Pattern formation at the ocean surface:

*Sargassum* distribution and the role of the eddy field

Yisen Zhong\(^1\), Annalisa Bracco\(^1\), Tracy A. Villareal\(^2\)

\(^1\)School of Earth and Atmosphere Sciences, Georgia Institute of Technology, 311 Ferst Dirve, Atlanta, Georgia 30332, USA

\(^2\)Marine Science Institute, The University of Texas at Austin, 750 Channel View Drive, Port Aransas, Texas 78373, USA

Correspondence to Yisen Zhong yzhong31@gatech.edu
Abstract

Positively buoyant organisms such as the macroalgae *Sargassum* and the cyanobacterium *Trichodesmium* often form surface accumulations visible in satellite imagery that have lateral scale separation of tens of kilometers and cannot be explained by Langmuir circulation. Here we discuss the accumulation of floating materials in the ocean in presence of meso- and submesoscale activity. Using high resolution simulations of the ocean mesoscale in both idealized (a 3-D box where coherent eddies are forced by small-scale winds) and realistic domains (western Gulf of Mexico), where extensive concentrations of floating *Sargassum* have been recorded in satellite images, we show that the distribution of tracers at ocean surface departs rapidly from the one observed few tens of meters below it. Such distribution does not resemble that observed for passive tracers in quasi-geostrophic turbulence.

The strong divergence and convergence zones generated at the surface by ageostrophic processes in the submesoscale range are responsible for the creation of areas where the floating material accumulates. Floating particles are expelled from the core of mesoscale eddies, and concentrate in convergence regions in patterns comparable to the ones observed through the satellite images. In light of those results, *Sargassum* and/or *Trichodesmium* may provide a useful proxy to track convergence/divergence processes resulting from ageostrophic processes at the ocean surface.

Keywords: Lagrangian transport, floating material, Gulf of Mexico, mesoscale circulation
Introduction

[1] In the ocean, a statistical description of the transport pathways usually requires Lagrangian observations in the form of drifter or float trajectories. Biological tracers can also help in the understanding of the physical processes at play and to this end the distribution of free-floating *Sargassum* (Phaeophyceae) has captured the interest of physical oceanographers and marine biologists for decades. *Sargassum* is a highly visible macrophytic algae widely distributed in the North Atlantic, Caribbean, and Gulf of Mexico that provides a crucial habitat for a myriad of marine organisms (Butler et al. 1983; Rooker et al. 2004). In 1923, Winge first reported that *Sargassum* algae always align themselves with local wind direction to form an irregular pattern of fractional parallel lines (aka windrows). Those streaks may extend to a few hundred meters in length with typical separations between wind-wave lines of 20 m to 50 m comparable to the boundary-layer depth (Faller and Woodcock 1964). Langmuir systematically investigated the lines of *Sargassum* through a series of experiments on Lake George and concluded that they were, in fact, dictated by a quasi-steady circulation, afterwards named Langmuir circulation (Langmuir 1938).

[2] More recently, extensive ‘lines’ of *Sargassum* of much larger magnitude have been detected in ocean color satellite images of Gulf of Mexico and the West Atlantic Ocean (Gower et al. 2006; Gower and King 2008). The Gulf of Mexico, together with the Sargasso Sea, has the largest concentration of floating *Sargassum* biomass in the world (Gower and King 2011). The satellite observations were the first to map the patchiness of the *Sargassum* at the ocean mesoscales, and note widespread lines, typically 10-100 kilometers long and kilometers wide (Fig. 1). The range of scales involved suggests that physical structures like
eddies, filaments and fronts are likely responsible for this phenomenon and may control the
dispersion and aggregation of Sargassum, as well as other floating material, at the ocean
surface. Similar patterns are also evident in reflectance images of the oceanic surface and have
been linked to the ageostrophic intensification of cold filaments by the horizontal shear
associated with mesoscale eddies (McWilliams et al. 2009a).

