### 3.6. Optimal interpolation of drifter velocity measurements

Optimal Interpolation (OI) was used to merge current velocities obtained from drifting buoy trajectories and 2dVar HFR velocities, resulting in the optimized 2dVar surface current product (hereinafter 2dVar-opt). Following a meticulous assessment of the accuracy of the various available surface current products (refer to Table 1), 2dVar was chosen for optimization due to its higher performance, as discussed in Section 4.1.

By performing a weighted least-squares fit of a background field to the drifter velocities, this method can provide a better estimation of the state of ocean dynamics. OI is generally used when observations are available at irregularly distributed points and are assumed to be affected by an observation error, e.g., drifter measurements. This observation error is assumed to be uncorrelated with 2dVar HFR fields error and is set to 0.05 m/s, which is twice the uncertainty due to drifter positioning. This adjustment ensures that potential errors attributed to the various buoy designs used in sea surveys are adequately accounted for.

OI provides a correction of a background velocity field  $\mathbf{u}_r(\mathbf{x}, t)$  by performing a linear combination of the weighted differences between the background trajectory  $\mathbf{u}_r$ , and the observed drifter velocities  $\mathbf{u}_i^*$  at point  $i$  (Sentchev and Yaremchuk, 2015). Weights chosen for minimization of the mean square difference between drifter and HFR velocities are a combination of 2dVar HFR ( $\mathbf{B} = \langle \mathbf{u}_r(\mathbf{x}, t) \mathbf{u}_r(\mathbf{x}', t') \rangle$ ) and drifter ( $\mathbf{R}_{ij} = \langle \mathbf{u}_i^*(\mathbf{x}, t) \mathbf{u}_j^*(\mathbf{x}', t') \rangle$ ) velocity covariances. The optimized velocities  $\mathbf{u}_{opt}$  at radar grid are computed as follows:

$$\mathbf{u}_{opt} = \mathbf{u}_r + \sum_j \mathbf{B} \mathbf{H}_j^T (\mathbf{H}_i \mathbf{B} \mathbf{H}_j^T + \mathbf{R}_{ij})^{-1} (\mathbf{H}_i \mathbf{u}_r - \mathbf{u}_i^*), \quad (2)$$

where  $\mathbf{H}_i$  corresponds to a linear operator projecting gridded velocity values from the apexes of the radar grid cell onto the  $i^{\text{th}}$  observation point location.

The performance of each velocity field, and particularly the quality of the interpolation scheme, is quantified by estimating the Root Mean Square Error (RMSE), corresponding to the mean relative difference between the drifter velocities  $\mathbf{u}^*$  and the desired velocity field  $\mathbf{u}$  as follows:

$$RMSE = \sqrt{\sum_i (\mathbf{H}_i \mathbf{u} - \mathbf{u}_i^*)^2}. \quad (3)$$

The success of OI relies on the background field, which should ideally be a close approximation to the truth. Nonetheless, the outcome is also linked to the composition of the covariance matrix, comprising a large number (80) of 40-h long trajectories known as ensemble members. These ensemble members are expected to represent similar circulation patterns as those observed during the survey (background current velocities and tidal phase).

For selecting ensemble members, the non-linear K-Means clustering algorithm (Hastie et al., 2009; Solabarrieta et al., 2015) is employed to analyze 2dVar HFR derived velocities from January 2020 to December 2022. The primary objectives of using this method are twofold: (i) to identify and categorize the dominant surface current groups present during both survey periods, and (ii) to extract ensemble members with similar dominant groups in the extended time series data, which are subsequently used to create the covariance matrix required for OI. The K-Means algorithm employs a fixed number of clusters, which was set to 12, with further details provided in Appendix B. Additionally, the number of ensemble members used was set to 80. Sensitivity tests on the number of ensemble members to OI results are presented in Appendix B.

An additional assessment of the performance of the 2dVar reconstruction using the given approach was completed by the comparison of the results with real trajectories (see Section 4.1) and by the examination of the interpolation errors with time and as the function of the number of available HFR observations (not shown). Although the relative error of 2dVar interpolation is sometimes relatively large (up to 0.7), it shows an overall mean of 0.5 during S1, and 0.4 during S2. During S1, the evolution in time of the interpolation error was found to be inversely proportional to the number of available HFR observations, with a correlation coefficient between the interpolation error and the number of available observations of  $-0.5$ . However, this was not the case during S2 where both quantities were found uncorrelated (correlation coefficient equal to  $-0.2$ ). This difference can be explained by the different dynamics observed during the two campaigns. During S1, surface current velocities were low and featured numerous eddies, more difficult to measure accurately by HFR, which explains the correlation between the number of observations and the 2dVar error. On the other hand, during S2, surface current velocities were driven by a strong coastal jet, much easier to measure by HFR, even in situations when the number of observations is not optimal.

### 3.7. Lagrangian metrics to estimate spreading and convergence of trajectories

The absolute (cluster) dispersion  $A^2(t)$  is generally used to quantify the intensity of the particle spreading induced by the surface current field.  $A^2(t)$  is defined as the variance of particle spreading with respect to the mean coordinate of particles in a cluster (the barycenter). In two-dimensions, the dispersion can be estimated along  $x$  and  $y$  axis (Berti et al., 2011; Enrile et al., 2019) following the expression:

$$a_{ij}^2(t) = \frac{1}{M} \sum_{m=1}^M \left\{ [x_i^m(t) - \bar{x}_i^m(t)] [x_j^m(t) - \bar{x}_j^m(t)] \right\} A^2(t) = a_{xx}^2(t) + a_{yy}^2(t). \quad (4)$$

Here,  $a_{ij}^2$  corresponds to the variance along  $i$  and  $j$  spatial coordinates,  $a_{xx}$  and  $a_{yy}$  corresponds to the absolute dispersion along  $x$  and  $y$  respectively,  $M$  is the number of drifters,  $x^m(t)$  is the position of the  $m$ -th drifter at time  $t$  and  $\bar{x}^m(t)$  is the coordinate of the barycenter.

Estimation of the absolute dispersion allows to identify two fundamental regimes of dispersion in the turbulent flow. At short times, i.e. during the first few hours of the drifter deployment, the growth of absolute dispersion follows generally a ballistic regime corresponding to  $A(t)^2 \sim t^2$  (Berti et al., 2011; Bouzaiene et al., 2020; Davis, 1983; Enrile et al., 2019; Poulain and Niiler, 1989; Taylor, 1922). This regime occurs when the pair velocities of the drifters are strongly correlated, influenced by strong and persistent currents such as a coastal jet. At larger timescales, when the memory of the initial condition is lost, and the absolute dispersion can grow as  $A(t)^2 \sim t$  (Berti et al., 2011; Bouzaiene et al., 2020; Davis, 1983; Enrile et al., 2019; Poulain and Niiler, 1989; Taylor, 1922), following a random-walk regime of dispersion.

The FSLE technique is used to analyze the ocean transport and mixing (Aurell et al., 1997; d'Ovidio et al., 2004; Hernández-Carrasco et al., 2011; LaCasce, 2008). The largest FSLE values concentrate along characteristic lines, identifying regions of maximum stretching, which corresponds to oceanic structures such as fronts, eddies boundaries, etc. As the aLCS cannot be crossed by particle trajectories, they act as barrier to transport, governing therefore the motion of the surrounding fluid particles around them. FSLE values, given by  $\lambda$ , are estimated as the inverse of the time  $\tau(\mathbf{x})$  required for two particles of fluid to separate from an initial distance  $\delta_0$  to a final distance  $\delta_f$ , and is expressed at position  $\mathbf{x}$  and time  $t$  as:

$$\lambda(\mathbf{x}, t, \delta_0, \delta_f) = \frac{1}{\tau(\mathbf{x})} \ln \alpha = \frac{1}{\tau(\mathbf{x})} \ln \frac{\delta_f}{\delta_0}. \quad (5)$$

In this study, values of  $\lambda$  were computed using the algorithm described in Hernández-Carrasco et al. (2011) with the amplification factor  $\alpha = 8$  in order to measure high stretching as well as to obtain reliable aLCS. The parameter  $\delta_0$  is fixed to 0.4 km to capture fine-scale structures. Thus,  $\delta_f = \alpha \delta_0 = 3.2$  km. Particle trajectories were integrated backward-in-time over 15 days, since the associated aLCS have a direct physical interpretation (i.e. nearby particle trajectories lie along the hyperbolic aLCS). This enables us to understand the fate and pathways of marine litter and identify potential CCS (Hernández-Carrasco et al., 2018a, 2011). A sensitivity test to the number of days of integration and to the amplification factor was conducted. We obtained that 15 days of integration and an amplification factor of 8 were a good choice to obtain reliable aLCS and detect transport barriers.

Note, however, that the drifter trajectories in both experiments had a relatively brief duration of approximately 40 h, providing only 40 h of 2dVar-opt velocity fields. The backward-in-time FSLE at  $t$  is computed using backward trajectories from  $t$  to  $t - 15$  days, relying on the velocity field from  $t$  to  $t - 15$  days. In instances where the 2dVar-opt does not exist ( $t$  inferior to the deployment time), 2dVar velocity field is used for the FSLE calculation.

## 4. Results

### 4.1. Optimization of surface current fields for dispersion quantification and CCS identification

Before optimizing the surface velocity fields, an extensive comparison exercise was conducted to assess the capabilities of various available Eulerian fields in reproducing the dynamics observed by drifting buoys (as explained in the previous section). This exercise led to the identification of the 2D Variational horizontal currents (2dVar) solution as the field to be optimized. The following results showcase the comparison of different current fields, including the 2dVar-opt field, with the aim to illustrate how the latter provides the best fit to drifting buoy data.

Above all, the performance of the optimization method was assessed by comparing velocity fields from different datasets with independent velocity measurements. Surface layer velocities (at 1 m depth) from a towed ADCP transect survey, conducted during S2, were used in comparison. The relative error between ADCP velocities and those derived from HFR measurements, interpolated into the ADCP track and sampling frequency, before and after optimization was reduced by a factor of two, from 0.26 to 0.11. In particular, the current direction was considerably improved. Then, other quantities were estimated to show improvements brought by data fusion.

In a qualitative approach, Fig. 3, provides two examples of comparison between real and virtual trajectories computed from the different surface current fields during S1 and S2 surveys. During S1 (Fig. 3a), the 2dvar-opt trajectory closely matches the real one in terms of separation distance, although not in terms of travel distance since it fails to reproduce the amplitude of the loops (due to a combination of tidal and inertial processes) exhibited by the real trajectory. OMA and 2dVar trajectories are mostly rectilinear and follow the 80-100 m isobaths. While the OMA trajectory does not show any loop, the 2dVar trajectories do depict small amplitude oscillations within the first hours after the beginning of the computation. The IBI trajectory is the one that shows the largest separation distance since it is advected in the opposite direction to the real drifter, depicting oscillations with similar amplitude of that of the real trajectory (Fig. 3a).

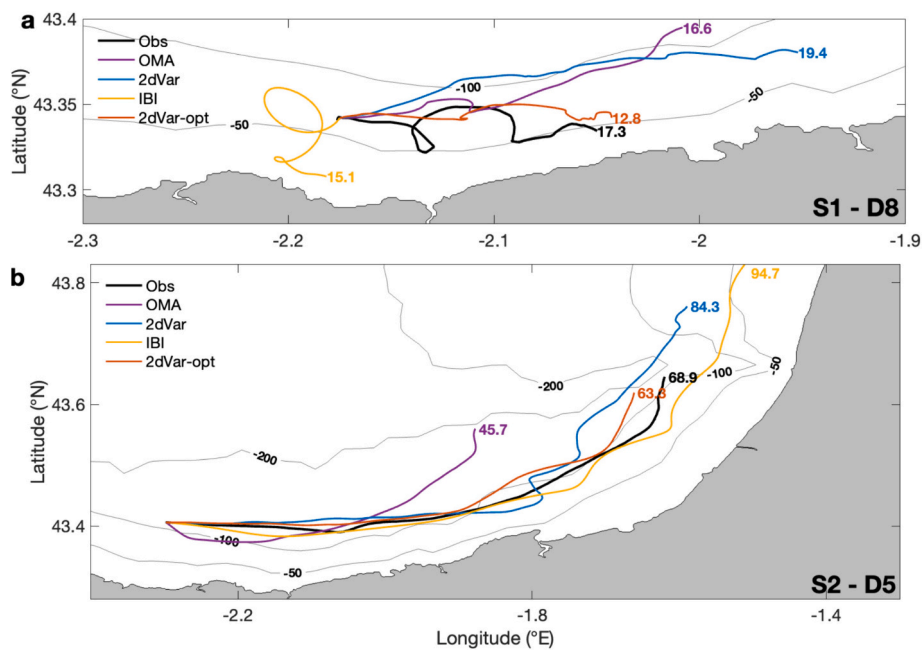
During S2 survey (Fig. 3b), the mean travel distance covered by the drifter was 68.9 km. In this period, the observed currents were much more intense and less variable, showing all the trajectories a net eastward drift along the 100 m isobath, with exception of the OMA trajectory. The 2dVar-opt trajectory is the closest to observations with a travel distance of 63.3 km. Travel distances are overestimated regarding IBI and 2dVar trajectories and underestimated regarding OMA trajectory. The 2dVar-opt trajectory follows the 100 m isobath, with very small separation distances from the real trajectory through all the paths. IBI and OMA trajectories show the worst fit.

Different statistics for real surface drifters during S1 and S2 surveys and the different current fields are presented in Table 2. Mean and maximum velocity values, mean travel distances, L index and RSME are computed from real and virtual trajectory pairs. Regarding mean velocity, all velocity fields yield similar results for S1 and S2 and are in accordance with the values measured by surface drifters. However, there is typically a visible difference of 0.1 m/s in maximum velocity when comparing these values with the data from surface drifters (Table 2). During S1, all the surface current fields underestimate the maximum velocity measured by the drifters by 0.1 m/s. During S2, 2dVar and IBI underestimate the maximum velocity by 0.1 m/s while OMA and 2dVar-opt give the same results as drifter measurements.

**Table 2**

Mean and maximum current speed, the average travel distance calculated from all the pairs of real and virtual trajectories pairs. L index and RMSE calculated using the OMA, 2dVar, IBI and 2dVar-opt fields.

		Drifters	OMA	2dVar	IBI	2dVar-opt
Mean velocity (m/s)	S1	0.1	0.1	0.1	0.1	0.1
	S2	0.3	0.3	0.3	0.3	0.3
Max velocity (m/s)	S1	0.3	0.2	0.2	0.2	0.2
	S2	0.5	0.5	0.4	0.4	0.5
Mean travel distance (km)	S1	19.5	16.7	14.6	18.4	12.9
	S2	62.0	46.3	80.0	95.6	61.0
L index	S1	-	0.6	0.5	0.7	0.3
	S2	-	0.3	0.2	0.3	0.1
RMSE	S1	-	1.3	0.9	1.4	0.7
	S2	-	0.6	0.4	0.5	0.3



**Fig. 3.** Real and virtual drifter trajectories from the 2dVar, IBI, OMA and 2dVar-opt fields during S1 (a) and S2 (b). Results are presented for drifter 8 of S1 and drifter 5 of S2. The travel distance is given in km at the end of each trajectory. The trajectory shown represents the typical behavior observed on all drifters during each survey. Zooms of (a) and (b) have been adjusted to provide the best representation of the trajectories.

During S1, the mean travel distance by drifters is 19.5 km. While OMA and IBI trajectories tend to give results close to observations, 2dVar and 2dVar-opt trajectories underestimate the travel distance. However, 2dVar-opt field respects the shape of the real trajectory (see Fig. 3a), which is not the case of 2dVar, OMA and IBI. During S2, the travel distance increased up to 61.6 km. Table 2 summarizes the results of the comparison of the different solutions used and shows that the 2dVar-opt field is the closest to observations. In contrast, IBI appears to underestimate the travel distance, while OMA and 2dVar tend to overestimate it.

Regarding the  $L$  index and RMSE, the 2dVar field once again appears to be the most realistic compared to OMA and IBI fields. The  $L$  index of the 2dVar field is 31 % smaller than that of IBI, and 15 % smaller than that of the OMA field during S1. During S2, the  $L$  index of the 2dVar field is 23 % smaller than that of IBI and 11 % smaller than that of OMA. In terms of RMSE, the RMSE of 2dVar is 29 % smaller than that of IBI and 21 % smaller than that of OMA for S1. Similarly, for S2, it is 16 % smaller than that of IBI and 32 % smaller than that of OMA. This reinforces confidence in the 2dVar results. 2dVar-opt shows the lowest values of the index  $L$ , and RMSE, improving those of the 2dVar field. Regarding  $L$ , this reduction attains 30 % during S1 and 63 % during S2. Regarding the RMSE, the average decrease is 23 % during S1 and 35 % during S2.

When assessing the errors of the different surface velocity fields with time (Fig. 4) we observe that the 2dVar-opt field still appears to be the most realistic regarding the time series of the Lagrangian error index  $L$  (Fig. 4a, b) and the RMSE (Fig. 4c, d) as stated in Table 2. During S1, the Lagrangian error index of all the surface current fields peaked around 5 h, at the time where surface currents were low, and drifters were subject to tidal and inertial oscillations. The RMSE was also larger during the first few hours of deployment for all the surface current fields. During S2, there is no such peak found in  $L$  index and RMSE time series, surface drifters being advected by the coastal jet and not subject to tidal or inertial oscillation. However, during both surveys and regarding both  $L$  and RMSE, 2dVar and 2dVar-opt are systematically presenting the best performance compared to IBI and OMA.

With the aim of assessing the impact of the OI spatially in the area used for FLSE maps computation, Fig. 5 shows the difference in velocity resulting from the application of OI to the 2dVar fields corresponding to the period of surveys S1 and S2. First, during S1, the mean velocity field shows the presence of eddies in the study area, with low velocities and a coastal current oriented mainly along the Spanish coast. During S2, the mean 2dVar-opt velocity field presents larger velocities (see Table 2), mainly oriented alongshore. During both surveys, the largest difference values between the initial and optimized velocity fields are found in the coastal region 20–30 km large. During S1, the mean absolute difference is close to 0.05 m/s (Fig. 5a). 2dVar field is characterized by the mean

and maximum velocity of 0.08 m/s and 0.5 m/s respectively, while for 2dVar-opt the respective values are 0.1 m/s and 0.4 m/s (Table 2). Therefore, the OI of drifter velocities induced modifications in the remotely sensed velocity fields up to 20 %, on average. Because coastal circulation during S1 was characterized by low velocities, inertial and tidal oscillations, the major changes were obtained in the eastern and central parts of the study region, covered by drifter observations (Fig. 5a). In contrast, during S2, the mean absolute difference between initial and optimized fields is twice larger, 0.09 m/s. Larger differences, 0.15–0.20 m/s, are found in the nearshore region, extending up to 20 km offshore (Fig. 5b). However, significant modifications in the velocity field (up to 0.1 m/s) observed at far ranges, ~40 km, demonstrate the global effect of drifter measurements near the shore on the velocity field within the whole study region.

For the 2dVar field, the mean and maximum surface current velocity are 0.2 m/s and 0.5 m/s, while in 2dVar-opt field they became respectively 0.3 m/s and 0.6 m/s (Table 2) indicating an increase of 45 %. The larger disparities observed during S2 can be related to higher current speed inside the current jet along the coast (Fig. 5b) occurring during this period of the year (Rubio et al., 2019).

Finally, as shown in Appendix C, the 2dVar-opt field proves to be the most accurate when compared to the absolute dispersion of virtual particles and to FSLE maps derived from alternative surface current fields. The results underline the accuracy of the 2dVar-opt field, which has the lowest Lagrangian and relative errors. Based on these results, the 2dVar-opt field will be used for further assessing CCS.

#### 4.2. Absolute dispersion and CCS identification

Current velocity fields during the two survey periods revealed very different coastal dynamics largely influenced by the local wind and remote forcing.

S1 took place during a period of low winds with a mean speed of 3 m/s (Fig. 6a). During this period, surface currents were weak, and some drifter trajectories showed the influence of tidal motions and inertial oscillations (Fig. 6c). During S1, up to five local CCS were identified by the naked eye (not shown), in the form of rectilinear lines parallel to the coast in the area where the westernmost cluster of drifters was deployed. The local CCS observed closest to the coast (showcased in Fig. 2, and which determined the location of the third cluster of drifters) was the only one showing foam. However, macro litter aggregation was (qualitatively) observed in several of them. The presence of CCS is also showcased in the Chl-a satellite images available during S1 survey dates (Fig. 6c). High values of Chl-a concentration are found close to the coast and associated with the presence of the main river plumes. The first alongshore line of high Chl-a values coincides with the CCS with foam

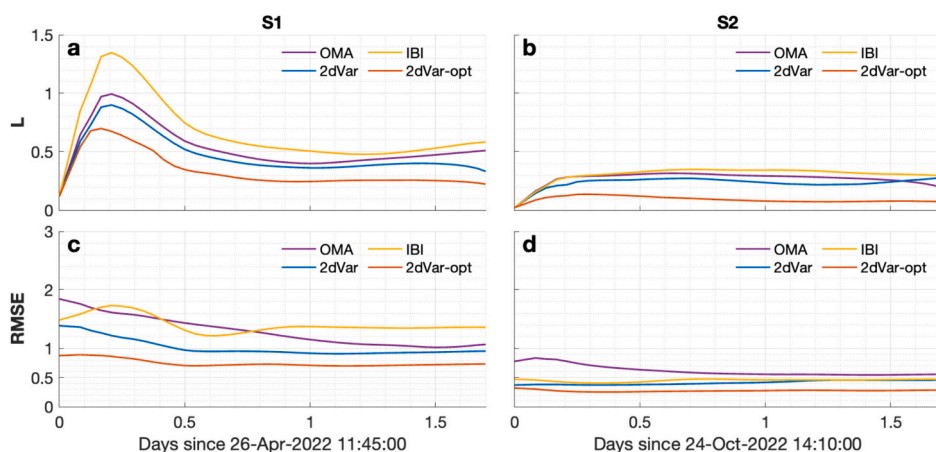
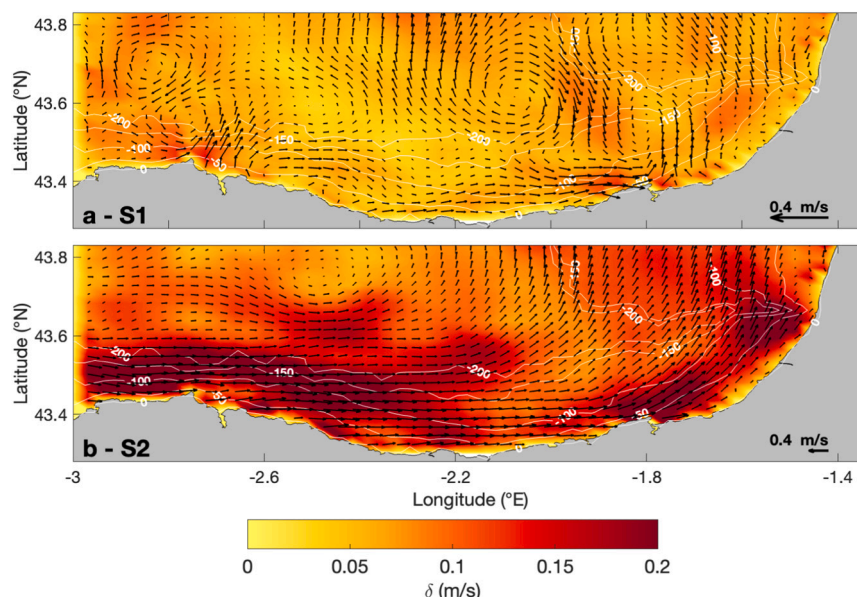


Fig. 4.  $L$  index (a, b) and RMSE (c, d) computed for the available fields. Figures related to S1 are given in the left column, while figures related to S2 are given in the right column.



**Fig. 5.** Map of the differences induced by OI on the 2dVar fields. Red vectors represent the mean surface velocity from the 2dVar-opt field. Velocity scale is different for the two periods and is given by the reference arrow in the right bottom corner of each panel. Color shading shows the absolute averaged velocity differences  $\delta = |\mathbf{u}_r - \mathbf{u}_{opt}|$  between 2dVar and 2dVar-opt fields over 40-h corresponding to the dates of S1 and S2 surveys, with the same color scale.

observed closest to the coast, and with the initial location of the drifters in the third cluster (Fig. 6c,  $-2.2^\circ\text{E}$ ;  $43.3^\circ\text{N}$ ). This high Chl-a strip could result from the convergence of the plumes of the Deba, Urola and Oria rivers around  $-2.05^\circ\text{E}$ ;  $43.33^\circ\text{N}$  (Fig. 6c). Chl-a concentrations observed are above  $1 \text{ mg}\cdot\text{m}^{-3}$ . This particular area corresponds also to the zone of convergence of three drifters observed during S1, shown in red in Fig. 6c. Two pairs of drifters, initially separated by 8 and 16 km, were found separated by  $<2 \text{ km}$  after a 40-h drift, which evidences convergence in the vicinity of the outer limit of the river plumes.

S2 took place under strong southwards winds, with a mean wind speed of  $6 \text{ m/s}$  (Fig. 6b). The current jet observed along the coast during this surveying period is at the origin of mostly rectilinear drifter trajectories (Fig. 6d). No CCS were identified during the survey, although (with much less extent than in S1) high Chl-a concentrations are found in the area around Deba and Urola rivers (where the drifters were deployed), and along the French coast. The S2 drifters did not encounter any visible local CCS along their drift, as observed in the Chl-a image for the corresponding date (Fig. 6d), at least until they arrived in the waters off the French coast, where the distribution of high Chl-a concentrations shows the Adour plume. No specific points of convergence for the drifters in S2 can be identified from the qualitative analysis of their trajectories (Fig. 6d).

To further investigate the influence of CCS on horizontal mixing, the absolute dispersion was computed based on analysis of real and virtual drifter trajectories (13 during S1, 9 during S2) reconstructed from 2dVar-opt fields. The absolute dispersion is computed along  $x$ - and  $y$ -axis directions (globally following the orientation of the coastline). First, the absolute dispersion computed from 2dVar-opt trajectories is consistent with observations regarding dispersion regimes, and dispersion rate in  $x$  and  $y$  directions. Nevertheless, an underestimation of the absolute dispersion by 37 %, during S1, and by 57 %, during S2, was found when 2dVar-opt trajectories were used, compared to that of real trajectories.

During the first 5 h of S1, the dispersion is nearly isotropic, with slightly larger alongshore dispersion, and the absolute dispersion follows a ballistic regime with  $A^2(t) \sim t^2$  (Fig. 7a, c). Then, starting from 6 to 7 h, the dispersion becomes anisotropic with ten times larger dispersion in the alongshore direction (along  $x$ -axis). During this period of time, the absolute dispersion curve demonstrates equivalent to the

diffusive dispersion regime with  $A^2(t) \sim t$ . This change in regime is due to the loss of correlation between the different trajectories as the real and virtual travel distances increase.

During S2, the dispersion is strongly anisotropic. The dispersion rate in the alongshore direction is found to be twelve times larger than in the cross-shore direction. The time evolution of absolute dispersion is found fundamentally different. For almost the entire period of S2, turbulent motions are characterized by ballistic regime with  $A^2(t) \sim t^2$  (Fig. 7b, d). This dispersion regime is representative for powerful current systems such as the current observed along the coast.

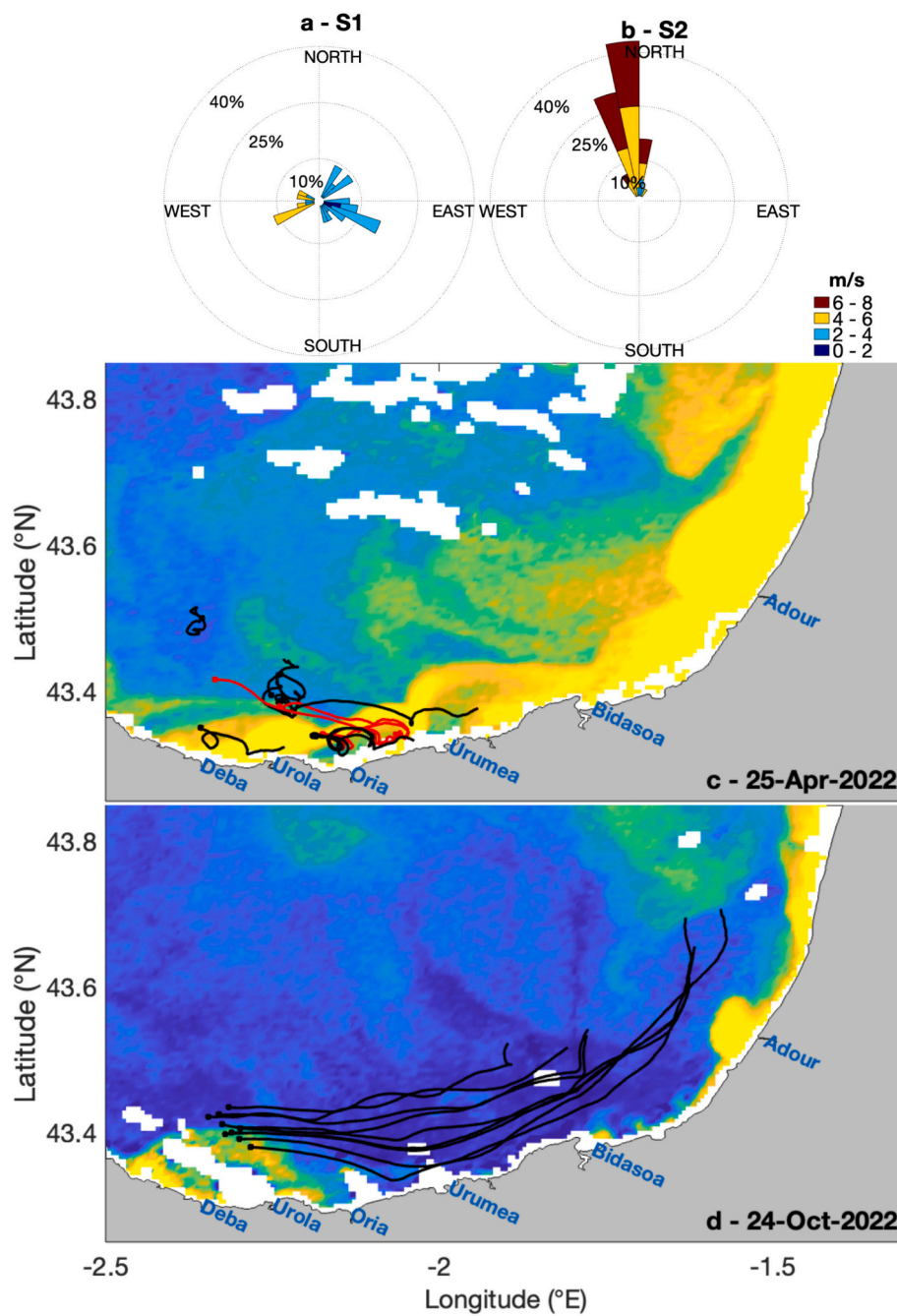
During S1 and S2, high current velocities observed influence the distribution of CCS and suggest strong dynamics along the coast, compatible with the presence of alongshore CCS.

To characterize the spatial distribution of CCS, 24 hourly FSLE maps were generated from 2dVar-opt and averaged over one day corresponding to S1 and S2 surveys (Fig. 8a, b). The spatial distribution of large FSLE values, revealing the presence of intense horizontal stirring in the turbulent flow, were qualitatively compared with Chl-a concentrations for the same period (Fig. 8c, d for the same dates, and e, f for previous dates with better coverage).

During S1, the FSLE ridgelines, located at short distance from the shore, are aligned with the Spanish coast (Fig. 8, red square), supporting that surface current structuring is affected by a coastal downwelling. During S2, less FSLE ridgelines are found close to the Spanish coast, showing lower values. This is probably because at the surface, the current velocities were larger, leaving no time for particles to accumulate near the Spanish coast. During S2, CCS are mostly found parallel to the French coast (Fig. 8, red triangle). During both surveys, in the middle of the study area, further from the coast, the FSLE ridgelines are less organized and sometimes meandering.

From the examination of hourly FSLE maps during both S1 and S2, coastal CCS persist during the whole time of the surveys and slightly move in the dominant surface current direction, i.e. northeastward (not shown).

During S1, the coastal CCS, where three drifter trajectories converged (Fig. 6c, red trajectories), is shown in all panels of Fig. 8 by red square. This location corresponds to aLCS, which coincides with a Chl-a front and where a local CCS was identified to concentrate foam and macrolitter. FSLE ridgelines (red lines in Fig. 8) delimit the regions



**Fig. 6.** Wind roses for (a) S1 and (b) S2 from the WFR model averaged over the S1 and S2 time periods and spatially-averaged over the study region. Trajectories of Lagrangian drifters released during (c) S1 and (d) S2 superimposed to the satellite Chl-a concentration from CMEMS at 300 m resolution on (c) April 25, 2022, and (d) October 24, 2022. Location of the main rivers is given by blue characters. The color bar is given in log scale. Drifter trajectories are shown in black, with a dot indicating the deployment location. The three drifters converging during S1 are shown in red in (c).

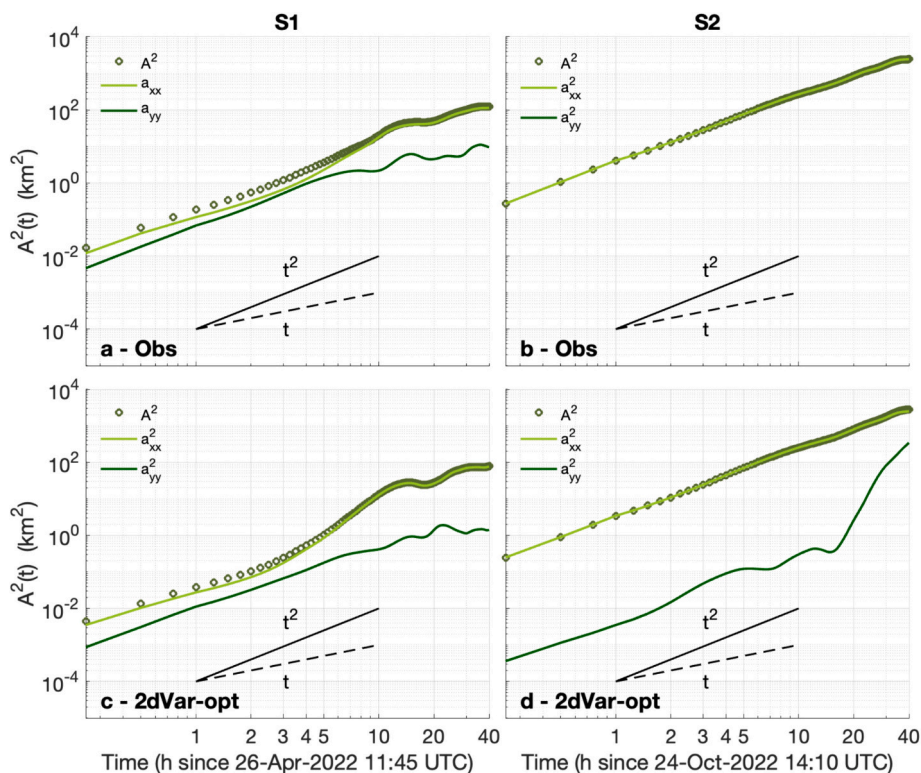


Fig. 7. Absolute dispersion considering observed drifter trajectories (a, b) and 2dVar-opt trajectories (c, d) during S1 and S2 surveys. The ballistic ( $t^2$ ) and diffusive ( $t$ ) regimes are shown by solid and dashed black lines, respectively.

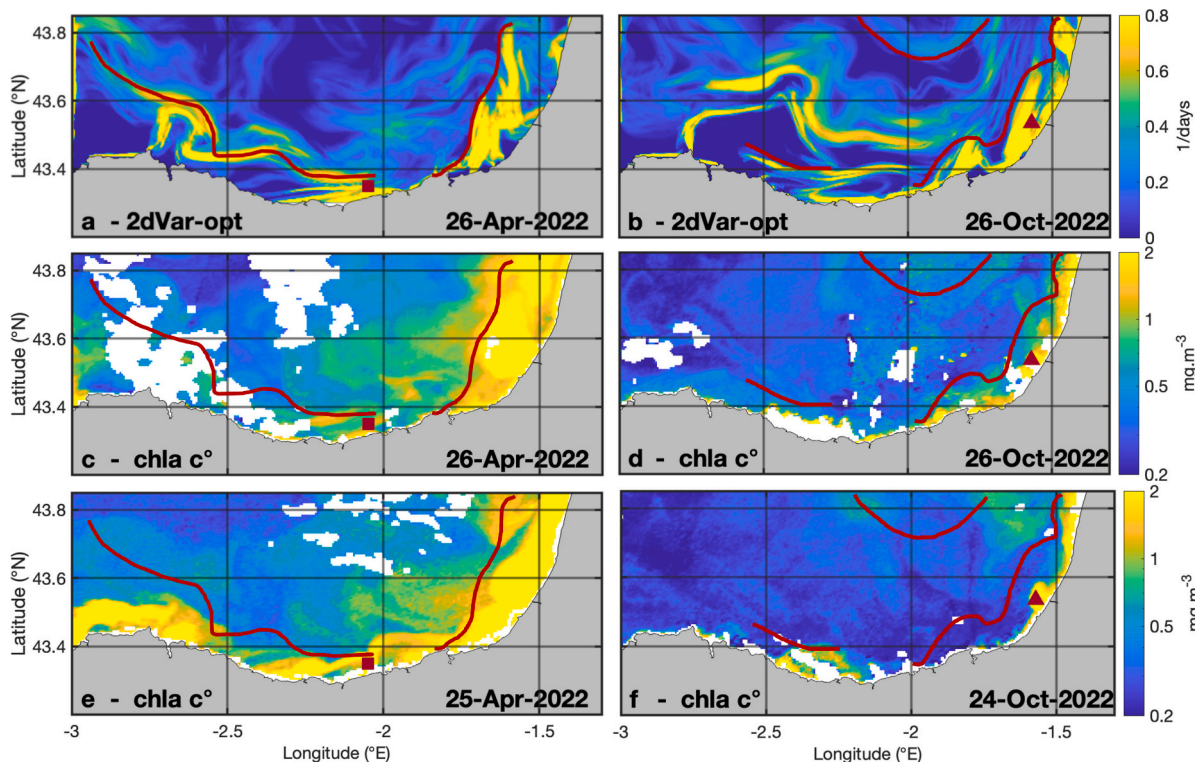


Fig. 8. FSLE maps averaged on April 26, 2022, and on October 26, 2022, computed from the 2dVar-opt field (a, b) respectively. Chl-a concentrations (c, d) derived from the 300 m resolution CMEMS product on (c) April 26, 2022, on (d) October 26, 2022, on (e) April 25, 2022, and on (f) October 24, 2022. Chl-a concentration is plotted in log scale. Specific locations mentioned in the text are shown with red squares and triangles and limits of some FSLE ridgelines are indicated by red lines.



of high Chl-a concentration, namely delineating the offshore limit of river plumes (ex. S1, red square) and in general of Regions of Freshwater Influence (ROFI). The eastern part of the FSLE maps and Chl-a concentration matches well on April 26, 2022. Nevertheless, due to the poor coverage of the image, the FSLE map was compared to another satellite image from the previous day, April 25, 2022. This second satellite image also matches the FSLE ridgelines, especially in the western part of the study area. The meander visible in FSLE ridgelines (Fig. 8a,  $-2.6^{\circ}\text{E}$ ,  $43.6^{\circ}\text{N}$ ) delimits a region of high Chl-a, mostly visible on Fig. 8e.

During S2, comparisons between the FSLE map and Chl-a concentrations are more consistent with the satellite image of October 26, 2022 (Fig. 8d) than with another cloud-free image from two days earlier (Fig. 8f). The northernmost and easternmost FSLE ridgelines delimit regions of high Chl-a, mainly visible in the satellite image of the same day (Fig. 8d), whereas the satellite image of the previous two days shows lower Chl-a concentrations overall, except in the Adour plume (red triangle) and near the Spanish coast ( $-2.4^{\circ}\text{E}$ ,  $43.3^{\circ}\text{N}$ ). The latter area is also delimited by the westernmost FSLE ridgeline, clearly visible in Fig. 8f. The location of the coastal jet (red triangle) shown in the FSLE map (Fig. 8b) is consistent with the high Chl-a concentration values visible in both satellite images (Fig. 8d, f).

## 5. Discussion

Improved identification of coastal CCS is achieved by mapping FSLE using a surface current field optimized by fusing HFR remote sensing data and surface drifter measurements, and completed by using satellite remote sensing data.

The efficiency of data assimilation schemes, based on OI of drifter velocity measurements in circulation models has been assessed in several studies (e.g. Belyaev et al., 2012; Molcard et al., 2003). Other techniques of surface current assimilation in circulation models, such as the Local Ensemble Transform Kalman Filter (Sun and Penny, 2019), 4DVar (Carrier et al., 2014; Muscarella et al., 2015), nudging and sequential methods (Isern-Fontanet et al., 2017), have also proven to be efficient. However, the approach proposed here, which involves merging velocity from HFR and drifter observations, is novel, easy to implement and requires minimal computational cost.

This approach can be applied in other coastal areas where surface velocity fields (from HFR observations or models) and surface drifter measurements are available. However, some adjustments may be necessary, particularly concerning the methods to obtain the covariance matrix used in the OI technique. These methods can be dependent on the targeted coastal area dynamics and the physical processes solved by the surface current available. For instance, in the southeastern BoB, a region with complex dynamics, including the inclusion of tides, inertial oscillation, and slope current was necessary to obtain accurate covariances. The same technique has been applied in the eastern English Channel (Bertin et al., 2024), a region mostly driven by tides, requiring fewer ensemble members to compose the OI covariance matrix due to the high correlation between ensemble members.

The good performance of the 2dVar method to retrieve accurate high-resolution gap-filled HFR velocity fields in the study area has been showcased for the targeted period. However, remote sensing of surface current velocity by HFR has limitations. The quality of HFR data can be affected by different issues related to hardware or software failures and unfavorable environmental conditions, like rough sea states, signal interference or modifications of the directional antenna patterns (Hernández-Carrasco et al., 2018b; Mantovani et al., 2020). This can compromise the availability of the data resulting in reduced range, incomplete spatial coverage or even periods without data. Moreover, the HFR baseline, the area between two radar sites in which the total currents cannot be reconstructed in an accurate way, is a permanent region of large uncertainty. The presence of spatio-temporal gaps prevents the use of the velocity data for Lagrangian analysis. So, the use of gap-filling methods is essential for the studying the transport and dispersion

processes in marine coastal environments. Even after application of gap-filling methods, the resulting velocity fields derived from HFR measurements are not free of errors (Hernández-Carrasco et al., 2018b). So, when an accurate description of small coastal features of coastal circulation is required, the use of an optimization method to reduce the impact of these errors is essential.

In this study, we first demonstrated the high performance of the 2dVar method for reconstructing the current velocity field. We also demonstrated that the OI of 2dVar fields provides the most realistic picture of surface current velocities, and therefore of the location of CCS in the study area. The shape, location and dimensions of these structures vary significantly with time, at seasonal scales, as shown by the comparison between the two surveys, and at shorter scales. It is therefore crucial to correctly characterize the CCS because of their role in the aggregation of marine litter and other materials such as particulate materials or pollutants (Cózar et al., 2021; Ruiz et al., 2020). The Lagrangian assessment of the coastal dynamics enables locating such CCS through the computation of FSLE, since their ridgelines indicate the location of converging Lagrangian trajectories (Hernández-Carrasco et al., 2018a). A notable advantage of FSLE is their ability to unveil oceanic structures at a finer scale than the nominal resolution of the velocity field under analysis, as demonstrated by Hernández-Carrasco et al. (2011). This allows FSLE to capture sub-mesoscale processes more effectively, and the results offer an accurate representation of oceanic transport properties.

FSLE ridgelines, obtained using the optimized velocity fields, were found mainly aligned with the coastline. The coastal CCS depicted scales ranging from 10 to 50 km in the alongshore direction and were much more intense closer to the coast. During S1, the presence of intense or persistent structures was concentrated mostly over the shelf, within a strip of 20 km from the coast, while during S2, CCS were also present in off shelf waters. Large horizontal stirring of real and virtual drifters in the alongshore direction is also supported by the anisotropic character of the absolute dispersion at the end of S1 and during the whole duration of S2. During these periods of time, the zonal component of the absolute dispersion is dominant, in good agreement with CCS orientation parallel to the Spanish Basque coast. Similar situation was observed along the French coast, where FSLE ridgelines were also aligned with the coastline. In both cases, FSLE ridgelines showed a strong alignment with the patterns of high Chl-a concentration and indicated the offshore limit of river plumes, or, more generally, ROFIs.

The comparison performed between FSLE maps and Chl-a spatial distribution is qualitative. First, the remote sensing of Chl-a concentration in coastal waters may still lack reliability and depends on the accuracy of the atmospheric correction, the Chl-a concentration retrieval models and the scale effects of these retrieval models (Chen et al., 2013). Second, to use Chl-a as a passive tracer is especially challenging in coastal areas, where nearshore biogeochemical processes are known to modulate Chl-a concentrations. A possibility to enable quantitative comparisons will be to decompose the Chl-a time series data (with an empirical orthogonal function method or a frequency analysis) into its different components, isolating the more conservative part of the signal. This approach required continuous time series of daily Chl-a concentration (Alvera-Azcárate et al., 2009; Beckers et al., 2006), which were not available during S1 and S2 due to the presence of clouds.

However, satellite ocean color products appear used for investigating the link between coastal CCS and the spatial distribution of particulate material. At larger scales, d'Ovidio et al. (2010, 2015) and Hernández-Carrasco et al. (2018a), demonstrated that FSLE ridgelines were found to act as dynamic barriers, exerting a significant influence on the structuring of the Chl-a field. Ridgelines delimit the Chl-a distribution in space and highlight the crucial role played by small-scale coastal hydrodynamics in shaping the intricate patterns observed in the Chl-a concentrations. During S1 survey, the FSLE ridgeline nearest to the Spanish coast aligned with elevated concentrations of Chl-a, primarily linked to the runoff of the Deba, Urola, and Oria rivers. This convergence

zone also corresponded to the area where drifters' trajectories tended to converge, and where noticeable quantitative concentrations of macro litter and foam were observed in surface waters. The observed CCS and aggregation of macro litter items observed during S1 survey are consistent with previous observations in the area made by (Ruiz et al., 2020). Further research is needed to study the occurrence, loads and fate of litter aggregating in CCS.