[3] In the ocean, coherent vortices or eddies form spontaneously by baroclinic or
barotropic instability under the influence of rapid rotation and stable stratification, or by
interaction of the main currents with topographic features. These vortices/eddies move in a
chaotic manner while embedded in a background turbulence field. Here ‘coherent’ refers to
the vortices’ lifetime, which is much longer than their turnover time. In the last three decades,
the transport and mixing properties associated with coherent vortices have been studied in
detail in the context of two-dimensional (2-D) (McWilliams 1990; Elhmaidi et al. 1993;
Provenzale 1999) and quasi-geostrophic (QG) (Smith and Vallis 2001; Bracco et al. 2004)
turbulence flows in both Eulerian and Lagrangian frameworks. Coherent vortices are
impermeable to inward and outward fluxes and can retain their physical and chemical
properties over their lifetime. They have been described as “islands of regular Lagrangian
dynamics in a chaotic background” (Provenzale 1999) and different dispersion behaviors have
been identified depending on where Lagrangian tracers reside – inside, outside or within the
circulation cells around the eddies. According to 2-D and QG theories, however, the
distribution of tracers (i.e., temperature, salinity, chemical or biological concentrations) is
uniform within each topological region (vortices, circulation cells or background turbulence)
and no aggregation or accumulation can take place. In other words, the mesoscale vortices
represent the only level of patchiness present in the Eulerian flow, while the distributions of Lagrangian tracers remain homogeneous. In agreement with those findings, in the ocean it has been observed that coherent eddies can maintain the characteristics of their source waters for several months and over hundreds of kilometers (Richardson 1993; Mcdonagh and Heywood 1999). They can trap floats (Richardson 1993; Paillet 1999; Shoosmith et al. 2005), and they are responsible for anomalous dispersion curves (Rupolo et al. 1996; Berloff et al. 2002; Reynolds 2002), Lagrangian power spectra (Rupolo et al. 1996) and horizontal velocity distributions (Bracco et al. 2000; Bracco et al. 2003) in agreement with 2-D and QG turbulence predictions. Those properties, all assessed below the ocean surface, have been confirmed in a recent study using a regional ocean model at 1 km horizontal resolution with an idealized configuration mimicking a baroclinic wind-induced vortex system (Koszalka et al. 2009). The same study, however, pointed out that the motion at the submesoscale (~ 1 km scale) within and immediately around eddies is strongly ageostrophic, in contradiction with the geostrophically balanced 2-D or QG approximations. This result contributed to the view of a surface ocean dominated by strong and complex vorticity patterns generated by ageostrophic processes, already established for frontal systems (Capet et al. 2008a,b).

[4] Fronts, i.e., narrow regions of very large lateral buoyancy gradients, are also ubiquitous in the upper ocean (Ullman and Cornillon 1999; Capet et al. 2008b). Fronts are characterized by strong lateral velocity shear and relative vorticity of the opposite sign on either side. Strong frontogenesis can occur also around the edge of the mesoscale eddies due to the high strain rate between the rapidly spinning eddy and background flow. The large vorticity gradients at front edges suggest that frontal motion is characterized by a loss of
geostrophic balance (Capet et al. 2008a,b,c). It has been shown recently that frontal instability acting at the submesoscales plays a crucial role in transferring energy down to the dissipation scale (McWilliams et al. 2003; Molemaker et al. 2005; Capet et al. 2008c). Additionally, large vertical velocities (substantially larger than mesoscale counterpart) develop inside the fronts (Thomas and Ferrari 2008), and they are instrumental in exchanging physical, chemical and biological properties between the ocean surface and the interior (Mahadevan et al. 2008).

[5] In this study, we discuss how ageostrophic submesoscale processes associated with eddies and fronts determine the distribution of tracers at the ocean surface and can explain the complex patterns revealed by the satellite observations of Sargassum. We will present two numerical integrations: The first representing an idealized wind-driven eddy-dominated portion of the open ocean, which allows us to investigate the physical mechanism at play, and the second covering the Gulf of Mexico, where we can verify the agreement with the satellite observations.