The CCS mapping and optimization methods developed here can be applied to operational data for different applications and in situations of vulnerability. For example, in the case of massive waste arrivals in certain areas or oil spill, or for applications such as fishing or search and rescue at sea, where CCS can be used as an indicator of high concentration of fishing resources or areas where search efforts could be intensified. This work can also be combined with satellite imagery data (Rußwurm et al., 2023; Cózar et al., 2024) to provide operational information on marine litter aggregation in coastal areas.

## 6. Conclusions

A method of efficient identification of coastal CCS in the south-eastern BoB is proposed in this study. The findings demonstrated, first, that the surface current field obtained from the 2dVar interpolation of radial HFR velocities provided the best performance compared to the other Eulerian fields, commonly used in the community, like HFR-based OMA product, and modeled fields (IBI). The best performance of 2dVar was observed regarding the Lagrangian error and RMSE obtained from real and virtual trajectories and regarding the identification of CCS. Then, the fusion of 2dVar surface current fields and drifter velocity measurements by means of OI further decreased the Lagrangian error index by 50 % and reduced the RMSE by 30 %. Notably, the optimized 2dVar field (2dVar-opt) was found to provide the most realistic location of CCS. We demonstrated the presence of intense horizontal stirring in the turbulent flow close to the Spanish and French coasts leading to the formation of coastal CCS, aggregating particulate matter. By using the Lagrangian approach, and more specifically FSLE estimation and mapping, we were able to determine the location of the CCS more accurately and we demonstrated that FSLE ridgelines delimit the spatial distribution Chl-a.

These results represent a significant advance in the identification of CCS from remotely sensed surface current velocity and have direct applications for the study of marine litter transport and aggregation. This knowledge is essential in physical oceanography because marine litter serve as tracers for comprehending sub-mesoscale ocean circulation. Inversely, our comprehension of ocean currents plays a significant role in determining floating marine litter pathways. This, in turn, facilitates research and pollution mitigation, which is particularly important in a coastal-dependent society.

The distribution and extension of CCS at subsurface layers and how these structures can affect the transport of floating marine debris of different buoyancy remains unknown. Future research will be focused on studying the vertical dimension of CCS by using 3D model outputs. Another interesting aspect will be to study the impact of different wind drag coefficients in the Lagrangian computations of trajectories and FSLEs in this area. This aspect may be relevant for floating marine debris with sections directly exposed to the wind. Understanding how the wind will influence the CCS spatial distribution and the transport and accumulation of floating marine debris of different nature could provide

significant information for more accurate simulations.

## CRedit authorship contribution statement

**S. Bertin:** Writing – original draft, Visualization, Software, Resources, Project administration, Methodology, Investigation, Formal analysis, Data curation. **A. Rubio:** Writing – review & editing, Validation, Supervision, Resources, Project administration, Methodology, Investigation, Funding acquisition, Conceptualization. **I. Hernández-Carrasco:** Writing – review & editing, Software, Methodology. **L. Solabarrieta:** Writing – review & editing, Software, Methodology. **I. Ruiz:** Writing – review & editing, Methodology, Investigation. **A. Orfila:** Writing – review & editing, Software, Methodology. **A. Sentchev:** Writing – review & editing, Validation, Supervision, Methodology, Investigation, Funding acquisition, Conceptualization.

## Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

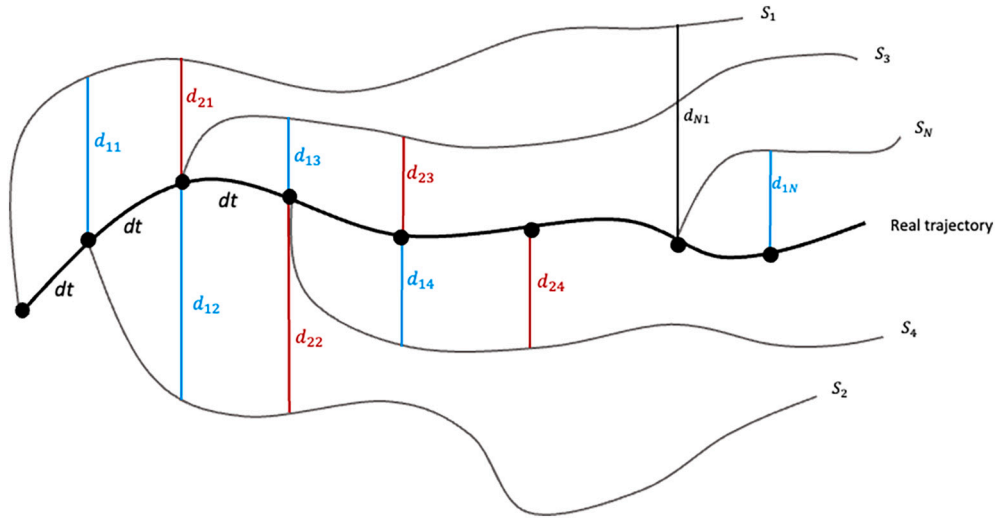
## Data availability

Data will be made available on request.

## Acknowledgements

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## Appendix A. Computation of the Lagrangian error



**Fig. A.** Schematic of the computation of the Lagrangian error for one drifter. The real drifter trajectory corresponds to the black line, and the  $N$  virtual trajectories deployed every  $dt$  ( $dt = 1$  h) corresponds to the grey lines.

Based on Fig. A, the steps of computation of the Lagrangian error are the following:

$$D_1 = D(1dt) = \frac{d_{11} + d_{12} + d_{13} + \dots + d_{1N}}{N}$$

$$D_2 = D(2dt) = \frac{d_{21} + d_{22} + d_{23} + \dots + d_{2N-1}}{N-1}$$

$$D_3 = D(3dt) = \frac{d_{31} + d_{32} + d_{33} + \dots + d_{2N-2}}{N-2}$$

...

$$D_{N-1} = D((N-1)dt) = \frac{d_{(N-1)1} + d_{(N-1)2}}{2}$$

$$D_N = D(Ndt) = d_{N1}$$

## Appendix B. Sensitivity test to the number of clusters in K-Means and to the number of ensemble members in OI covariance matrix

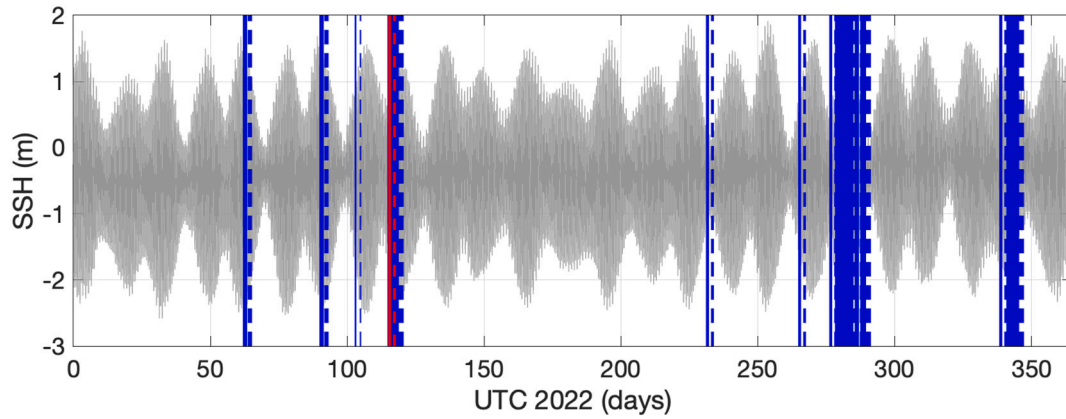
As described in Chapter 3.6, the non-linear K-Means clustering algorithm was utilized for two main objectives. These objectives included identifying dominant surface current groups during each survey and extracting ensemble members that shared similar dominant groups as observed during each survey period, accounting for 90 % of the variability. In order to identify the optimal number of clusters for each survey, we conducted experiments involving varying cluster counts within the K-Means method (refer to Table B1).

It was decided that 8 and 6 clusters of one hour each were sufficient to represent around 40 h of survey time (Table B1, grey shaded lines), and the sensitivity test was continued by taking 35 and 12 clusters in the K-Means method. Simultaneously to the sensitivity test for the number of clusters in the K-Means method, we also conducted a sensitivity test for the number of ensemble members comprising the OI covariance matrix. To construct this covariance matrix, we first identified the predominant groups observed during each survey. Subsequently, we extracted ensemble members that exhibited the same predominant groups in a similar order of magnitude from the year-long 2dVar field (covering January 2022 to December 2022), which included data from both surveys. It is important to notice that every chosen ensemble member aligned with the same tidal stage as the survey in question. To prevent any tidal influence, we exclusively opted for ensemble members synchronized with the tidal oscillations, even though tides contribute to a relatively small portion of the total kinetic energy in the study area (approximately 10 to 15 %, as indicated by Solabarrieta et al., 2014).

**Table B1**

Number of representative clusters found during S1 and S2 using different number of K-Means clusters, with 90 % of variability explained. Grey shaded lines correspond to the number of K-Means clusters we decided to further test.

Variability explained (%)	Number of clusters	Number of clusters in S1	Number of clusters in S2
90	50	10	9
	45	9	10
	35	8	8
	25	6	7
	12	6	5



**Fig. B1.** An example of 80 ensemble members (in blue) extracted from the 1-year long 2dVar field. S1 period is shown in red. Sea surface height (SSH) from IBI model run is given in grey.

Subsequently, we calculated the RMSE (Table B2) and the Lagrangian errors (Table B3) for both the initial 2dVar field and the optimized 2dVar field, with the covariance matrix comprising 20, 40, 60, and 80 ensemble members. Notably, Tables B2 and B3 underscore that the outcomes of the OI technique are notably influenced by the number of ensemble members used in the OI covariance matrix, rather than the number of clusters employed in the K-Means clustering method. Across 35 and 12 K-Means clusters, transitioning from 20 to 80 ensemble members in the OI covariance matrix yielded a 12 % reduction in RMSE during S1 and a 17 % reduction during S2 (as indicated in Table B2). Conversely, opting for 35 K-Means clusters instead of 12 only marginally decreased the Lagrangian error (averaged over the number of ensemble members in the OI covariance matrix) by 1.6 % during S1 and 4.6 % during S2. Regarding the Lagrangian error, averaged over 35 and 12 K-Means clusters, increasing the number of ensemble members in the OI covariance matrix from 20 to 80 resulted in a 10 % decrease in error during S1 and a substantial 40 % reduction during S2 (as detailed in Table B3). In contrast, choosing 35 K-Means clusters over 12 led to a modest reduction of the Lagrangian error index by 1 % during S1 and 9 % during S2. It is not surprising that the OI outcomes were particularly responsive to the number of ensemble members comprising the covariance matrix, given the intricate dynamics prevalent in the southeastern BoB. This region exhibits a complex interplay of various influences, including freshwater inputs, swell, inertial oscillations, and tides.

**Table B2**

Mean RMSE computed from Eq. 3 for both surveys considering 35 and 12 K-Means clusters and a number of ensemble members varying from 20 to 80. The mean error was computed from the initial 2dVar field (column 3) and the optimized 2dVar field (column 5). The standard deviation of the mean error of the optimized field is given at the 5th column.

	Number of clusters K-Means	Number of ensemble members OI	$RMSE_r$	$RMSE_{opt}$
S1	35	20	0.94	0.80
		40		0.77
		60		0.75
		80		0.73
	12	20		0.81
		40		0.79
		60		0.72
		80		0.72
S2	35	20	0.43	0.32
		40		0.30
		60		0.29
		80		0.28
	12	20		0.34
		40		0.31
		60		0.29
		80		0.28

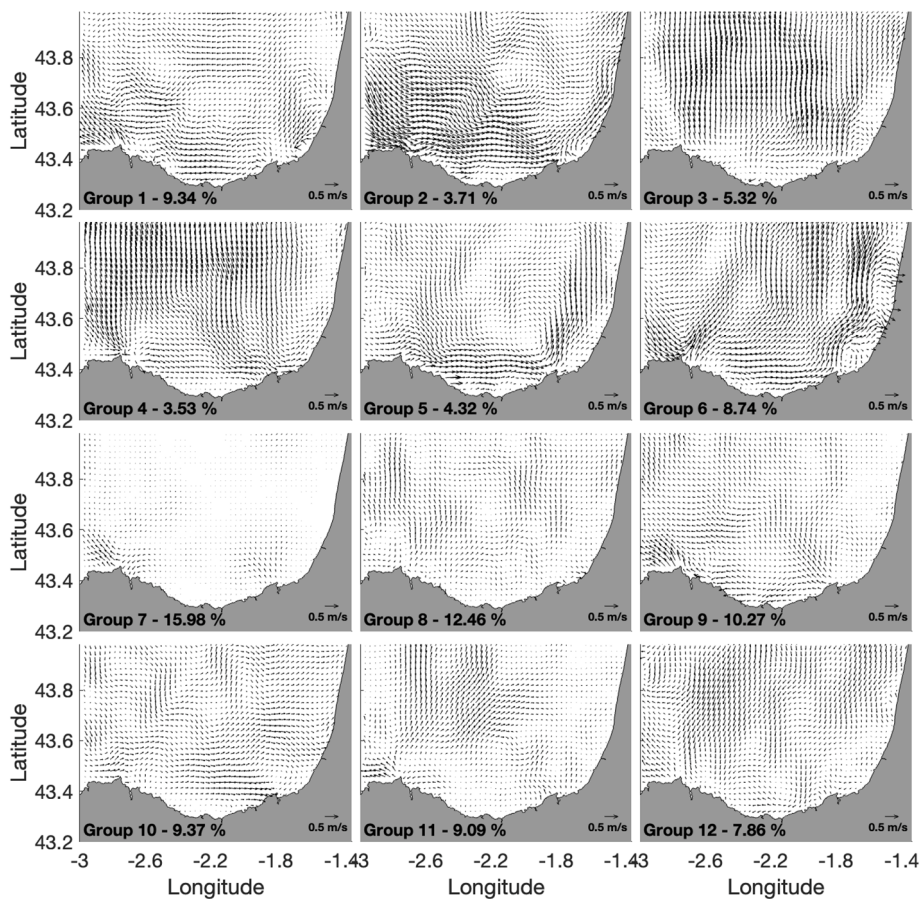
Therefore, in the sensitivity test regarding the number of K-Means clusters, we opted for 12 clusters, as it yields nearly identical results to using 35 clusters, with the added advantages of reducing computational costs and aligning with the findings of (Solabarrieta et al., 2015), who demonstrated that 12 K-Means clusters were adequate for assessing variability in the southeastern BoB.

Hence, regarding the sensitivity test to the number of K-Means clusters, we chose to employ 12 as it is giving approximately the same results as 35 for practical reasons: it reduced the computational cost, and it was proven that 12 K-Means clusters were sufficient to assess for the variability in the southeastern BoB. The 12 clusters are shown in Fig. B2. In the sensitivity analysis regarding the number of ensemble members forming the OI covariance matrix, the study's results were derived using 80 ensemble members, as it consistently produced the most favorable outcomes in both survey periods.

**Table B3**

Mean Lagrangian error index  $L$  computed for both surveys considering 35 and 12 K-Means clusters and a number of ensemble members varying from 20 to 80.

	Number of clusters K-Means	Number of OI ensemble members	$L_r$	$L_{opt}$
S1	35	20	0.5	0.37
		40		0.36
		60		0.35
		80		0.32
	12	20		0.37
		40		0.35
		60		0.36
		80		0.32
S2	35	20	0.2	0.13
		40		0.10
		60		0.09
		80		0.10
	12	20		0.15
		40		0.10
		60		0.10
		80		0.09



**Fig. B2.**  $4 \times 3$  lattice of the K-Means algorithm applied to HFR surface currents from January 2020 to December 2022. The percentage of variability explained by each group is given in the title.

Appendix C. Lagrangian metrics of the different surface current fields

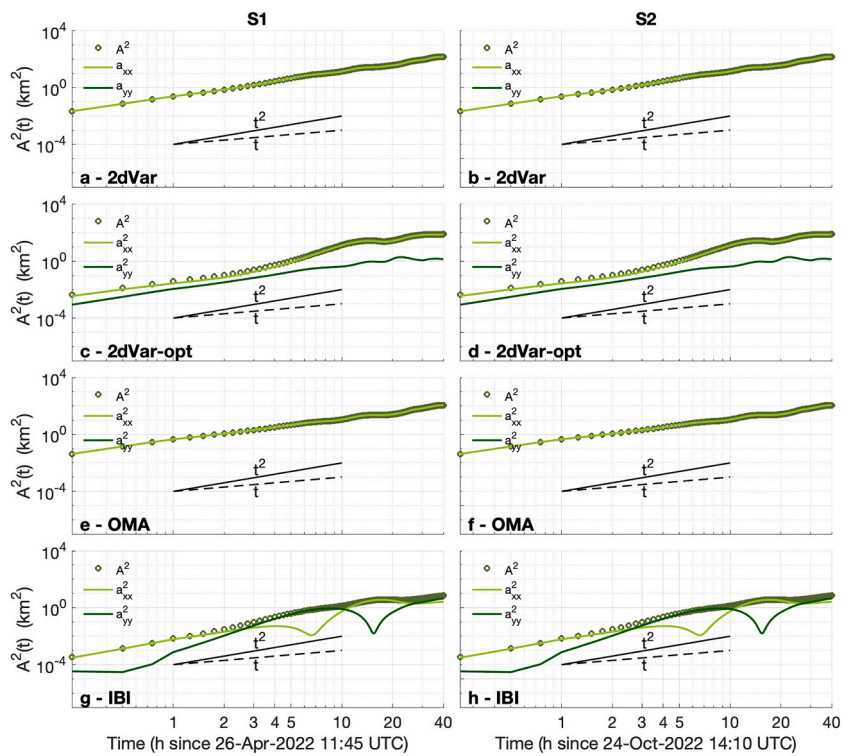


Fig. C1. Absolute dispersion along x and y computed from Eq. (4) for the modeled trajectories using 2dVar, 2dVar optimized, OMA and IBI surface current fields during S1 (left column) and S2 (right column). The ballistic ( $t^2$ ) and diffusive ( $t$ ) laws are shown by the black line and dashed black line respectively.

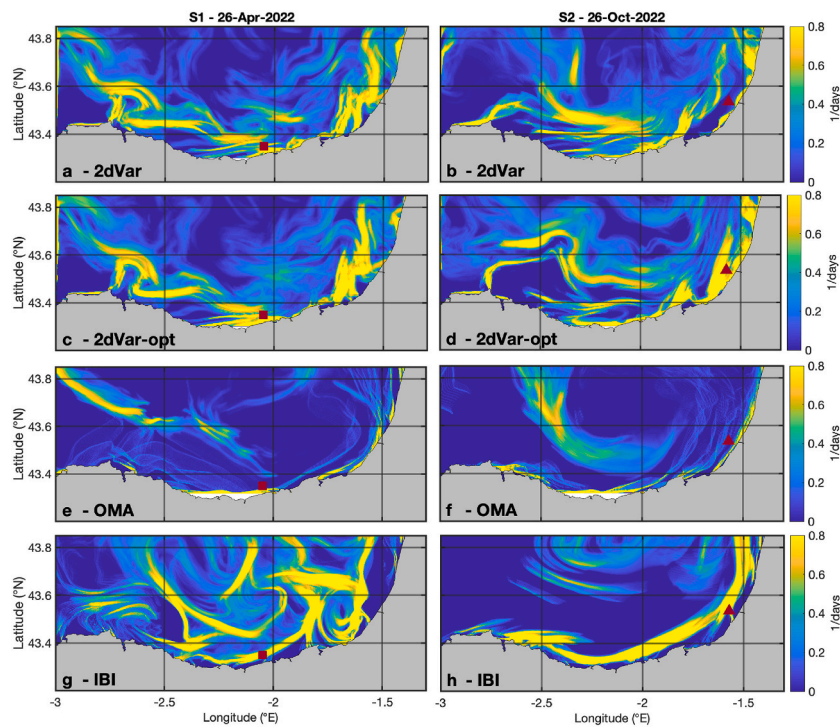


Fig. C2. FSLE maps averaged on the 26th of April 2022 (left column) and on the 26th of October 2022 (right column) computed from the initial 2dVar field (a, b), the optimized 2dVar field (c, d), OMA field (e, f) and IBI field (g, h). Specific locations mentioned in the text are shown with red square and triangle symbols.

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## Chapitre 5

# Effet du vent sur les structures de convergence

### 5.1 Introduction

#### *Motivations de l'article*

Dans la continuité de l'étude précédente dans le sud-est du golfe de Gascogne, portant sur l'identification des structures de convergence des courants, cette étude se focalise particulièrement sur le transport de matière sous l'effet conjugué des courants et du vent afin de cibler les déchets marins flottants. Ainsi, l'objectif était d'étudier l'impact du vent sur les structures de convergence des courants susceptible d'accumuler les déchets marins flottants. Seulement la campagne B1, incluant le lâcher de bouées dérivantes avec et sans ancre flottante, a été considérée pour cette étude.

Les diagnostics lagrangiens utilisés dans les études précédentes se sont révélés efficaces pour cartographier la distribution spatiale des structures de convergence, se basant sur des champs de courants de surface fiables. Ils m'ont permis d'identifier les structures lagrangiennes cohérentes agissant comme des barrières de transport convergentes où une plus grande concentration de déchets marins flottants est susceptible de se produire. Ces structures de convergence affectent la matière particulaire soumise seulement aux courants de surface. Cependant, étant donné qu'une partie des déchets marins flotte en surface, son mouvement peut également être affecté par le vent. Ainsi, l'analyse des métriques lagrangiennes doit tenir compte de ce transport modifié par le vent.

En effet, la dérive induite par le vent sur les matériaux flottant à la surface libre de l'océan joue un rôle crucial dans le transport en surface de matériaux biologiques et de contaminants. De nombreuses études identifient les structures cohérentes lagrangiennes pour étudier des scénarios de transport à la surface des océans tels que les marées noires, les déchets marins ou les traceurs biogéochimique. Cependant, l'influence du vent sur ces structures organisationnelles influençant le transport de matière particulaire n'a été que très peu documenté.

Tenir compte du vent permet aussi de différencier les déchets marins flottants en fonction de leur prise au vent. Le courant induit par la prise au vent, ou windage en anglais, peut être estimé comme un pourcentage du vent à 10 m de hauteur. Par exemple, un conteneur à la dérive aurait une prise au vent de 1.4%, un jerrican 0.8%, un fragment de plastique 2%, ou encore une bouteille en plastique vide et bouchonnée 4%. Ainsi, il est possible de prédire le comportement

de différent type de déchet marin flottant sous réserve de considérer la prise au vent adéquate lors des calculs.

### *Méthodes d'analyse*

L'étude précédente, présentée au chapitre 4, a démontré que les champs de courants les plus fiables étaient ceux provenant des données radar HF issus de la méthode d'interpolation variationnelle bi-dimensionnelle 2dVar. Basée sur l'utilisation de ces courants de surface, et l'ajout d'une vitesse induite par le vent, cette étude a pour objectif de caractériser l'impact du vent sur le transport de particules virtuelles, assimilables à des déchets marins flottants.

Dans un premier temps, l'algorithme de clustering SOM a été utilisé afin de distinguer cinq régions de la zone d'étude présentant une dynamique distincte, en se basant sur trois ans de champs de courant de surface radar (produit 2dVar). A partir de ces cinq clusters, des particules virtuelles, assimilables à des déchets marins flottants, ont été déployées et advectées par les courants de surface (avec et sans influence du vent), afin de calculer les taux d'échouage et de sortie du domaine d'étude, ainsi que les zones de passage (calcul des PDF présenté au chapitre 2.5) favorisées par les structures de convergence.

Ensuite, la dispersion absolue, les FSLE et la FDL D ont été calculées avec et sans vent. Les PDF et la FDL D ont été superposées aux lignes de crêtes des FSLE afin de visualiser l'impact du vent sur les structures de convergence.

Le coeur de l'article présente des résultats obtenus à partir des vents réalistes issus du modèle WRF de l'agence météorologique de Galice. Cependant, quatre autres tests ont été réalisés avec des vents théoriques typiques du sud-est du golfe de Gascogne : vent de nord, nord-est, nord-ouest et sud-ouest. Les résultats de ces tests sont présentés en annexe.

### *Résultats*

Les résultats de cette dernière étude sont multiples et significatifs. En effet, nous avons démontré que le vent exerce une influence significative sur la distribution, la provenance et les trajectoires d'accumulation des déchets marins flottants dans la zone d'étude. Une dispersion hautement anisotrope a été observée, tant pour les bouées équipées d'ancre flottante que pour celles sans ancre, avec une dispersion dix fois plus importante dans la direction principale du courant. De plus, le taux de dispersion est 50 fois plus élevé pour les bouées de surface plus exposées au vent (sans ancre flottante).

Par ailleurs, les indicateurs FSLE et FDL D se sont avérés particulièrement efficaces pour localiser et suivre l'évolution des structures de convergence des courants marins, ainsi que pour analyser leur impact sur les trajectoires de transport des déchets marins flottants.

Lorsque le windage est inférieur à 2-3%, les structures d'agrégation des déchets marins flottants sont principalement influencées par l'effet combiné des courants océaniques et des vents dominants, entraînant le déplacement de ces structures en accord avec ces forces.

En revanche, lorsque la dérive éolienne dépasse 3%, une augmentation marquée des taux d'échouage et de dispersion des déchets marins flottants est observée, accompagnée d'une réduction du nombre de structures d'agrégation distinctes. Cette diminution est due à un windage

plus élevée, qui favorise à la fois l'échouage et la dispersion, tout en entraînant la fusion des structures d'agrégation.

Malgré les variations de direction du vent, les zones de passage intense des déchets marins flottants tendent à se situer dans des zones constantes. L'origine des déchets au sein de ces zones d'agrégation peut varier en fonction du niveau de windage, suggérant une interaction complexe entre le courant induit par le vent et le transport des déchets marins flottants.

En intégrant cette méthode, les pratiques de gestion côtière pourraient bénéficier d'une amélioration considérable. Cette approche fournit des perspectives plus réalistes et détaillées sur les structures de transport sous-jacentes déterminant le transport des déchets marins flottants, permettant ainsi de mieux comprendre les dynamiques complexes en jeu. Une vision plus précise des voies de transport et des zones d'accumulation permettrait de formuler des stratégies plus efficaces pour atténuer les impacts environnementaux, optimiser les interventions de dépollution, et mieux protéger les écosystèmes côtiers et estuariens.

Différents aspects de l'étude, brièvement présentés ci-dessus, ont été développés plus en détail dans un article de recherche intitulé : « *The impact of windage on the transport and accumulation of marine litter : a Lagrangian analysis in the Bay of Biscay* » soumis dans la revue *Journal of Marine Systems*, Special Issue de la conférence ISOBAYs 2024.

**5.2 Article : *The impact of windage on the transport and accumulation of marine litter : a Lagrangian analysis in the Bay of Biscay***

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# The impact of windage on the transport and accumulation of marine litter: a Lagrangian analysis in the Bay of Biscay

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## Abstract

The southeastern Bay of Biscay (SE-BoB) is known as an accumulation zone for marine litter. Coastal Current Convergence Structures (CCS), which are linked to vertical movements at the edges of river plumes, estuarine fronts, or other frontal processes, have been observed in this area to act as aggregators of floating marine litter (FML). Observed at the ocean surface, CCS often form narrow filament-like streaks with variable loads of FML. Understanding the role of CCS in the transport and aggregation of FML is crucial for quantifying and mitigating pollution. In the Lagrangian framework, the use of Finite Size Lyapunov Exponents (FSLE) and the Finite Domain Lagrangian Divergence (FDLD) has proven in a previous work to be a good approach to study and locate CCSs in the area by using the velocity field data from the EuskOOS High-Frequency Radar. Backwards-in-time FSLE and FDLD unveil Lagrangian Coherent Structures (LCS), which act as convergent transport barriers, a proxy of CCS where larger concentration of FML of different buoyancy is likely to occur. Since a fraction of FML is buoyant and its motion can be driven also by windage, additional computations are conducted considering the effect of the wind drag force on FML trajectories and Lagrangian quantities. Escaping and beaching rate of particle clusters, delineating areas of different dynamic regimes, are also assessed considering different windages. Our results suggest that the wind has a significant influence on the FML distribution and accumulation paths in the study area, causing the aggregation of materials in areas not coinciding with CCS driven only by currents. Different locations of barriers to transport are obtained, depending on the wind regime and windage considered. Below 2-3% windage, aggregation structures are shifted in the direction of currents and winds. Above 3% windage, beaching and escaping ratios are increasing and aggregation structures tend to merge.

## Key words

Coastal Current Convergence Structures, Windage, floating marine litter, Lagrangian diagnostics, escape rates, beaching

## Highlights

- Wind has a significant influence on the buoyant FML distribution, provenance and accumulation paths.
- Highly anisotropic dispersion is observed in the Southeast Bay of Biscay by means of drifters and high-frequency radar data, with ten times larger dispersion along-shore, i.e. along the main flow direction.

- The dispersion rate for surface drifters more exposed to the wind (undrogued drifters) is 50 times larger, although dispersion regimes are similar for drogued and undrogued drifters
- Below 2-3% windage, aggregation structures are shifted in the direction of currents and winds. Above 3% of windage, beaching and escaping ratios are increasing and fewer aggregation structures are observed, as they merge.

## 1. Introduction

Coastal regions, defined as the transitional areas between land and sea, are subjected to numerous anthropogenic pressures, mainly as the result of industrial activities, fishing or tourism. These regions are complex environments whose dynamics can evolve over a wide range of spatio-temporal scales. Different physical mechanisms such as fronts and small eddies control the transport of physical, chemical and biological substances (Nencioli et al., 2011; Zhang, 2017). At the sub-mesoscale, with characteristic spatial scales less than 10 km, various physical processes drive current variability and generate fronts and eddy-like structures. These sub-mesoscale processes are known to be very important locally for the transport of passive quantities (McWilliams, 2016; Poje et al., 2014). Several mechanisms are responsible for the formation of kilometer-scale fronts, such as vertical movements at the edges of river plumes, estuarine fronts, or other frontal processes (Ayouché et al., 2020; Mann and Lazier, 2005). These small-scale fronts are associated with the intense convergence of horizontal flow induced by sub-mesoscale dynamics, and can impact the distribution of buoyant material (D'Asaro et al., 2018; Hernández-Carrasco et al., 2018; Huntley et al., 2015). Sub-mesoscale surface ocean structures present a challenge to both numerical models and observations due to their chaotic nature and because high resolution is required to properly solve the related variability (Capó et al., 2021).

In the Lagrangian framework, quantities such as the Finite Size Lyapunov Exponent (FSLE) and Finite Domain Lagrangian Divergence (FDLD) are very effective to unveil Lagrangian Coherent Structures (LCS) and to provide information on transport patterns associated with the flow, such as coherent convergence and divergence structures at the ocean surface (Haller, 2015; Hernández-Carrasco et al., 2018; Hernández-Carrasco and Orfila, 2018). Attracting LCS often coincide with Current Convergent Structures (CCS), where the accumulation of floating marine litter (FML) is likely to occur (Hernández-Carrasco et al., 2018; Huntley et al., 2015). However, a fraction of FML is buoyant and its motion can be driven also by windage, the additional direct wind-induced drift of material floating at the free surface of the ocean (also called wind-drift factor or leeway). Thus, relevant Lagrangian analysis metrics must account for this modified transport (Allshouse et al., 2017). In coastal ocean regions, the Lagrangian properties of the flow have been assessed using velocity data from models (Bertin et al., 2024b; de Mello et al., 2023; Kumar et al., 2023), from satellite altimetry (Siegelman et al., 2020), as well as from High-Frequency Radars (HFR) (Bertin et al., 2024a; Hernández-Lasheras et al., 2024; Tran et al., 2022; Veatch et al., 2022). The use of HFR technology has the potential to provide observational information with the necessary resolution in both space and time to unveil small scale structures such as CCS (Bertin et al., 2024a).

The southeastern Bay of Biscay (hereinafter SE-BoB) has been identified by global and regional models as an accumulation zone for marine litter (Pereiro et al., 2019; Rodríguez-Díaz et al., 2020). Previous studies on macrolitter (Ruiz et al., 2020) and microlitter (Davila et al., 2021), as well as biota, have shed a light on the quantities and behaviors of litter at the water surface in the SE-BoB, reinforcing the hypothesis of litter accumulation in the area. These recent studies also underscored the significant

seasonal variability of this accumulation (Declerck et al., 2019; Pereiro et al., 2019; Ruiz et al., 2022). In a recent study in the area, Bertin et al. (2024a) have showcased the presence of CCS with FML loads, mostly associated with river plume edges, i.e. aligned with remotely sensed high concentrations of Chlorophyll-a, and in agreement with the in-situ observations made by Ruiz et al. (2020). Understanding the formation and evolution of the CCS and their role in the transport of FML is essential to better quantify and to help mitigate marine litter pollution. Bertin et al. (2024a) demonstrated that estimating FSLE from velocity data provided by HFR data allowed a realistic characterization of CCS, their location and evolution. The authors also reported the important role of the CCS in the accumulation of marine litter observed in windrows distributed alongshore, identified by filament-like structures of high FSLE values. However, the effect of windage on the computation of these CCS and on the precise detection of floating particle aggregation zones still remains unclear.

This paper focuses on the impact of wind on the transport of FML under different windage and on how wind affects the distribution and evolution of CCS driving the accumulation of FML in coastal waters of the SE-BoB. To do so, we use a combination of surface currents from drifters and HFR as well as realistic and theoretical modeled winds. Several simulations are performed taking into account a range of windage coefficients, representative of different types of FML (for example, a drifting container is subject to a windage of 1.4%, while a capped plastic bottle is subject to a windage of 4% (Breivik et al., 2011; Park and Seo, 2021)) and also accounting on FML from different provenance. We estimate beaching and escaping ratios in the study area, aligned with absolute dispersion, FSLE and FDLT to showcase the impact of windage on the transport, fate and aggregation of FML.

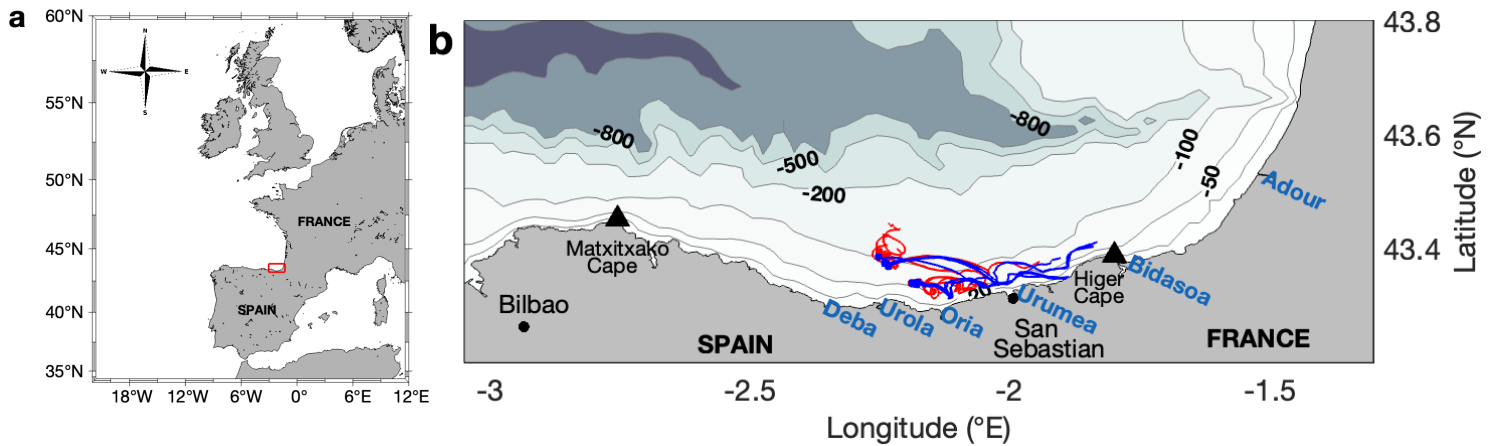
The core of the article presents results obtained using realistic winds from an operational meteorological system. Additionally, four academic configurations, carried out using theoretical winds typical of the SE-BoB in the period of the observations, provide a detailed insight to the effect of wind variability on the structuration and transport of FML.

## **2. The southeastern Bay of Biscay**

The main current regime is along-slope and depicts a strong seasonal variability (Charria et al., 2013; Rubio et al., 2019; Solabarrieta et al., 2014; Vandermeirsch et al., 2012), dominated by the Iberian Poleward slope Current (hereinafter IPC). Coastal circulation in the SE-BoB shows a complex pattern. Along the slope, the IPC flows eastward off the Spanish coast and northward off the French coast. In autumn and winter, it has a relatively intense northward flow (2-5 cm/s at the surface), carrying warm waters from the southeastern corner of the Gulf northwards, resulting in a warm tongue stretching from the Basque country to Brittany. In spring and summer, the IPC weakens and becomes more variable, moving mainly eastwards (Charria et al., 2013; Rubio et al., 2019; Solabarrieta et al., 2014; Vandermeirsch et al., 2012). The interaction of the IPC with the abrupt bathymetry often generates eddies (e.g. Pingree and Le Cann, 1992).

On the continental shelf, circulation is mainly determined by wind variability (Fontán et al., 2009; González et al., 2004; Solabarrieta et al., 2015), and is strongly influenced by the orientation of the coastline. Thus, the effect of wind adds complexity to this seasonality (Rubio et al., 2018; Solabarrieta et al., 2015), with prevailing southwesterlies in autumn and winter (reinforcing the winter poleward flow), and northeasterlies in spring and summer (contributing to a weaker and highly variable equatorward flow). Transport and retention conditions at the surface vary at interannual, seasonal and even daily scales, influenced by the interplay of the aforementioned processes (Rubio et al., 2020).

The most important freshwater influence on the SE-BoB shelf comes from the Adour river in France (Figure 1b), with a mean annual discharge of  $300 \text{ m}^3 \text{ s}^{-1}$  (Valencia et al., 2004). On the Spanish coast, the rivers have much lower discharges and are torrential in character, with very short time lags between rainfall and the resulting river flow. The main rivers draining the Spanish coast are the Nervión, Oria and Bidassoa (Figure 1b), with average annual flows of 20 to  $30 \text{ m}^3 \text{ s}^{-1}$  (Ferrer et al., 2009). The highest discharges are, in both cases, observed mainly in spring and autumn, while the lowest river flows occur in late summer.



**Figure 1.** Map of the study area: the bathymetry is shown by blue shading and the black triangles display the location of the two HFR stations at Matxitxako (west) and Higer (east) Capes. The location of the major river mouths is given in blue. Trajectories of surface drifters with and without drogue are shown in red and blue, respectively. The left panel shows the location of the study area as the red rectangle in the BoB Cantabric coast.

### 3. Materials and methods

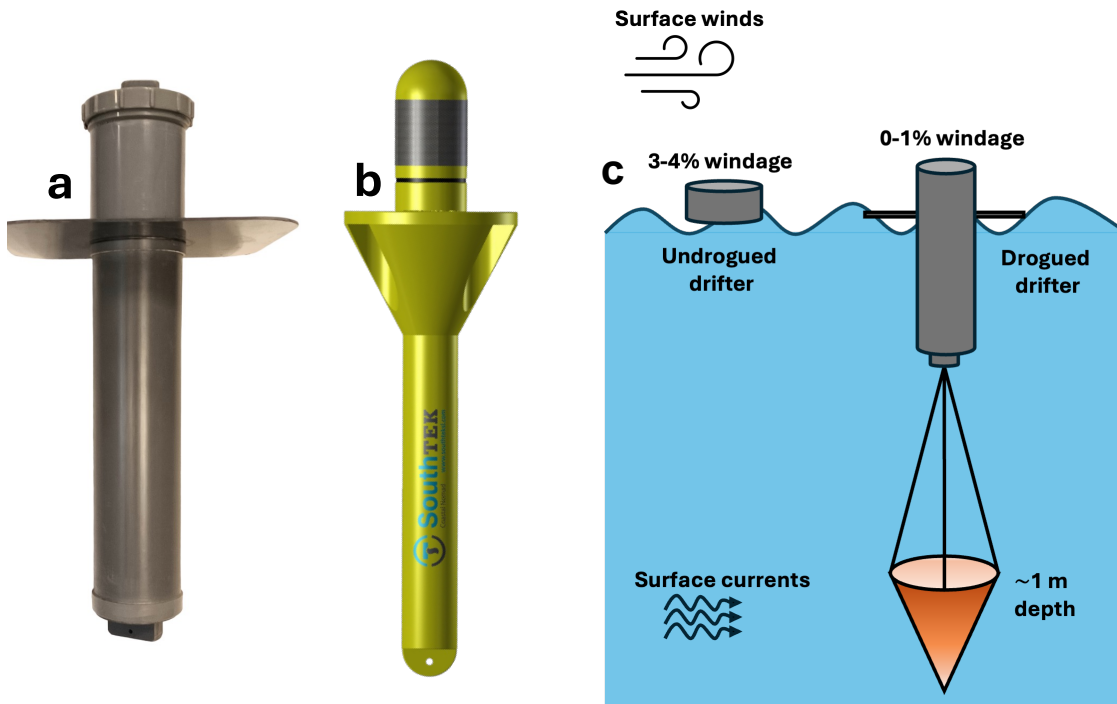
#### 3.1. Lagrangian drifter surface current data

An in-situ experiment was conducted in the coastal waters of the SE-BoB during spring 2024, in an area where occurrence of coastal CCS was previously reported (BOBLIT0.1, see Bertin et al 2024b, for more detail). From 26<sup>th</sup> April 11:45 UTC to 28<sup>th</sup> April 07:00 UTC, 20 surface drogued and undrogued drifters were deployed in pairs (pairing one drogued and one undrogued drifter) in two different clusters (centered at  $43.4^\circ\text{N}$ ,  $-2.25^\circ\text{E}$  and  $43.35^\circ\text{N}$ ,  $-2.3^\circ\text{E}$ , see Figure 1b, red and blue dots for locations). In both clusters, the drifter pairs were deployed in areas within visible CCS and where qualitative aggregation of foam and macrolitter was observed (Bertin et al., 2024a). The mean separation distance between drifters was 1.7 and 0.6 km in the first and second cluster, respectively.

Regarding drogued drifters, three types of drifters were used: the coastal Nomad surface buoys, cylinder-shaped manufactured by SouthTek (<https://www.southteksl.com/>); ODI surface buoys formerly manufactured by Albatros Marine Technologies and a set of cylinder-shaped home-made drifters (whose design was similar to the coastal Nomad drifters (see Figure 2a, b)). All drifters were equipped with an anchor of 0.5 m in length positioned in the water column between 0.8 and 1.3 m depth, allowing them to drift with surface currents and minimize the direct effect of wind drag (Figure 2c).



Undrogued drifters were home-made, featuring a plastic container of 9 cm in height and 9.5 cm in diameter, weighing approximately 200 g (see Ruiz et al. (2022) for a detailed description and Figure 2c for the schematic).



**Figure 2.** Schematic of the drogued (a, b, c) and undrogued (c) drifters used during the survey and the windage to account for the impact of surface wind.

Observed surface current velocities were estimated from drogued and undrogued drifter trajectories with a timestep of 15 min, nominal period of drifter positioning via GPS. Potential differences in the behavior between different drogued drifter designs were neglected in the computation of Lagrangian diagnostics taking into account the short-term window considered (40 hours) and that all drifters were equipped with the same drogues.

### 3.2. HFR surface current data

Hourly surface velocity current fields were obtained from a HFR system, consisting of two CODAR Seasonde stations situated along the Spanish Basque Country coast in the SE-BoB. EuskOOS HFR network is part of euskoos observatory (<https://doi.org/10.57762/T4WH-DQ48>) and JERICO (<https://www.jerico-ri.eu/>) and it is operated following JERICO standards and recommendations (Mantovani et al., 2020; Rubio et al., 2018; Solabarrieta et al., 2016). The two individual radar stations, operating at a frequency of 4.86 MHz, enable velocity measurements within the range of  $\sim 150$  km with 5 km along-beam resolution, and  $5^\circ$  angular resolution. Velocity data are obtained at 1 h time step and distributed as part of Copernicus Global Ocean-in-situ near real-and delayed-time surface ocean currents products (<http://doi.org/10.48670/moi-00041> & <http://doi.org/10.17882/86236>).

Based on the measured radial velocities, surface current fields were generated using the two-dimensional variational interpolation method (2dVar) (Yaremchuk and Sentchev, 2009) covering an area between  $-3^\circ\text{E}$  to  $-1.3^\circ\text{E}$  and between  $43.3^\circ\text{N}$  to  $44^\circ\text{N}$ . The 2dVar technique enables to obtain hourly currents with

2.5 km spatial resolution. 2dVar is a non-local and kinematically constrained interpolation method which creates surface current velocity maps by using a combination of all measured velocities, within a given grid, reconstructing the velocity vector in one location. This interpolation technique has been used in a previous study in the SE-BoB and proved its reliability (see Bertin et al. (2024a) for detailed description).

### 3.3. Winds and windage computation

Mean wind velocity was obtained from the Weather Research and Forecasting model (WRF) provided by the Galician Meteorological Agency (Meteo Galicia). Hourly zonal and meridional components of wind velocity at a height of 10 m were used, featuring a spatial resolution of 4 km. Model details are given for completeness in Skamarock et al. (2019).

The force exerted by the wind on the emerging part of an object is known as windage and can be estimated by a simplified model, following Yoon et al. (2009). In brief, windage is computed at the surface as a fraction of the wind velocity at 10 m height:

$$U_{windage} = C_x U_{10} \quad (1)$$

$$U_T = U_{HFR} + U_{windage} \quad (2)$$

where  $U_{windage}$  is the velocity of current due to the winds,  $U_{10}$  is the wind speed at 10 m height,  $U_{HFR}$  the current velocity measured by HFR, and  $U_T$  is the total velocity. The wind drift factor  $C_x$ , depends on the geometry and the percentage of the emerged surface of the object (Sainte-Rose et al., 2016). This assumption considers a flat ocean surface, a purely downwind windage (symmetric object, no crosswind component) and wind velocities higher than both windage currents and oceanic currents.

### 3.4. SOM clustering

Self-Organizing Map (SOM) clustering was used to obtain a map of different dynamical regions in the study area in order to seed particles in clusters for the analysis of escape rates and beaching. The SOM technique consists in an unsupervised learning neural network especially suited to extract nonlinear patterns in large datasets (Kohonen, 1998). This classification tool is used to map high-dimensional input data onto a two-dimensional space, while preserving the topological relationships between the input data (Liu et al., 2006). The input data are clustered in units, called neural units or neurons. The number of units is the same as the size of the SOM array (or lattice), where each unit has a weight vector that has the number of components equal to the dimension of the input sample data.

SOM learning process involves inputting velocity fields into a neural network, which undergoes modification through an iterative procedure. Each neuron is characterized by a weight vector, matching the dimensions of the input sample data. During each iteration, the neuron with the weight vector closest to the input data vector, identified by minimum Euclidean distance, is adjusted along with its neighboring neurons according to a specified neighborhood relationship defined by a mathematical function. By the end of the training process, SOM approximates the probability density function of the input data, assigning each neuron a reference pattern.

SOM technique can be applied both in the spatial (e.g. Orfila et al., 2021) and temporal domains (e.g. Morales Márquez et al., 2021), according to the objectives of the study to be conducted. In this work, the SOM technique has been applied in the temporal domain in order to classify regions of the SE-BoB according to the temporal variability of surface currents measured by HFR. The input dataset consisted of a time series of three years of 2dVar HFR surface current velocities, which was normalized before starting the learning process and which included the land mask (with constant values of 999). The number of neurons was fixed to six, (in a topology of 2 x 3) to distinguish six regions with distinct dynamical behaviors (one of the six being the land mask). We used a rectangular map lattice and opted for a linear mode for the initialization, a batch algorithm for the training process, and an “gaussian” type of a neighborhood. The output of the analysis was a map of different sub-regions characterized with a particular temporal variability. The temporal patterns associated with each coherent region are presented in Appendix A.

### 3.5. Virtual particles advection and estimation of Lagrangian error

The potential impact of CCS and windage on the spatial distribution of FML was assessed by analyzing the trajectories of virtual particles advected in the 2dVar HFR velocity fields. Trajectories were computed with OpenDrift<sup>1</sup> code (Jones et al., 2016) and windage analyzed by imposing different windage coefficients. A 4<sup>th</sup> order Runge-Kutta scheme with a bilinear interpolation and a timestep of 1 h has been used with no horizontal diffusion.

Virtual particle advection was used with three objectives: (i) to evaluate the reliability of surface current fields with different windages through the Lagrangian error, (ii) to estimate the rate of FML beaching and escaping the domain and (iii) to compute the PDF, FSLE and FDL D.

Considering  $d_{tk}$ , the separation distance between the real and the  $k$ -th simulated trajectory at time step  $t$ , the Lagrangian error was computed at each time step and averaged over all the simulated trajectories, following the expression:

$$\varepsilon_L(t) = \left\langle \frac{\sum_{k=1}^{N-(t+1)} d_{tk}}{N-(t+1)} \right\rangle, \quad (4)$$

where  $N$  is the maximum number of time steps in the real trajectories. Simulated trajectories were launched at each time step of the real trajectories. Brackets indicate the average over all the simulated trajectories.

To facilitate comparisons, the Lagrangian error was divided by the mean drift distance  $D$  of the real drifters (travel distance averaged over all the drifters), giving rise to a dimensionless index:  $L = \frac{\varepsilon_L}{D}$  (Bertin et al., 2024a).

To evaluate the rate of particles beaching on the French and Spanish coast and escaping from the northern and western parts of the domain, virtual particles were seeded and advected as dynamically distinct clusters highlighted by the SOM technique.

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<sup>1</sup> <https://opendrift.github.io/>

### 3.6. Lagrangian statistics: Absolute dispersion, Probability density functions, Finite-Size Lyapunov Exponents and Finite-Domain Lagrangian Divergence

#### *Absolute dispersion*

The absolute (cluster) dispersion  $A^2(t)$  was used to quantify the intensity of the single-particle spreading induced by the surface current field.  $A^2(t)$  is defined as the variance of particle spreading with respect to the barycenter of particles in a cluster. In two dimensions, the dispersion can be estimated along  $x$  and  $y$  axis following the expression (Berti et al., 2011; Enrile et al., 2019):

$$a_{ij}^2(t) = \frac{1}{M} \sum_{m=1}^M \left\{ \left[ x_i^m(t) - \underline{x_i^m(t)} \right] \left[ x_j^m(t) - \underline{x_j^m(t)} \right] \right\} \quad A^2(t) = a_{xx}^2(t) + a_{yy}^2(t), \quad (6)$$

where  $a_{ij}^2$  corresponds to the variance along  $i$  and  $j$  spatial coordinates,  $a_{xx}$  and  $a_{yy}$  correspond to the absolute dispersion along  $x$  and  $y$ , respectively,  $M$  is the number of drifters,  $x^m(t)$  is the position of the  $m$ -th drifter at time  $t$  and  $\underline{x^m(t)}$  is the coordinate of the barycenter.

Estimating the absolute dispersion allows us to distinguish between fundamental dispersion regimes in a turbulent flow. At short timescales (i.e., during the initial hours of drifter deployment in coastal areas,  $> 1$  day in the open ocean), the absolute dispersion generally exhibits a ballistic growth pattern, following  $A(t)^2 \sim t^2$  (Berti et al., 2011; Bouzaiene et al., 2020; Davis, 1983; Enrile et al., 2019; Poulain and Niiler, 1989; Taylor, 1922). This regime is characterized by strong correlations in the velocities of drifter pairs, influenced by persistent and robust currents such as a coastal jet. Over longer timescales, when the influence of the initial conditions dissipates, the absolute dispersion tends to follow a random-walk regime, growing as  $A(t)^2 \sim t$ .

#### *Probability Density Functions*

To estimate the zones of intense passage of the virtual particles in the study region, the probability density function of the number of particles present in each grid cell during the simulation time were computed. This metric enables to unveil the most probable location of passage of particles and is calculated as follows:

$$PDF(i, j, t) = \sum_{i=1}^{N_i-1} \sum_{j=1}^{N_j-1} \sum_{k=1}^{N_k} Ind[x(i, j), x(i+1, j)](lon(t, k)) \cdot Ind[y(i, j), y(i, j+1)](lat(t, k)), \quad (5)$$

where  $t$  corresponds to the time step,  $N_t$  the maximum number of timesteps,  $lon$  and  $lat$  the longitudes and latitudes of the virtual particles respectively.  $i$  and  $j$  are the size of the grid in longitude and latitude, respectively, and  $x$  and  $y$  the longitude and latitude coordinates of the HFR grid, respectively.  $Ind[a, b](x)$  is the indicator function of the interval  $[a, b]$ , which equals 1 if  $x$  is within the interval and 0 otherwise.

#### *Finite-Size Lyapunov Exponents*

The study of turbulent transport and dispersion mechanisms can be approached by the characterization of the coherent structures in the flow field. The development of the coherent turbulent structures such

as jets, eddies and fronts, is determined by the intrinsic spatial and temporal variability of the surface current field interacting with the external forcings (wind, waves and freshwater inflow, (Morales Márquez et al., 2023)). These dynamical structures dictate a large part of the transport characteristics (Haza et al., 2008).

The FSLEs help to identify these transport structures, providing information on velocity fluctuations and mixing in the marine flow (Aurell et al., 1997; d’Ovidio et al., 2004; Hernández-Carrasco et al., 2020; LaCasce, 2008). This method identifies the LCS that organize transport in a flow field, by locating regions of the flow that evolve on time scales different from the temporal variability of the surrounding field. Typically, these characteristic LCSs appear to be surrounded by lines of large FSLE values, also known as ridgelines (d’Ovidio et al., 2004; Hernández-Carrasco et al., 2011). As FSLEs cannot be crossed by particle trajectories, they act as barriers to transport, thus governing the movement of particles in the surrounding fluid flow around them. In this study, FSLEs are computed backward in time because, in these conditions, the resulting LCSs can be directly interpreted as attractive hyperbolic structures. Thus they have a physical interpretation and can be used to identify CCS (Bertin et al., 2024a; Hernández-Carrasco et al., 2018).

FSLE values, given by  $\lambda$ , were estimated as the inverse of the time  $\tau(x)$  required for two particles of fluid to separate from an initial distance  $\delta_0$  to a final distance  $\delta_f$ , and is expressed at position  $x$  and time  $t$  as:

$$\lambda(x, t, \delta_0, \delta_f) = \frac{1}{\tau(x)} \ln \frac{\delta_f}{\delta_0}. \quad (7)$$

The  $\lambda$ -values were computed using the algorithm described in Hernández-Carrasco et al. (2011) with  $\delta_0 = 0.4$  km and  $\delta_f = 3.2$  km (using an amplification factor  $\alpha = \delta_f/\delta_0 = 8$  in order to measure high stretching and obtain reliable attracting LCS). Particle trajectories were integrated backward-in-time over 10 days.

### *Finite-Domain Lagrangian Divergence*

FDDL computed backwards provides insights into flow regions that consistently exhibit divergence or convergence in their Lagrangian motion, revealing coherent regions of flow divergence or convergence associated with vertical dynamics. High negative values of FDDL indicate regions of particle aggregation on the sea surface, while LCS highlights areas of intense fluid stirring, where particles are not only aggregated but also dispersed (Hernández-Carrasco et al., 2018). FDDL ravines (i.e. high gradients) coincide with the attracting LCS revealed by FSLE (Hernández-Carrasco et al., 2018). Thus, using these two metrics together enables the mapping of CCS, as shown in Bertin et al. (2024a).

Assuming that the velocity field is divergent, the horizontal divergence ( $\nabla_H \cdot v$ ) accumulates along a trajectory  $s(x_0, y_0, t_0)$  in the finite domain at the sea surface. FDDL values were calculated by integrating and averaging horizontal divergence over fluid particle trajectories as follows:

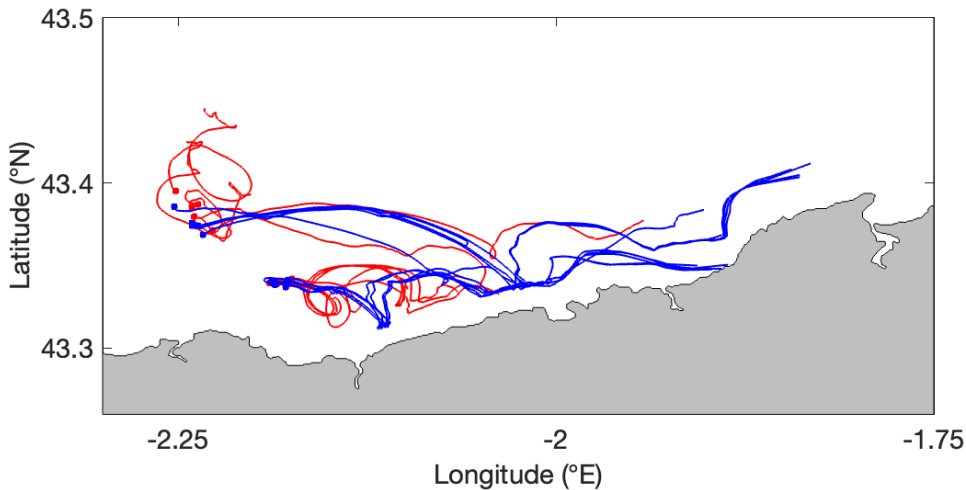
$$FDDL(x_0, y_0, t_0, t_f) = \frac{1}{t_f - t_0} \int_{t_0}^{t_f} \nabla_H \cdot v(x(t), y(t), t) dt, \quad (8)$$

with  $t_f - t_0$  the time interval of the trajectory integration and  $x_0$  (resp.  $y_0$ ) the initial longitude (resp. latitude) of the particle trajectory considered. Particle trajectories were integrated backward-in-time over 7 days.

## 4. Results

### 4.1. Trajectories of drogued and undrogued drifters

Drogued and undrogued drifter pairs trajectories are shown in Figure 3. During the first ten hours of the trajectories, moderate winds (speed up to 5 m/s) from south-southwest were observed (Figure 4a), while lighter winds (less than 3 m/s) from east-southeast (Figure 4b) dominated afterwards. Surface current velocity from HFR and drogued drifters was generally low, with a mean speed of 0.09 m/s (Figure 4c). However, surface current velocity from undrogued drifters was twice larger, with a mean speed of 0.2 m/s (Figure 4c). The predominant direction of the surface currents was east-northeast (Figure 4d).

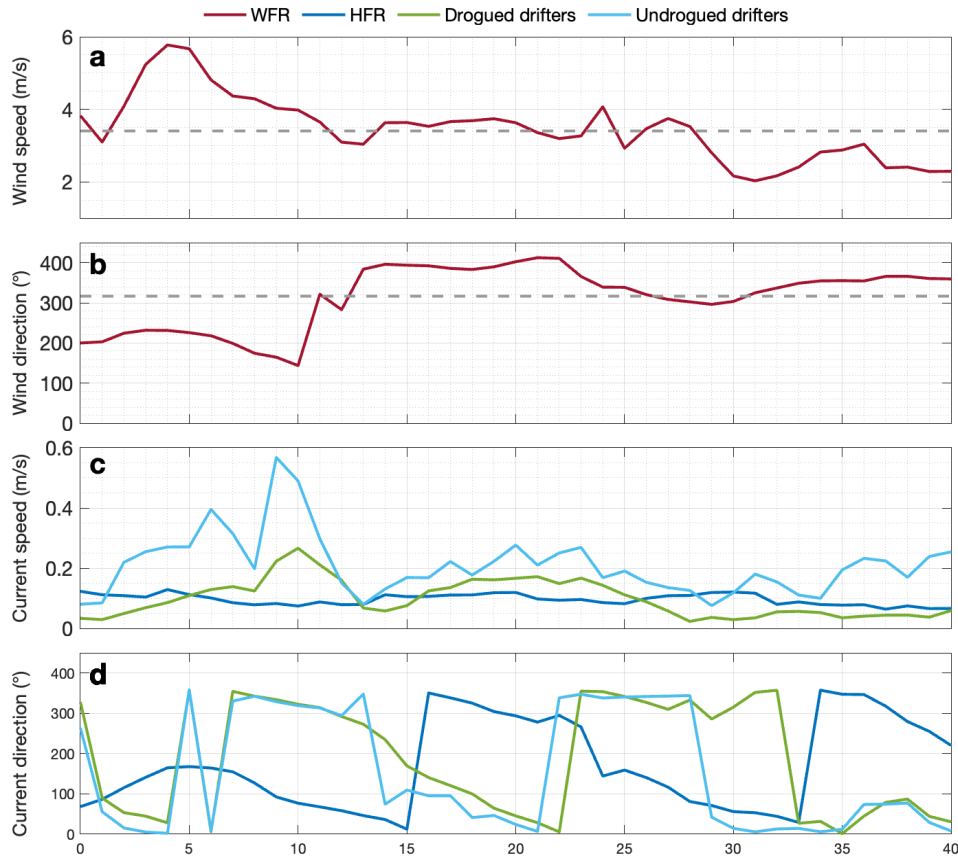


**Figure 3.** Trajectories of surface drifters with drogue (red) and trajectories of surface drifters without drogue (blue).

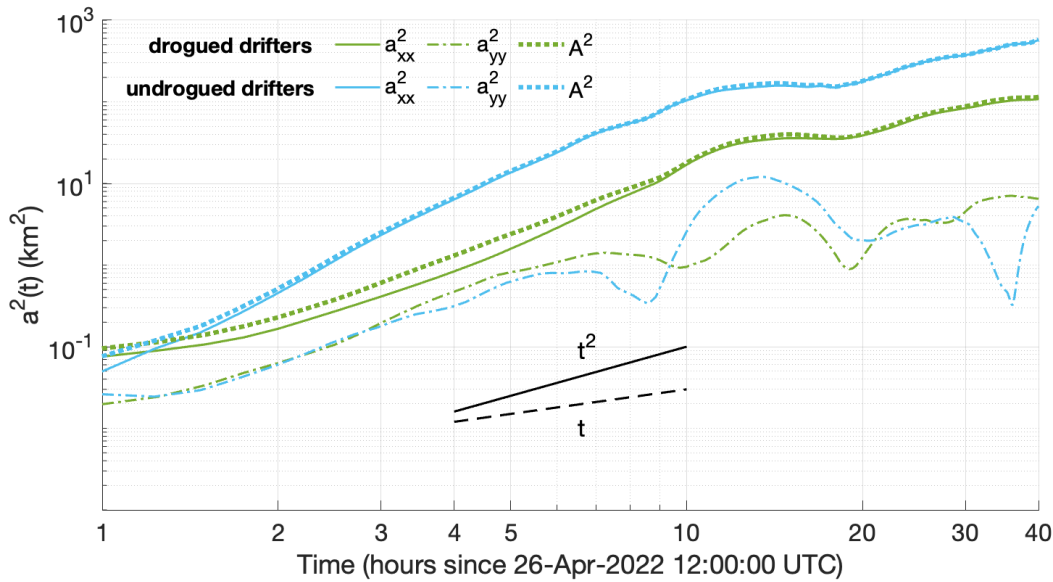
The variability of both wind and surface currents explains the observed different behaviors between drogued and undrogued drifters (Figure 4c). Figure 5 shows the absolute dispersion of real drifters along the  $x$  and  $y$  axes (aligned with along- and cross-shore directions). Both sets of drifters exhibit significantly greater dispersion along the  $x$ -axis. The dispersion along the  $x$ -axis is 16 times larger than along the  $y$ -axis for drogued drifters and 47 times larger for undrogued drifters. The difference in dispersion along the  $x$  and  $y$  axes is attributed to the shape of the Spanish coastline and the coastal jet, which generates a predominantly eastward along-shore transport. With only minor oscillations along the cross-shore  $y$ -axis, these results highlight a marked anisotropy in spreading. Additionally, the varying levels of dispersion between drogued and undrogued drifters can be explained by the windage effect.

Furthermore, the absolute dispersion,  $A^2(t)$  is 5 times larger for undrogued drifters than for drogued drifters. Undrogued drifters experience an additional wind-induced velocity, leading to increased greater travel distances: 33.6 km for undrogued drifters and 19.4 km for drogued drifters.

Except for the difference in order of magnitude between the absolute dispersion of drogued and undrogued drifters, the dispersion regimes are similar. Indeed, during the first ten hours of the survey, the absolute dispersion of both types of drifters follows a ballistic regime characterized by  $A^2(t) \sim t^2$  (Figure 5). Subsequently, as the separation distance between drifters increases, the absolute dispersion transitions to a diffusive regime, characterized by  $A^2(t) \sim t$ .



**Figure 4.** Time series of (a) wind speed and (b) direction. Dash lines in panels (a) and (b) correspond to the temporal mean of each variable. (c) Surface current speed, and (d) direction, spatially averaged over the study obtained from HFR and from averaged drogued and undrogued drifter data.

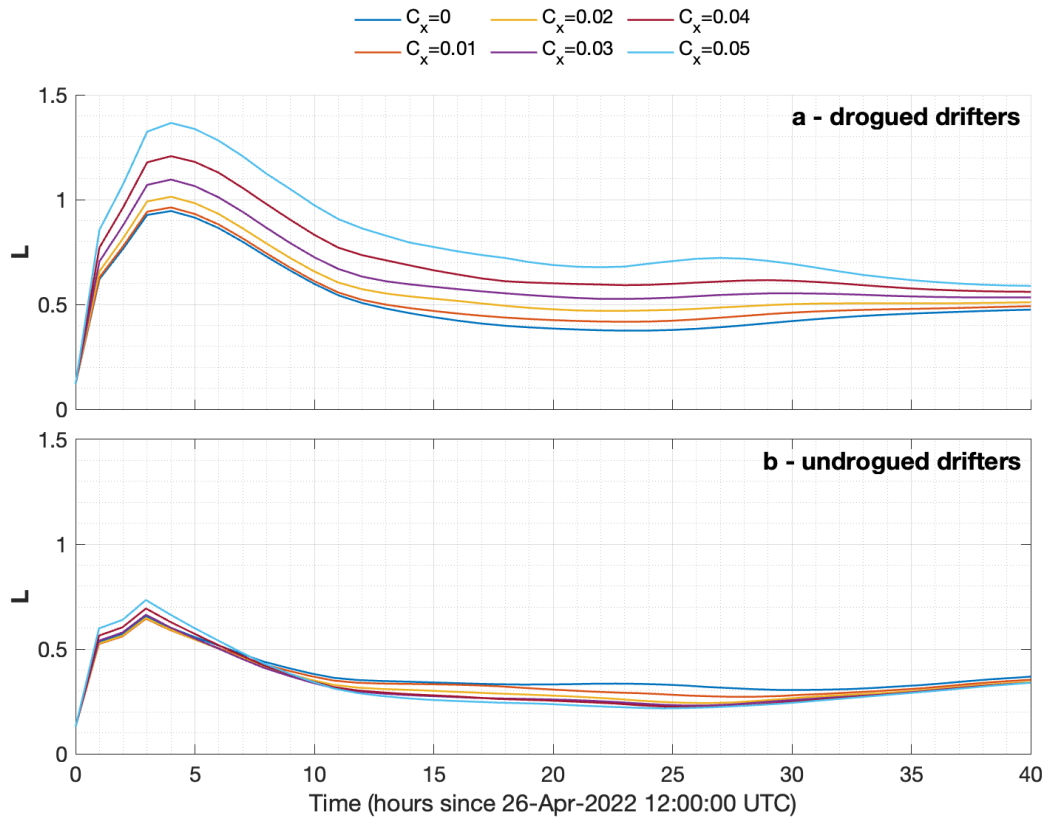


**Figure 5.** Absolute dispersion considering drogued and undrogued drifters. The absolute dispersion along  $x$ -axis (*along-shelf*) is given in solid line, the absolute dispersion along  $y$ -axis (*cross-shelf*) is given in dash line and the total absolute dispersion is given by the bold dot line. The ballistic ( $t^2$ ) and diffusive ( $t$ ) regimes are shown by solid and dashed black lines, respectively.

Considering the different dispersion experienced by drogued and undrogued drifters, it is useful to compute the Lagrangian error index ( $L$ ) between real and virtual drifters to determine the windage coefficient most suited to each type of drifter. Thus, the calculation of the Lagrangian error was performed with surface currents including a varying wind-induced velocity (from  $C_x = 0$  to  $C_x = 0.05$ ) and results are presented in Figure 6 and Table 1. For drogued drifters, the lowest  $L$  value (0.52) is obtained using  $C_x = 0$ . This is consistent because drogued Lagrangian drifter motions are designed to be driven solely by surface currents.

For undrogued drifters, the lowest  $L$  value (0.33) is obtained for  $C_x = 0.03$ . This is consistent with the study by Ruiz et al. (2022), where they used the same handmade surface drifters and obtained  $C_x = 0.04$ . The slight difference can be attributed to the different datasets used for the calculations (wind and surface currents).

The difference in  $L$  between drogued and undrogued drifters (0.52 for drogued drifters versus 0.33 for undrogued drifters) can be explained by the trajectory shapes of the undrogued drifters, which are more rectilinear and easier to resolve by the HFR than the oscillating trajectories of drogued drifters (Figure 3).



**Figure 6.**  $L$  index computed with drogued drifters (a) and undrogued drifters (b) upon trajectories with various windage coefficient  $C_x$ .

These results highlight the necessity of considering windage effects when studying the transport of FML at sea. Depending on their shape, and particularly the area of emerging parts, different types of FML can exhibit significantly different trajectories, impacting on the absolute travel distances, pathways, beaching, preferential accumulation and escape ratios.



**Table 1.** Mean Lagrangian error index  $L$  depending on the windage coefficient for both drogued and undrogued drifters. Lower  $L$  values are underlined.

	$C_x = 0$	$C_x = 0.01$	$C_x = 0.02$	$C_x = 0.03$	$C_x = 0.04$	$C_x = 0.05$
<b>Drogued drifters</b>	<u>0.52</u>	0.59	0.57	0.64	0.71	0.80
<b>Undrogued drifters</b>	0.39	0.36	0.34	<u>0.33</u>	0.34	0.34

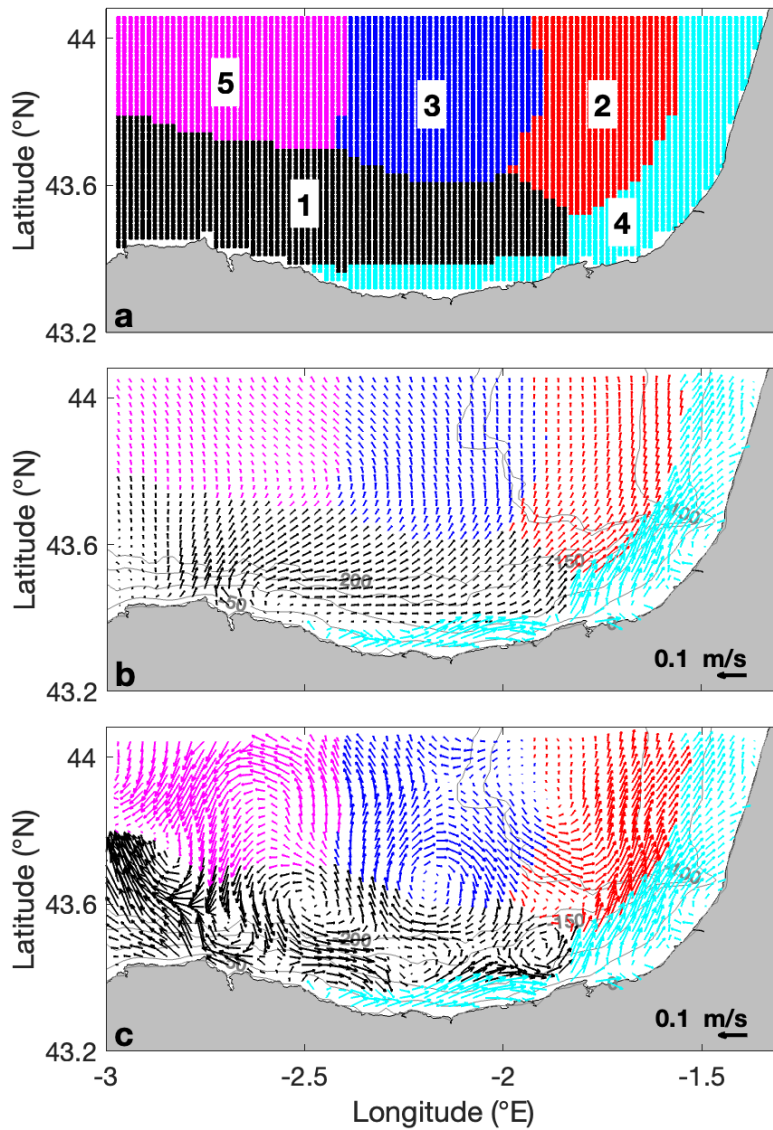
## 4.2. Coastal water dynamics under different windages

The SOM technique enabled to unveil different coherent regions of similar velocity field variability. The SOM regions 1 and 2 (Figure 7a in dark and red) correspond to the area of influence of the slope current (Rubio et al., 2019). SOM regions 3 and 5 correspond to regions where surface currents are weak and variable in direction due to the presence of eddies. SOM region 4 corresponds to the Region of Freshwater Influence (ROFI) (Reverdin et al., 2013). The deployment locations of drifters during the survey are in region 4.

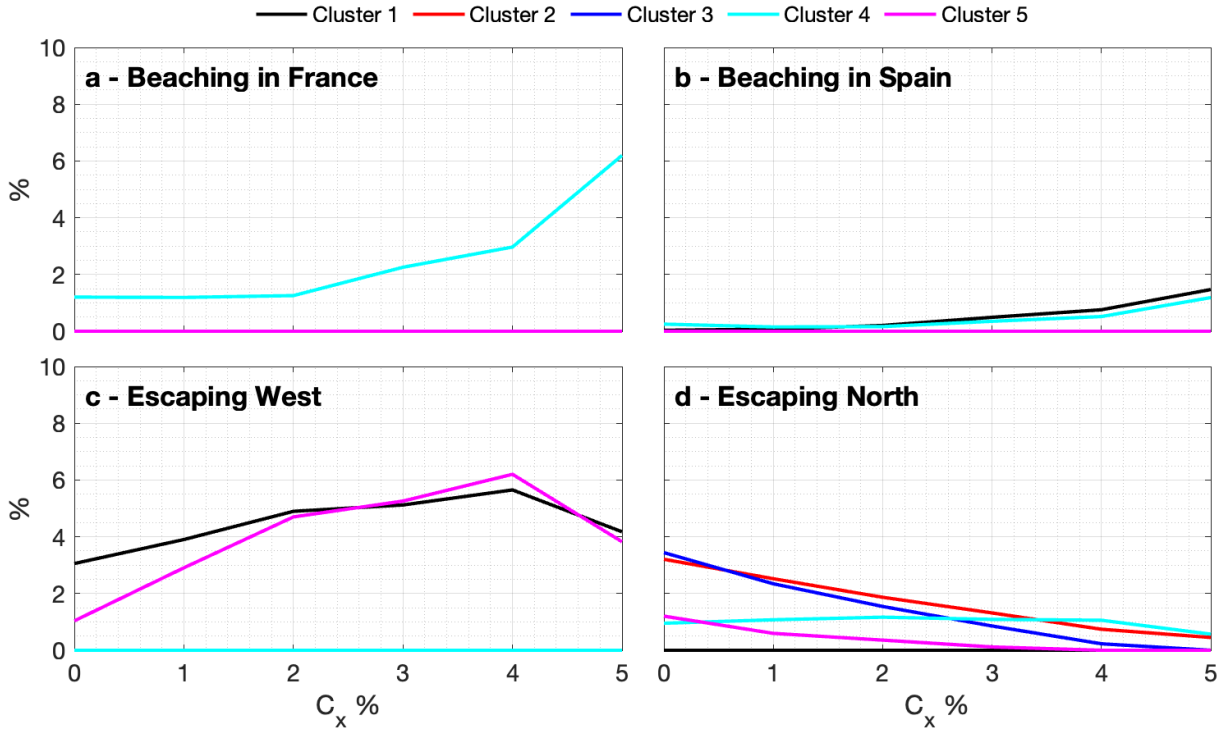
Figure 7b shows the mean surface current velocity (averaged over three years), which reveals a significant coastal jet in SOM regions 1, 2, and 4. Low mean velocities are observed in drifters deployed in regions 3 and 5 (under 1 m/s), with an overall transport towards the north-northeast. Figure 7c shows the mean surface current velocity observed by HFR during the survey, revealing weak variable currents and gyres in regions 1, 3, and 5 and a coastal jet which remains present and intense in regions 1, 2, and 4.

Virtual particles were deployed in all the study area and differentiated in clusters based on the different dynamical regions unveiled by the SOM (Figure 7a). Particles were advected with surface currents and considering various windage coefficients. A total of 5,988 particles have been deployed: 1,768 in cluster 1 (SOM region 1), 976 in cluster 2 (SOM region 2), 1,176 in cluster 3 (SOM region 4), 984 in cluster 4 (SOM region 4) and 1,084 in cluster 5 (SOM region 5). This deployment strategy enabled the study of the effect of wind in the particle's trajectories depending on their initial location in relation with the main dynamical regions, so the behavior of particle clusters located in regions dominated by the coastal jet can be differentiated from that of particles deployed within the ROFI or at offshore locations.

Figure 8 presents the beaching ratios in France and Spain, as well as the ratios of virtual particles escaping north and west of the study area for different windages and considering realistic modeled winds. The beaching ratios in France and Spain increase with windage, primarily due to contributions from cluster 4 in France and clusters 1 and 4 in Spain. The rate of virtual particles escaping from the western side of the study area also increases with windage up to  $C_x = 0.04$ , influenced by clusters 1 and 5, before decreasing from  $C_x = 0.04$  to  $C_x = 0.05$  due to the wind direction (east-southeast during most of the survey - Figure 4b). Conversely, the rate of virtual particles escaping from the northern part of the study area decreases with increasing windage.



**Figure 7.** Distinct dynamic regions revealed by the SOM clustering method, with the number of each region ranked in order of occurrence (a). Mean surface current velocity averaged over the 3-years of training data (b) and averaged over the survey period (c).



**Figure 8.** Beaching and escaping ratios of numerical particle clusters deployed in the five SOM regions shown in Figure 7a, depending on the windage coefficient and considering realistic modeled winds.

### 4.3. Effect of windage on the coastal Current Convergence Structures

FSLEs were computed from a combination of surface currents and winds by using different windage coefficients. FSLE snapshots presented in Figure 9 show that for  $C_x = 0$ , ridgelines are located parallel to the French and Spanish coasts, which is consistent with the alongshore filamental aggregation of FML observed in the study area (Ruiz et al., 2020).

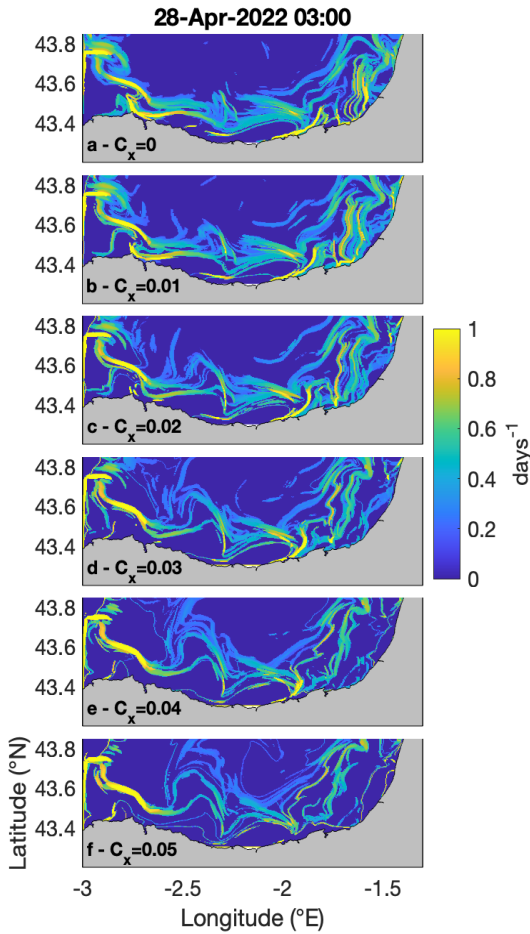
When accounting for windage below  $C_x = 0.03$ , FSLE ridgelines remain present but are overall less defined, in accordance with the FSLE mean values presented in Table 2 ( $0.51 \text{ days}^{-1}$  for  $C_x = 0$ , and  $0.47 \text{ days}^{-1}$  for  $C_x = 0.01$  and  $C_x = 0.02$ ). The FSLE maximum slightly increases with windage (Table 2 - from  $2.29 \text{ days}^{-1}$  with  $C_x = 0$  to  $2.43 \text{ days}^{-1}$  with  $C_x = 0.03$ ) but is always located in the same area on the Spanish coast (i.e.  $-1.9^\circ\text{E}$ ,  $43.4^\circ\text{N}$ ).

In contrast, above  $C_x = 0.03$ , FSLE ridgelines become less defined, and the thin filaments observed at  $C_x = 0$  appear to merge at windages of  $C_x = 0.04$  and  $C_x = 0.05$ , particularly on the French side of the study area. However, it also causes the disappearance of small-scale structures, especially on the French side of the domain, leading to a decrease in the FSLE maximum (Table 2 - from  $2.43 \text{ days}^{-1}$  at 3% windage to  $1.59 \text{ days}^{-1}$  at 5% windage). Although the spatial distributions of the main LCS are rather similar for different windage coefficients, new LCS emerge as  $C_x$  increases (i.e. in the central region) while other LCS disappear near the French coast.

The decrease in the number of FSLE ridgelines with higher windage is associated with increased beaching and escaping ratios, but also with the fusion of ridgelines. As windage increases, the numerous

convergent structures present in the study area (i.e.  $-1.6^{\circ}\text{E}$ ,  $43.5^{\circ}\text{N}$ ) with  $C_x = 0$  gradually merge into a single, wider structure. It is important to note that the disappearance of transport barriers and the easier passage of particulate matter are much more pronounced on the French side compared to the Spanish side, for the wind conditions corresponding to the field survey.

Knowing that FSLE ridgelines structure the transport of particulate matter, FSLE values above  $0.7 \text{ days}^{-1}$  were superimposed on the PDF and FDL D fields. Snapshots are presented in Figure 10 with PDF superimposed on FSLE (left panel) and FDL D superimposed on FSLE (right panel), only for  $C_x = 0$ ,  $C_x = 0.03$  and  $C_x = 0.05$ .



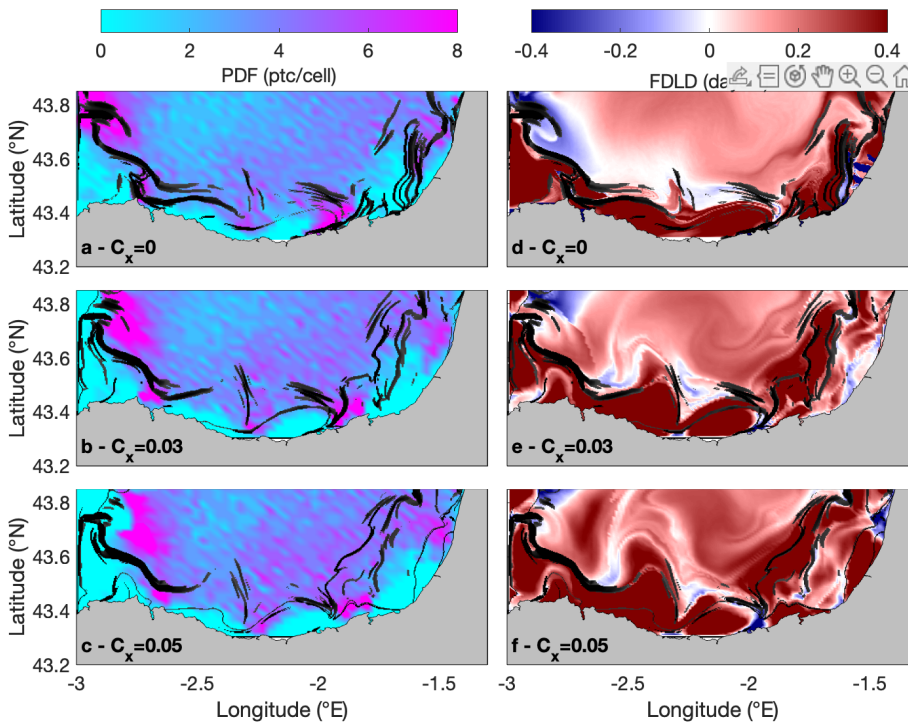
**Figure 9.** FSLE snapshots under different windage coefficients and considering modeled realistic winds.

**Table 2.** FSLE maximum and mean for the snapshot (April 28th, 2022, 03:00) presented in Figure 9.

	$C_x = 0$	$C_x = 0.01$	$C_x = 0.02$	$C_x = 0.03$	$C_x = 0.04$	$C_x = 0.05$
<b>Max</b>	2.29	2.32	2.30	2.43	1.84	1.59
<b>Mean</b>	0.51	0.47	0.47	0.49	0.50	0.51

FSLE ridgelines align closely with the spatial distribution of PDF (Figure 10, left panel), often surrounding areas with maximum values of PDFs. Maximum values of PDFs delineate areas of intense passage of virtual particles and are located in the northwest of the study region and close to the coast. These intense passage zones tend to be located in similar areas, regardless of the windage and wind direction considered, as presented in Appendix (Figure B2). FSLE ridgelines also align closely with the ravines of FDL D, indicating significant gradients in the FDL D field (Figure 10, right panel). Considering the windage effects, both the FSLE ridgelines and the areas with high values of PDF and FDL D ravines shift towards the northwest. Despite this shift, the FSLE ridgelines continue to align closely with the PDF and FDL D fields.

Additionally, a pronounced divergence alongshore is observed, which is consistent with the dominant coastal jet in the study region (Figure 7b, c).



**Figure 10.** PDF snapshots on April 28<sup>th</sup>, 2022, at 03:00 with superimposed FSLE ridgelines (a, b, c) and FDL D snapshots on the same date with superimposed FSLE ridgelines (d, e, f). The FSLE ridgelines have been plotted above the threshold of 0.7.

## 5. Discussion

Our results showcase how windage affects ocean surface transport, influencing the distribution and fate of FML, particularly. This impact is manifest in the varying fates and trajectories of virtual particles influenced by increasing windage. We suggest that wind-induced velocity plays a significant role in the behavior of FML transport and aggregation at the ocean surface. Although both drogued and undrogued drifters experience the same dispersion regime (a ballistic regime during the first ten hours of deployment, followed by a random-walk regime, as shown in Figure 5), the magnitude of absolute dispersion for undrogued drifters is five times greater. Wind tends to increase both the absolute

dispersion, and the distance traveled. Moreover, it significantly influences both the beaching and escape ratios and locations.

LCSs, provided by FSLE ridgelines, seem to act as transport barriers for FML, as evidenced by the distribution of high values in the PDF fields aligned with FSLE ridgelines. Moreover, high coherence is observed between the FSLE ridgelines, FDL and transport patterns for different windages, with a pronounced divergence alongshore in the studied period. Allshouse et al. (2017) were the first to assess the impact of windage on LCS. They demonstrated that the windage was influencing the transport of material patches by altering their initial positions and the arrangement of strainlines around Lagrangian hyperbolic saddle-points. Windage shifts and reshapes these structures rather than creating or destroying them, which is crucial for interpreting ocean transport dynamics. The calculation of FSLE and FDL with surface currents and wind, using different windage coefficients, demonstrated the effect of wind in CCSs location and intensity. This result underlines that the trajectories of different types of FML, submitted to different windage, are subject to different barriers to transport, depending on the wind drag they experience along their trajectory. However, the accurate implementation of the combined effect of currents and wind in the computation of transport is challenging, since it relies heavily on the quality of surface velocity and wind field data (Bertin et al., 2024b, 2024a). The use of drifters and HFR radar data have proven to provide good approximation to LCSs for the period analyzed, but the characterization of transport and aggregation paths along seasonal or interannual periods would require longer and reliable data sets. The progress in improving coastal models through atmosphere ocean coupling, the implementation of real time riverine runoff, and multi-platform data assimilations (including HFR data) or the use of data blending techniques is a key element towards accurate monitoring and forecasting of FML hotspots in the coastal ocean.

In this study, we estimated the impact of windage on the transport of FML using realistic surface winds. However, the results are only representative for the specific conditions during the analyzed period. In order to discuss the effect of wind from different directions, we have also computed escape rates, beaching and Lagrangian quantities under different theoretical wind patterns dominant in the study area. The results of these calculations are presented in Appendix B, and Table 3 presents the conclusions compared to those obtained with realistic winds.

Results in table 3 show that winds from the north (N, NE and NW) are the least favorable for particle evacuation in the area as they lead to a decrease in escape through the northern border (also the western border, in the case of north-westerly winds) and are the most favorable for beaching on the Spanish coast. South-westerly winds increase escape through the northern border but also increase beaching on the French coast. The impact of different wind directions on FSLE, FDL, and PDF can be summarized by a shift in the locations of LCSs and regions of higher passage, in the direction of the prevailing winds. In general, we observe no significant changes in the number of observed LCSs or areas of intense passage for most of the cases, excepting the area off the French coast where the number of LCS drastically decreases with north-westerly winds. North-westerly winds have the worst impact in the area, inducing a decrease in escape through both the northern and western borders and an increase in beaching on both the French and Spanish coasts. These results highlight the importance of this type of analysis to provide a basis for planning cleanup or mitigation strategies to face the issue of pollution by FML. The aggregation of FML along LCSs is crucial for active fishing for litter activities which would be extremely inefficient otherwise, as discussed in Ruiz et al. (2020). To understand the complex distribution of FML due to the effect of LCSs is also important for the interpretation and planning of future sampling surveys aiming to quantify and map this pollution at the sea. The presence of sub-

mesoscale LCSs is also relevant for other types of particles and passive tracers, thus our results may have other applications related to ocean biogeochemical cycles or marine biology.

**Table 3.** Summary of impacts of different wind directions with increasing windage on the beaching, escaping rates, PDFs, FSLE and FDLF compared with those for WRF realistic winds. Conclusions about escaping and beaching ratios are based on Figures 8 for WRF winds and B3 for theoretical winds. Conclusions about the FSLE and FDLF are based on Figures 9 and 10 for WRF winds, and Figures B4 and B5 for theoretical winds. Conclusions about PDF are based on Figure B2.

	<b>WRF winds</b>	<b>N winds</b>	<b>NE winds</b>	<b>NW winds</b>	<b>SW winds</b>
<b>Escaping N</b>	Decreasing with windage	Decreasing with windage	Slightly decreasing with windage	Decreasing with windage	Increasing with windage
<b>Escaping W</b>	Increasing with windage under 4%, then decreasing	Constant with windage	Increasing with windage	Decreasing with windage	Decreasing with windage
<b>Beaching F</b>	Increasing with windage	Increasing with windage	Decreasing with windage	Increasing with windage	Increasing with windage
<b>Beaching S</b>	Slightly increasing with windage	Increasing with windage	Increasing with windage	Increasing with windage	Slightly decreasing with windage
<b>FSLE &amp; FDLF</b>	- Displacement towards north-west under 3% windage - Fusion of LCS and less structures observed above 3% windage	- Displacement towards south under 3% windage - Fusion of LCS and less structures observed above 3% windage on the French side	- Displacement towards south-west under 3% windage - Fusion of LCS and less structures observed for 3-4% windage	- Displacement towards south-east under 3% windage - Fusion of LCS observed above 3% windage on the Spanish side - Less LCS observed on the French side	- Displacement towards north under 3% windage - Fusion of LCS and less structures observed above 3% windage
<b>PDF</b>	Intense passage zone located in the same areas under 3% windage, vanishing above 3% windage	Intense passage zones located in similar areas regardless of windage	Intense passage zones slightly shifted towards west with increasing windage	Intense passage zones located in similar areas regardless of windage	Intense passage zones shifted towards north-east with increasing windage

In addition to wind drag, another process which may have a significant effect on the transport of passive particles in the coastal area is the Stokes drift—the current velocity induced by waves. Research indicates that this additional velocity can significantly alter the transport pathways of passively advected particles (Curcic et al., 2016; Dobler et al., 2019; Morales Márquez et al., 2023; van den Bremer and Breivik, 2018). Stokes drift is estimated to be about 1% of the wind speed (Ardhuin et al., 2018, 2012). The inclusion of Stokes drift in HFR measurements remains contentious. Some researchers argue that

HFR do not measure Stokes drift (Röhrs et al., 2015; van den Bremer and Breivik, 2018), while others disagree (Ardhuin et al., 2009; Chavanne, 2018; Dussol et al., 2022). However, Breivik and Allen (2008) showed that wave drift forces on small objects (i.e., less than 30 m) decay rapidly as the ratio of the dominant wavelength to the object's length increases. These forces become negligible compared to wind forces when the wavelength is approximately six times the object's length (Hodgins and Hodgins, 1998; Mei, 1989). For the target period analyzed here, mean significant wave height was around 1 m (source euskoos.eus), so we expect the impact of Stokes drift on the transport of FML to be very small. However, the effect of Stokes drift could be significant in other periods and study areas with high exposure to energetic waves and deserves further attention in future works. Lagrangian simulations performed in this work have also some limitations which could be improved in future works, such as more accurate beaching parameterization and more realistic behaviors of particles at the open boundaries (for instance, allowing particles escaping through the open boundaries to reenter the domain under favorable conditions).

Perspectives inherent to this study are to focus on the three-dimensional aspects of the CCS. Current convergence and divergence processes are closely linked to vertical movements, which can impact the distribution of particulate matter in the water column. Sub-mesoscale structures are associated with horizontal density gradients (i. e., frontal zones) with strong surface convergence and water subduction. These structures can easily generate a three-dimensional distribution of material initially drifting to the surface (D'Asaro et al., 2018; Essink, 2019). Understanding these three-dimensional transport processes is crucial, as they contribute to vertical mixing. These processes control the supply of nutrients necessary for the growth of marine organisms, impacting the fate of many biogeochemical variables (Gove et al., 2019). On the other hand, toxic materials such as hydrocarbons or marine waste, accidentally spilled at sea, can be carried to depth by downward movements, amplifying their impact on the ecosystem (Cózar et al., 2014; D'Asaro et al., 2018). Plastic waste, in particular, is subject to biofouling (the accumulation of microorganisms, plants, algae and animals on submerged structures) and tends to fragment, increasing its presence in the water column (Lobelle et al., 2021). In the SE-BoB, samplings were performed in coastal CCS enabling the gathering of 16.2 tons of FML where plastic was the predominant material, making up 96% of the items collected (Ruiz et al., 2020). Thus, the three-dimensional aspect of coastal CCS is an area for further study. In addition, Lagrangian diagnostics such as FSLE and FDL, which have proven their worth in identifying surface current convergence structures, can also be used to infer three-dimensional LCS and the impact of the vertical dynamics on the transported particles, such as plastics, organisms, etc (Bettencourt et al., 2013; Taani et al., 2019).

## 6. Conclusion

Our analysis provides significant insight into the impact of windage on FML transport and accumulation in coastal environments. When windage is below 2-3%, the aggregation structures of FML are primarily influenced by the combined effect of ocean currents and prevailing winds, resulting in a shift of these structures in alignment with these forcings. In contrast, when windage exceeds 3%, we observe a marked increase in beaching and escaping ratios of FML, accompanied by a reduction in the number of distinct aggregation structures. This reduction occurs because higher windage causes higher escape and beaching and also causes these structures to merge. Despite variations in wind direction, the zones with intense FML passage tend to be located in consistent areas. The origin of the FML loads within these aggregation zones can vary depending on the specific characteristics of FML items and their exposure to windage, suggesting the importance of accurately accounting for the interplay between windage and



surface currents when using Lagrangian tools to study FML transport. Tools such as FSLE and FDL D prove to be invaluable for identifying CCS and elucidating the transport pathways of FML. By integrating this method, coastal management practices could be significantly enhanced, providing clearer insights into the underlying transport structures and ultimately leading to more informed decisions in managing coastal and estuarine environments.

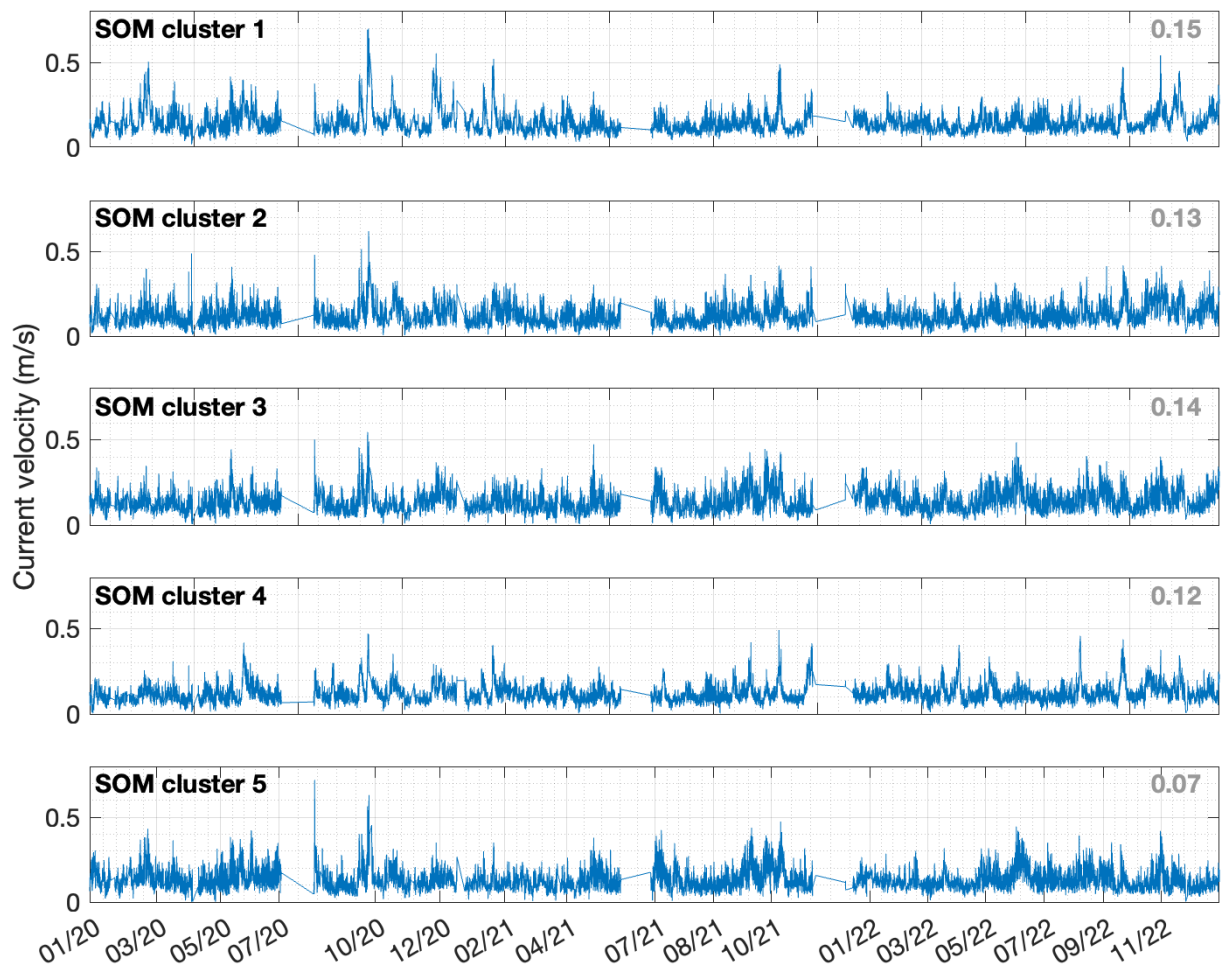
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## **Declaration of interests**

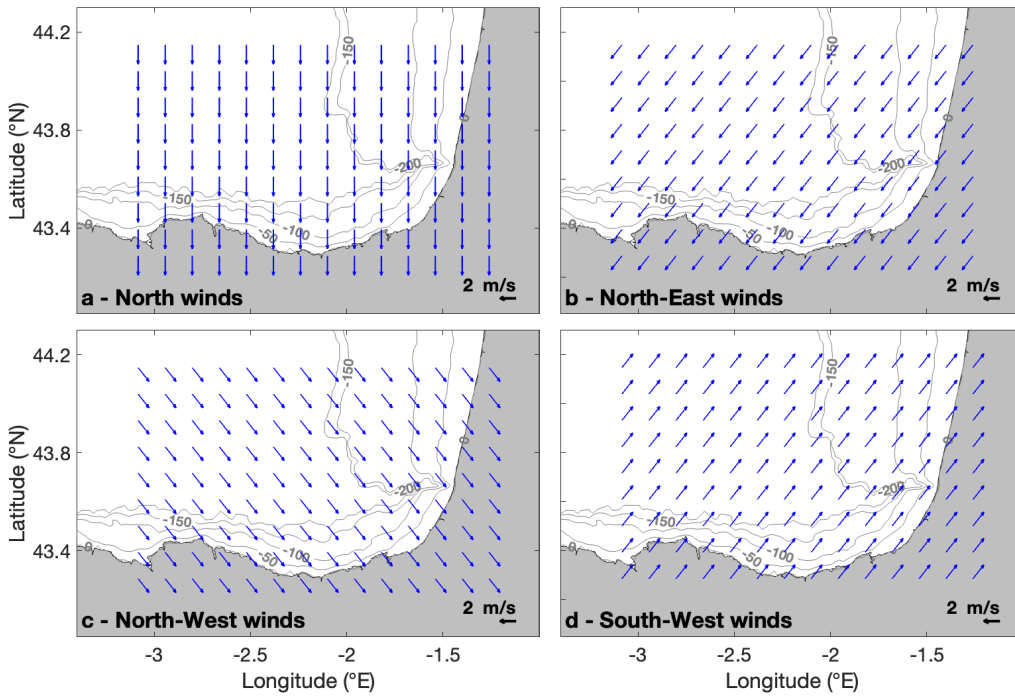
The authors declare no competing interests.

## Appendix A: Temporal patterns of SOM clusters

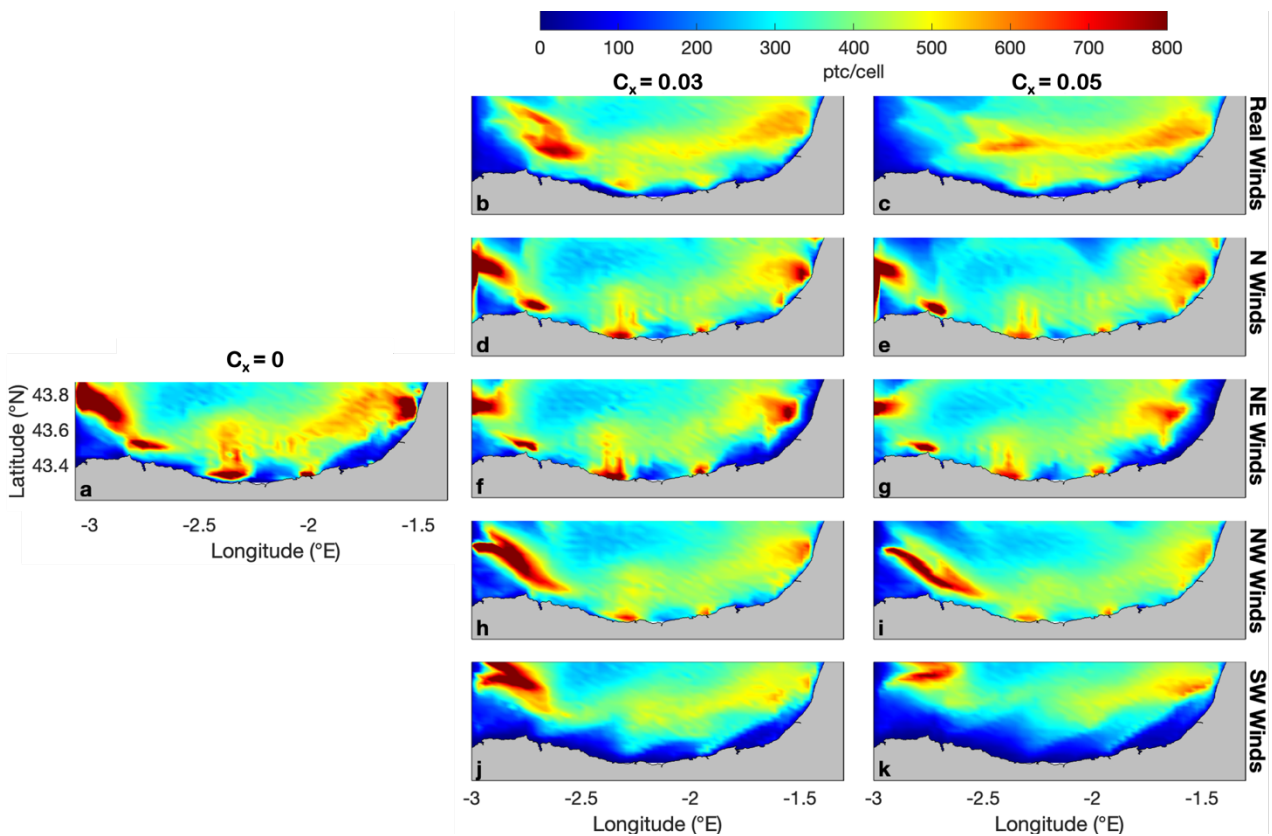


**Figure A.** Temporal patterns of surface current velocity associated with each coherent region during the 3 years of SOM training. The mean velocity is shown in gray at the top right of each figure.

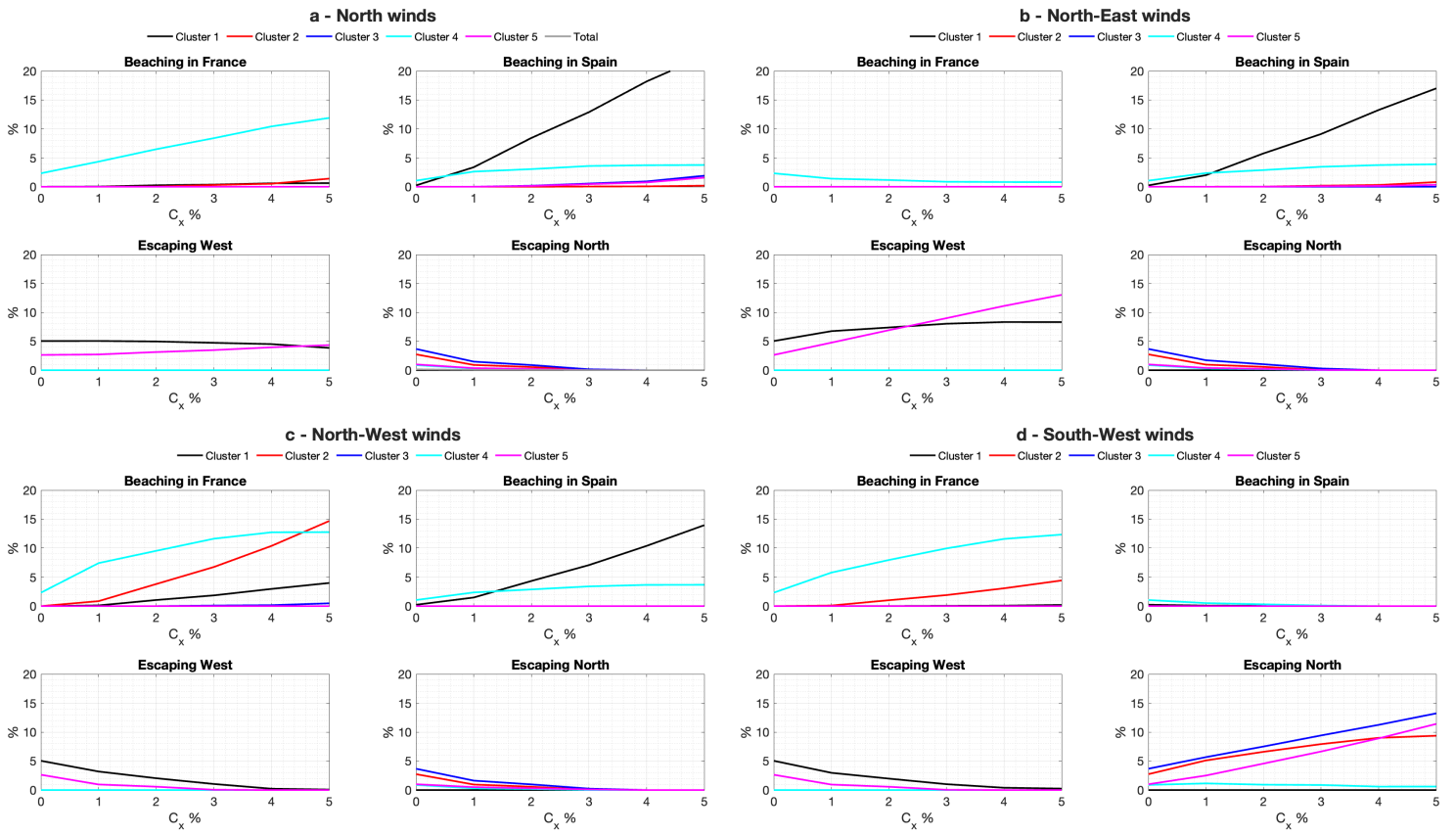
## Appendix B: Theoretical wind regimes



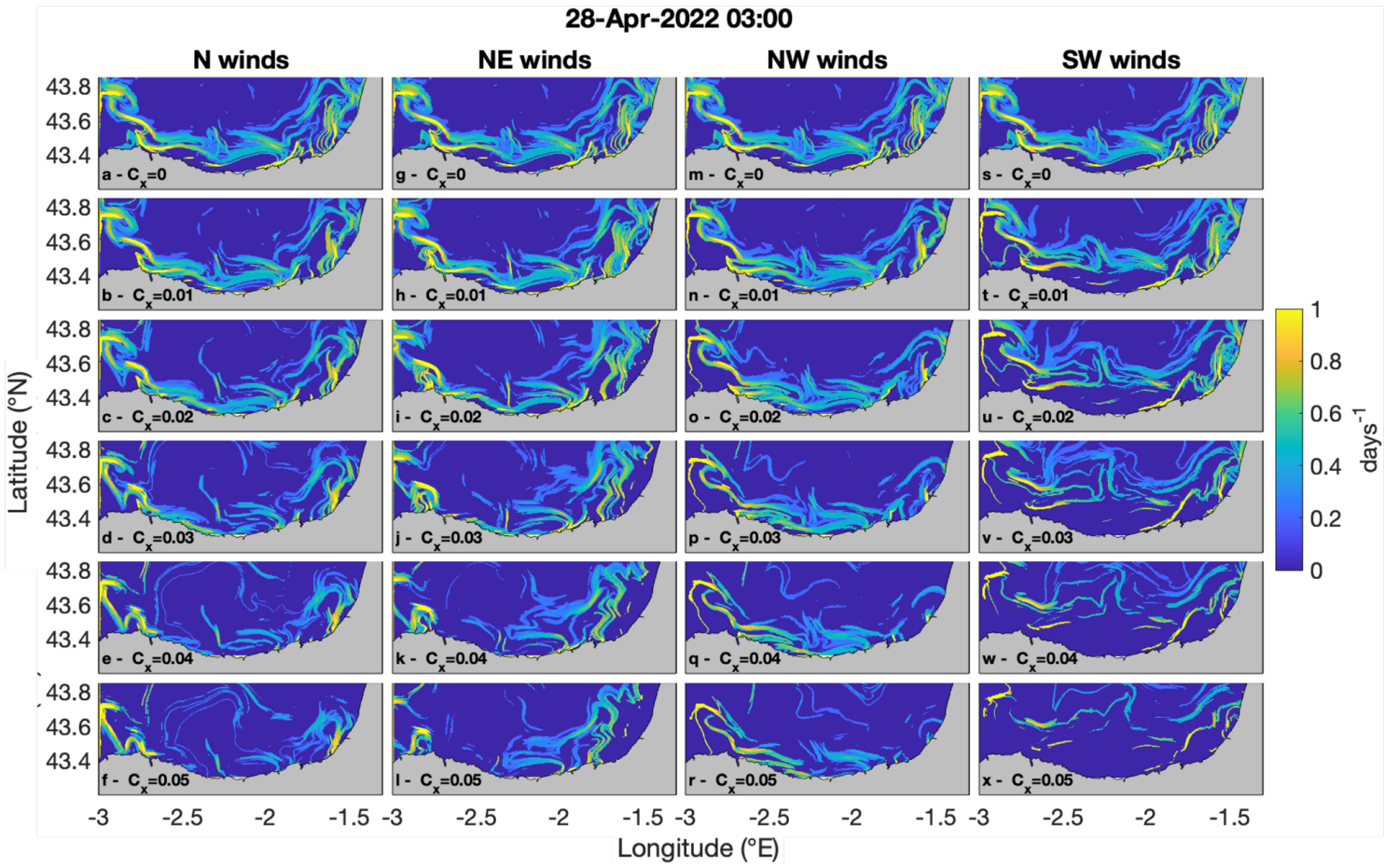
**Figure B1.** Four theoretical wind regimes used in the computation of surface current fields with windage used to estimate escaping and beaching ratios, PDF, FDL and FSLE fields.



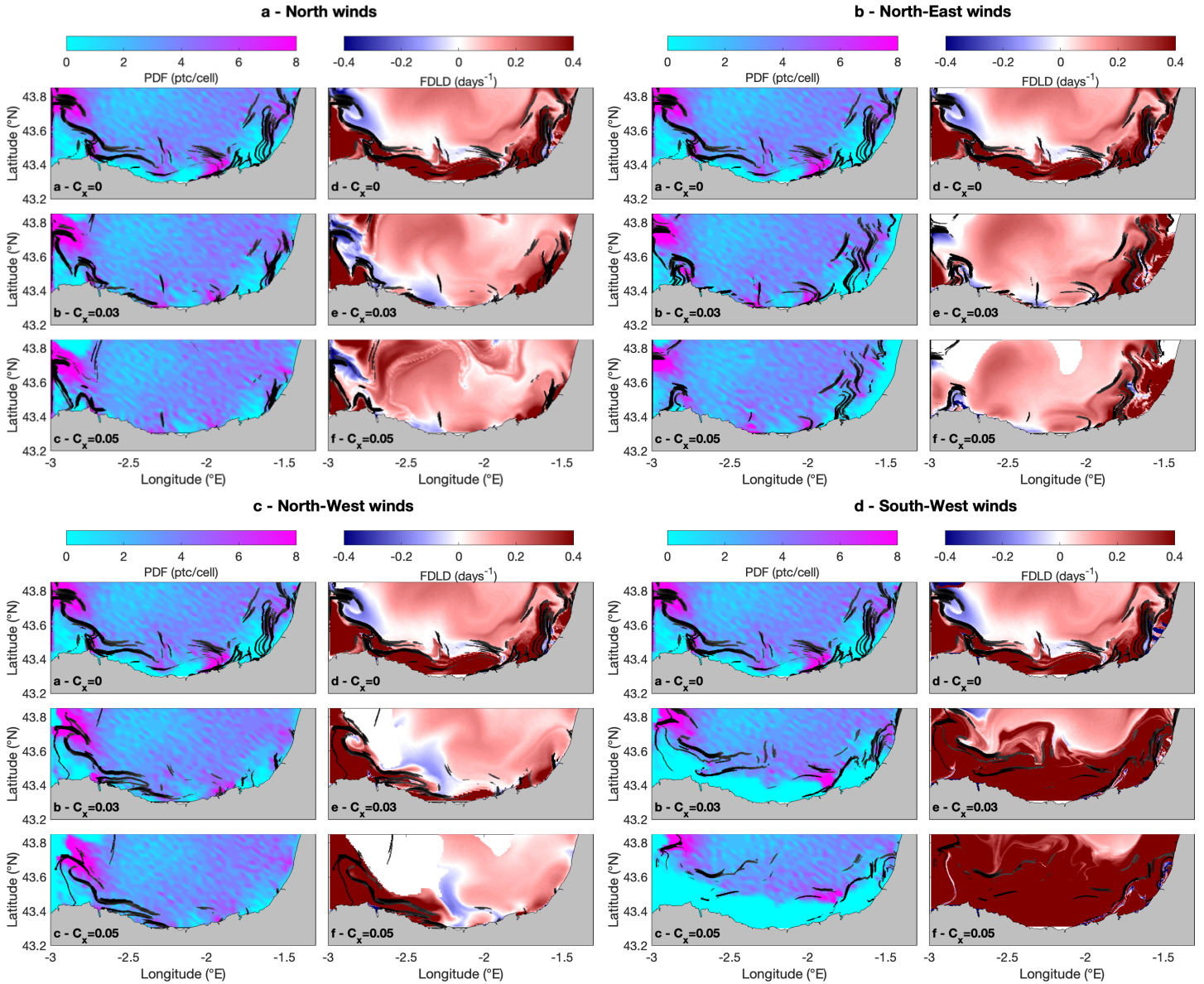
**Figure B2.** Summed PDF for the different theoretical wind directions and for a windage of 0% (left panel), 3% (middle panel) and 5% (right panel).



**Figure B3.** Beaching and escaping ratios depending on the windage coefficient and the theoretical wind direction.



**Figure B4.** FSLE snapshots computed with the 2dVar HF radar fields with different windage coefficients considering four theoretical winds.



**Figure B5.** PDF snapshots on April 28<sup>th</sup>, 2022, at 03:00 with superimposed FSLE ridgelines (a, b, c) and FDL D snapshots on the same date with superimposed FSLE ridgelines (d, e, f). The FSLE ridgelines have been plotted above the threshold of 0.7. PDF, FDL D and FSLE were computed using theoretical winds: north winds (A), north-east winds (B), north-west winds (C) and south-west winds (D).

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## Chapitre 6

# Conclusions générales et perspectives

L'objectif de cette thèse était de proposer une méthode de caractérisation des structures de convergence des courants marins en zone côtière, ainsi que de leur distribution spatiale et de leur variabilité temporelle, en utilisant une approche de fusion de données hétérogènes. Ce travail s'est focalisé sur deux régions d'étude bien différentes : la Manche orientale et le sud-est du golfe de Gascogne. La dynamique côtière en Manche orientale et dans le sud-est du golfe de Gascogne est complexe, avec des structures à petite échelle qui agrègent les particules passives sur des échelles de temps courtes.

À sous-mésoéchelle, en particulier dans les zones côtières, la détection des structures de convergence des courants constitue un défi. Leur nature chaotique complique leur représentation précise par les modèles de circulation ou leur capture par les systèmes d'observation, souvent limités par la résolution spatio-temporelle. Dans cette thèse, j'aborde ce défi en mettant en place une méthode d'optimisation efficace combinant des données lagrangiennes à la modélisation numérique en Manche orientale, et aux radars HF dans le golfe de Gascogne, afin d'obtenir une estimation plus fiable de la dispersion turbulente. Dans les deux zones d'étude, cette technique s'est avérée efficace et a permis une réduction de moitié de l'erreur entre les vitesses des bouées et les vitesses modélisées ou observées par radar HF.

Cette technique d'optimisation m'a permis de quantifier de manière plus précise la convergence des courants marins en calculant diverses métriques lagrangiennes, telles que la dispersion absolue, les FSLE et la FDL, en s'appuyant sur des champs de courants de surface plus réalistes. L'utilité de ces grandeurs lagrangiennes pour identifier les structures de convergence des courants marins a été démontrée. En effet, il a été prouvé que les lignes de crête des FSLE ainsi que les gradients importants de FDL correspondent à des structures cohérentes lagrangiennes, qui organisent le transport de matière particulaire à la surface de l'océan. Ces dernières délimitent en particulier les eaux provenant du continent et riches en Chl-a et ont un rôle crucial dans le transport et l'accumulation des déchets marins flottants.

La variabilité spatiale de la dispersion et des zones de convergence a été étudiée, permettant d'identifier des zones très turbulentes, des zones de passage intense, ou encore des zones de brassage. Effectivement, pendant la période étudiée, le nord du Cap Gris Nez, en Manche orientale, s'est révélé être une zone très turbulente où se forme un tourbillon résiduel dû aux courants de marée, impactant fortement le transport de matière particulaire. Dans le sud-est du golfe de Gascogne et durant les deux campagnes de mesures, les résultats ont révélé un brassage très intense le long des côtes françaises et espagnoles, attribuable à la présence d'un fort jet côtier. De plus, cette étude a mis en évidence trois zones de passage particulièrement intense situées près

des côtes françaises et espagnoles, ainsi qu'au nord-ouest du domaine, structurant fortement le transport de matière particulaire dans la région.

J'ai également étudié l'impact du vent sur les structures de convergence des courants marins qui influencent le transport des déchets flottants dans le sud-est du golfe de Gascogne. Entre 1% et 2-3% de dérive due au vent, les structures de convergence des courants se déplacent dans la direction des courants et des vents. Avec un tel forçage dû au vent, les zones de passage intense simulées se trouvent généralement dans des régions similaires, indépendamment de la direction du vent. En revanche, au-delà de 3% de dérive due au vent, les taux d'échouage et de sortie du domaine augmentent de manière significative, et les structures de convergence ont tendance à s'agréger entre elles.

Ces résultats représentent une avancée significative dans l'identification de la dispersion et des structures de convergence des courants, avec des applications directes pour l'étude du transport et de l'agrégation des déchets marins. Cette connaissance facilite la recherche et l'atténuation des effets de la pollution marine, ce qui revêt une importance particulière dans une société dépendante des ressources côtières et soucieuse de préserver les écosystèmes marins. Effectivement, durant l'été, le golfe de Gascogne est missionné pour ramasser les déchets en mer afin de prévenir la pollution côtière (RUIZ et al., 2020).

Cependant, cette thèse démontre combien il est essentiel d'acquérir des observations pour comprendre la dynamique des structures de convergence à fine échelle, qui n'ont été que très peu documentées en Manche orientale et dans le sud-est du golfe de Gascogne. En effet, le manque d'observations systématiques est probablement la principale raison qui empêche une bonne compréhension de la dynamique océanique côtière dans ces deux régions.

Les observations lagrangiennes, notamment le déploiement de bouées dérivantes, offrent une valeur ajoutée significative. En effet, l'approche lagrangienne s'est révélée très efficace, comme cela a été démontré, pour acquérir une connaissance plus approfondie et réaliste de la dynamique côtière. Cette technique de mesure est peu coûteuse et facile à déployer. De plus, elle permet d'observer l'océan dans une zone prédéfinie et sur de plus ou moins longues périodes, dépendant fortement de la région d'étude. Il existe deux types d'instruments lagrangiens : les bouées dérivantes suivant l'écoulement dans la couche de surface (utilisées dans cette étude), et celles pouvant couler pour atteindre une profondeur spécifique et se déplacer à cette profondeur (D'ASARO et al., 2020; RYPINA et al., 2021). L'acquisition de ces deux types de données est utile pour plusieurs raisons. D'une part, déployées en clusters, elles peuvent être utilisées pour valider les modèles de circulation océanique ou des mesures radar HF. D'autre part, elles permettent de réaliser des études détaillées sur la dispersion turbulente à fine échelle, à différentes profondeurs, dans une région d'étude spécifique.

En l'absence d'observations lagrangiennes, ou en complément de celles-ci, les mesures radar HF apparaissent comme un outil de premier plan, indispensable pour l'étude de la turbulence marine, de la dispersion turbulente et de la dynamique à sous-mésoéchelle grâce à leur capacité à fournir une couverture spatiale étendue avec une résolution spatio-temporelle élevée. Ils permettent une surveillance continue et en temps réel des courants de surface, ce qui est crucial pour comprendre les dynamiques rapides et variables des zones côtières. Ces radars mesurent directement les vitesses des courants de surface, capturant ainsi des détails importants des structures à fine échelle, comme les tourbillons et les fronts jouant un rôle clé dans le transport de matière particulaire. Bien que ces systèmes soient opérationnels, il est envisageable de dévelop-

per différents indicateurs utiles pour l'aide à la prise de décision en cas de pollution par exemple.

L'efficacité des schémas d'assimilation de données, basés sur l'IO des mesures de vitesse des bouées dérivantes dans les modèles numériques de circulation, a été évaluée dans plusieurs études (e. g., BELYAEV et al., 2012; MOLCARD et al., 2003). D'autres techniques d'assimilation des courants de surface dans les modèles de circulation, telles que le filtre de Kalman à transformation d'ensemble local (SUN & PENNY, 2019), 4DVar (CARRIER et al., 2014; MUSCARELLA et al., 2015), le nudging et les méthodes séquentielles (ISERN-FONTANET et al., 2017), se sont également révélées efficaces. Cependant, l'approche proposée au chapitre 4, qui consiste à fusionner la vitesse des observations de radar HF et des bouées dérivantes, est nouvelle, facile à mettre en œuvre et nécessite un coût de calcul minimal. La méthode d'optimisation proposée dans cette étude peut aussi être appliquée dans d'autres zones côtières où des champs de vitesse de surface (provenant d'observations radar HF ou de modèles) et des mesures lagrangiennes sont disponibles. Cependant, certains ajustements peuvent être nécessaires, en particulier en ce qui concerne les méthodes d'obtention de la matrice de covariance utilisée dans la technique d'IO. Ces méthodes peuvent dépendre de la dynamique de la zone côtière ciblée et des processus physiques résolus par les champs de courant de surface disponibles. Ainsi, la procédure appliquée en Manche orientale est bien différente de celle utilisée dans le sud-est du golfe de Gascogne car les deux régions d'étude présentent des dynamiques distinctes.

En s'appuyant sur les champs de courant réalistes, cette thèse se focalise sur l'impact des courants marins et du vent sur les structures de convergence. Dans un premier temps, il a été démontré dans les chapitres 3 et 5 que le vent n'est pas négligeable lors de l'étude du transport de matière particulaire, et en particulier des déchets marins flottants. En Manche orientale en particulier, nous avons montré que la paramétrisation du vent dans le modèle MARS prend toute son importance. Effectivement, en zone côtière dans une région comportant de nombreuses falaises, les bouées dérivantes ou les particules fictives sont parfois protégées du vent. Ainsi, une méthode de correction de l'effet induit par le vent a été proposée pour reconstruire des trajectoires plus réalistes. Cela souligne l'importance d'une représentation précise du vent et de sa variabilité spatio-temporelle, ainsi que de son effet dans les modèles numériques de circulation côtière à haute résolution. Cependant, de nombreux autres phénomènes, non traités dans cette thèse, sont connus pour avoir un impact sur le transport de matière particulaire. Par exemple, l'effet de la dérive de Stokes sur les traceurs passifs, dérivant dans la couche de surface, devrait être pris en compte. La vitesse du courant induit par les vagues, estimée à 1% de la vitesse du vent (ARDHUIN & ACCENSI, 2012; ARDHUIN et al., 2018), peut modifier considérablement le transport des particules advectées passivement. DOBLER et al. (2019), van den BREMER et BREIVIK (2018), et CURCIC et al. (2016) ont également mis en évidence l'impact de la dérive de Stokes sur le comportement des traceurs passifs, des micro-plastiques ou des marées noires, notamment en cas de vents forts. Dans un second temps, la diffusion horizontale reste un phénomène à prendre en compte lors des simulations d'advection lagrangienne (KULLENBERG, 1972; SIEGEL et al., 1990), ou encore l'étude du transport océanique en utilisant une approche 3D.

Les processus de convergence et de divergence des courants sont étroitement liés aux mouvements verticaux et à la répartition de matière particulaire dans la colonne d'eau. En effet, les structures à sous-mésoéchelle sont associées à des gradients horizontaux de densité (zones frontales) avec une forte convergence en surface et une subsidence de l'eau. Ces structures peuvent facilement générer une répartition tridimensionnelle du matériel dérivant initialement en surface (D'ASARO et al., 2018; ESSINK, 2019). Aussi, comprendre ces processus de transport tridimensionnel est crucial car ils sont à l'origine du mélange vertical. Ces processus contrôlent

l'apport de nutriments nécessaires à la croissance des organismes marins, impactant le devenir de nombreuses grandeurs biogéochimiques (GOVE et al., 2019). D'autre part, des matières toxiques, telles que les hydrocarbures ou les déchets marins, déversées accidentellement en mer, peuvent être entraînées en profondeur par des mouvements descendants, amplifiant ainsi leur impact sur l'écosystème (CÓZAR et al., 2014; D'ASARO et al., 2018). Les déchets plastiques, en particulier, sont soumis au biofouling (accumulation de micro-organismes, de plantes, d'algues et d'animaux sur des structures immergées) et ont tendance à se fragmenter, augmentant ainsi leur présence dans la colonne d'eau (LOBELLE et al., 2021). Ainsi, l'aspect tridimensionnel des structures de convergence des courants côtier est une étude à approfondir. Tout en sachant que les diagnostics lagrangiens comme les FSLE et FDL, qui ont prouvé leur performance pour identifier les structures de convergence des courants en surface, sont aussi utilisables en trois dimensions (BETTENCOURT et al., 2013; TAANI et al., 2019).

En outre, dans le contexte climatique actuel, il serait intéressant d'étudier l'impact du changement climatique sur la variabilité saisonnière et la localisation des structures de convergence à fine échelle. Le dérèglement climatique, par l'augmentation des températures globales et les perturbations des cycles hydrologiques, modifie l'hydrographie et les courants marins qui jouent un rôle clé dans la régulation du climat terrestre. Son impact à grande échelle est plus ou moins connu mais son impact sur les structures de convergence reste à investiguer car les structures de convergence, où les courants se rencontrent et interagissent, peuvent être sensibles à ces changements. Comprendre l'effet du changement climatique sur les structures de convergence des courants marins pourrait aider à mieux prévoir les conséquences écologiques et socio-économiques, et contribuer à l'élaboration de stratégies d'adaptation et de mitigation. De plus, le changement climatique provoque l'augmentation en fréquence et en intensité des événements extrêmes. Une étude dans le golfe du Tonkin, au Vietnam, a démontré que les flux de matière particulaire sont parfois inversés lors du passage de cyclones (PITON et al., 2021). Ainsi, bien que le sud-est du golfe de Gascogne et la Manche orientale ne soient pas des régions sujettes aux cyclones, il serait pertinent d'étudier l'effet des tempêtes et vagues de chaleurs marines sur les structures de convergence des courants côtiers.

En outre, l'amélioration des techniques d'observations de l'océan, et en particulier le lancement du satellite SWOT en décembre 2022, ainsi que l'utilisation croissantes des techniques d'intelligence artificielle, est de bonne augure pour améliorer nos connaissances sur l'hydrodynamique complexe de l'océan côtier. De plus, un nouveau capteur, conçu comme une preuve de concept, a été développé pour détecter systématiquement les déchets marins sur une large gamme d'échelles spatio-temporelles (CÓZAR et al., 2024). Ce capteur cible les agrégats de débris flottants de la taille d'un mètre générés par les lignes de convergence de la surface de la mer, ces agrégats ayant été signalés comme une cible fiable pour la détection par satellite. Ces observations sont indispensables pour de nombreuses applications océanographiques, telles que la gestion des ressources côtières, la sécurité maritime, et la prévention des risques environnementaux, dans le but de renforcer la durabilité des écosystèmes côtiers.

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# Valorisations du travail

## *Publications scientifiques*

**P1** - S. Bertin, A. Sentchev, E. Alekseenko - 2024. Fusion of Lagrangian drifter data and numerical model outputs for improved assessment of turbulent dispersion. Ocean Science.  
*doi* : 10.5194/os-20-965-2024

**P2** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, I. Ruiz, A. Orfila, A. Sentchev - 2024. Coastal current convergence structures in the Bay of Biscay from optimized High Frequency radar and satellite data. Science of the Total Environment.  
*doi* : 10.1016/j.scitotenv.2024.174372

**P3** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, I. Ruiz, A. Orfila, A. Sentchev - 2024. Coastal current convergence structures in the Bay of Biscay from optimized High Frequency radar and satellite data. Paralia. [https://www.paralia.fr/jngcgc/18\\_02\\_bertin.pdf](https://www.paralia.fr/jngcgc/18_02_bertin.pdf)

## *Publications grand public*

**G1** - S. Bertin - 2024. Une nouvelle méthode pour localiser les déchets en mer. The conversation. <https://theconversation.com/une-nouvelle-methode-pour-localiser-les-dechets-en-mer-223896>

## *Communications orales*

**C1** - A. Rubio, A. Sentchev, S. Bertin, O.C. Basurko, I. Ruiz, I. Hernandez-Carrasco, I. Lizarán, B. Casas, A. Orfila. Lagrangian approach for studying marine litter aggregations in the coastal area of the South-East Bay of Biscay. EOF, Expanding Ocean Frontiers Conference, Las Palmas de Gran Canaria (Espagne), juillet 2022.

**C2** - S. Bertin, A. Sentchev, E. Alekseenko. Towards a new decision support tool for marine litter monitoring in the eastern English Channel. SeaTech week, Brest (France), octobre 2022.

**C3** - A. Rubio, I. Hernandez-Carrasco, S. Bertin, I. Ruiz, A. Sentchev, O.C. Baskuro, A. Orfila. Lagrangian transport of MARine litter from modeling, analysis, and observations in coAstal waters of the Bay of Biscay. ASLO Aquatic Sciences Meeting, Palma de Majorque (Espagne), juin 2023.

**C4** - S. Bertin, Z. Hubert, E. Alekseenko, A. Sentchev, S. Monchy, F. Artigas. Optimization

of a coastal ocean model for the study of the influence of turbulent transport and dispersion processes in the Eastern English Channel on the distribution and dynamics of phytoplankton communities. EGYMANCHE, Wimereux (France), juillet 2023.

**C5** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, A. Orfila, A. Sentchev. Impact of coastal Current Convergence Structures and windage in the spatial distribution of floating marine litter in the southeastern Bay of Biscay. ISOBAY conference, La Rochelle (France), juin 2024.

**C6** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, I. Ruiz, A. Orfila, A. Sentchev. Identification des zones de convergence par le calcul de métriques lagrangiennes à partir de champs de courant de surface issus de modèles tridimensionnels et radar HF : Application à l'étude du transport des déchets marins dans le golfe de Gascogne. JNGCGC conférence, Anglet (France), juin 2024.

**C7** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, I. Ruiz, A. Orfila, A. Sentchev. Operational observations of coastal current convergence structures driving floating marine litter aggregation in the southeastern Bay of Biscay. EOF conference, Valence (Espagne), juillet 2024.

**C8** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, I. Ruiz, A. Orfila, A. Sentchev. Identifying coastal Current Convergence Structures in the southeastern Bay of Biscay by a combination of HF radar and Lagrangian measurements of surface current velocities : application to Floating Marine Litter. International Radiowave Oceanography Workshop, ROW, Plymouth, (Royaume-Uni), septembre 2024.

**C9** - G. Lopez, S. Bertin, E. Reyes, A. Rubio, A. Sentchev, A. Barth. An evaluation of gap-filling methodologies for HF radar. International Radiowave Oceanography Workshop, ROW, Plymouth, (Royaume-Uni), septembre 2024.

### *Communications par affiche*

**A2** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, A. Orfila, A. Sentchev. Improvement of a high-resolution oceanic circulation model using Optimal Interpolation of Lagrangian drifters in the Southeast Bay of Biscay for assessing the turbulent dispersion. 54<sup>th</sup> International Liege Colloquium of Ocean Dynamics, Liège (Belgique), mai 2023.

**A2** - S. Bertin, A. Rubio, I. Hernandez-Carrasco, L. Solabarrieta, A. Orfila, A. Sentchev. Improving the frontal zone detection in the southern Bay of Biscay by a combination of remote sensing and Lagrangian measurements of surface current velocities. Ocean Science Meeting, New Orleans (Etats-Unis), février 2024.

### ***Autres***

**X1** - Participation à l'organisation de campagnes de mesures en Manche orientale et dans le sud-est du golfe de Gascogne.

**X2** - Participation à l'encadrement d'un stage de M1 au LOG intitulé "Utilisation de bouées lagrangiennes pour améliorer un modèle de circulation".

**X3** - Enseignement d'un UO de 20h d'introduction à l'océanographie physique et d'un UE de 10h de physique à l'ULCO.

**X4** - Mise en place d'une formation à l'utilisation des logiciels OpenDrift et OceanParcels à l'AZTI.

**X5** - Mise en place du code 2dVar sur le GitHub du projet national espagnol LAMARCA.

**X6** - Communication sur les différents projets et publications sur divers réseaux sociaux professionnels.

## Résumé en français

Les processus de transport et de dispersion dans l'océan sont cruciaux car ils déterminent le devenir des quantités biologiques et chimiques dérivant avec les courants océaniques. Les zones côtières, situées entre le continent et l'océan ouvert, subissent de nombreuses pressions anthropiques causant pollution, prolifération d'algues toxiques et d'espèces invasives. Comprendre les processus de transport et de dispersion de matière particulaire dans ces zones est essentiel pour préserver ces environnements et permettre leur gestion durable. Les modèles numériques de circulation et les systèmes d'observations sont capables de quantifier avec précision l'écoulement turbulent et étudier les processus de dispersion. Mais ils sont limités en raison de la complexité de l'environnement océanique côtier et le manque de résolution spatio-temporelle. Cette thèse vise à améliorer la représentation et la caractérisation réaliste des structures de convergence des courants de surface en recourant à une méthode innovante de fusion de données hétérogènes. Cette méthode consiste à interpoler de manière optimale dans le temps et l'espace les vitesses de courants de surface issues du déploiement de bouées dérivantes afin de contraindre différents champs de courants de surface réguliers (modélisés, observés par radar haute-fréquence). Dans l'objectif de tester la performance de cette technique, elle est appliquée dans deux régions d'étude bien différentes - la Manche orientale et le sud-est du golfe de Gascogne. L'optimisation proposée améliore significativement la reconstruction des champs de courants de surface, réduisant de 50% environ l'erreur lagrangienne, renforçant ainsi la compréhension des processus de transport et de dispersion de matière particulaire. Le calcul de différentes quantités lagrangiennes telles que la dispersion, les exposants de Lyapunov et la divergence lagrangienne permet d'identifier les zones de forte turbulence et de cartographier la distribution spatiale des structures de convergence des courants marins. Aussi, grâce à l'optimisation, l'identification des structures de convergence des courants marins est plus précise, ce qui permet de mieux comprendre le transport de matière particulaire à la surface des zones côtières.

**Mots-clés :** Dispersion turbulente, Structures de convergence des courants côtiers, Diagnostics lagrangiens, Manche orientale, Sud-est du golfe de Gascogne

## Abstract in english

Transport and dispersion processes in the ocean are crucial, as they determine the fate of biological and chemical quantities drifting with ocean currents. Coastal zones, located between the continent and the open ocean, are subject to numerous anthropogenic pressures causing pollution, toxic algal blooms and invasive species. Understanding the transport and dispersion processes of particulate matter in these areas is essential to preserve these environments and enable their sustainable management. Numerical circulation models and observation systems have a significant potential to study dispersion processes. However, they are limited by the complexity of the coastal ocean environment and by the lack of spatio-temporal resolution. This thesis aims to contribute to the improvement of the realistic representation and characterization of coastal current convergence structures using fusion of heterogeneous data. This method consists to optimally interpolate in space and time surface current velocities derived from Lagrangian drifter deployments to constrain different regular surface current fields (modeled, observed by high-frequency radar). In order to test the capabilities of this technique, it is applied in two very different study regions - the eastern English Channel and the southeastern Bay of Biscay. The proposed optimization significantly improves the reconstruction of surface current fields, reducing by 50% the Lagrangian error, thus enhancing our understanding of particulate matter transport and dispersion processes. By calculating various Lagrangian quantities such as dispersion, Lyapunov exponents and Lagrangian divergence, it is possible to identify areas of high turbulence and map the spatial distribution of coastal current convergence structures. Hence, thanks to optimization, the identification of coastal current convergence structures is more precise, enabling a better understanding of the transport of particulate matter at the surface of coastal zones.

**Key words :** Turbulent dispersion, Coastal Current Convergence Structures, Lagrangian diagnostics, eastern English Channel, southeastern Bay of Biscay