**Methods**

[6] In this study, we implemented two ocean model configurations. The first configuration simulated an idealized eddy-dominated flow away from coastal boundaries, similar to that described in Koszalka et al. (2009) except for a deeper mixed layer. Its circulation was forced only by a wind field described by an analytical expression. This idealized configuration allowed us to investigate in detail the dynamical mechanism that controls the distribution of the surface particles. The second one simulated the circulation in the western Gulf of Mexico, where extensive lines of Sargassum have been observed using satellite measurements (Gower et al. 2006), and allowed us to validate our conclusions in a realistic domain. Both
configurations include the integration of Lagrangian trajectories to mimic the behavior of passively-advected material at the ocean surface.

[7] The Regional Ocean Modeling System (ROMS) (Shchepetkin and McWilliams 2003; Shchepetkin and McWilliams 2005) is a free-surface eddy-resolving primitive equation model based on the Boussinesq approximation and hydrostatic balance. It is discretized using an Arakawa C grid with orthogonal curvilinear coordinates on the horizontal and has a generalized terrain-following coordinate system (S- or sigma coordinate) with a staggered Arakawa C grid in the vertical. The depth of each layer depends therefore on the local topography and sea surface height (SSH). ROMS has a split-explicit time step scheme, which implements a short time step for surface elevation and barotropic momentum, and a longer one for baroclinic momentum and tracer variables (including temperature and salinity). Horizontal 3rd-order upstream-biased and vertical 4th-order centered difference schemes were used to advect both momentum and tracer equations. Given the high resolution of our runs, we selected horizontal biharmonic friction and diffusion to reduce small-scale noise. The vertical mixing was parameterized accordingly to the non-local K-Profile Parameterization (KPP) boundary-layer scheme (Large et al. 1994). ROMS has been successfully used to simulate the ocean circulation in regions characterized by high mesoscale activity as, for example, the California Current System (Di Lorenzo et al. 2004; Capet et al. 2008a), the Caribbean Seas (Jouanno et al. 2008) and the Labrador Sea (Luo et al. 2011).

[8] For the idealized configuration, we adopted a periodic domain with lateral size 256 km and a constant depth of 1000 m on the \( f \)-plane \( (f \) has constant value \( 10^{-4} \text{s}^{-1} \)). The horizontal resolution of the grid was 1 km and the vertical discretization was 30 layers, 10 of
which confined to the first 110 m. The depth of each layer was approximately constant because the topography is flat and the SSH variations are small compared to the total water depth. The top layer covered the first 1.93 m of the water column. An idealized wind forcing field with a narrow-band continuous spectrum centered at about 40 km with radial wave number \( \sqrt{k_x^2 + k_y^2} \approx 8 \) provided a close approximation to that used in previous studies to force 2-D and QG turbulence fields (Elhmaidi et al. 1993; Bracco et al. 2004). Different realizations of the wind stress field with identical power spectra were linearly interpolated every 20 days. We initialized the model from rest by imposing a vertical stratification that follows sample profiles in the extra-tropics (Conkright et al. 2002) below the top 50 m and is uniform within the surface mixed layer. The solution was nudged to the initial vertical profile of temperature and salinity every 60 days to forestall flow barotropization and energy accumulation.

[9] Passive Lagrangian particles were deployed in the ocean top layer at about 1 m depth as approximation of the surface, and at 106 m, once the stationary state was achieved (in the model the horizontal velocity field used to advect the Lagrangian tracers is available at the half-sigma levels). Their advection in time is described by

\[
\frac{dx}{dt} = u(x,t)
\]

here \( x \) is the particle position vector and \( u \) is the corresponding Eulerian velocity vector at the particle position. The time integration of the above equation was performed on-line using a 3\textsuperscript{rd}-order Milne predictor and 4\textsuperscript{th}-order Hamming corrector scheme. Off-line Lagrangian diagnostics of upper ocean mixing in presence of an energized submesoscale field has been
shown to lead to significant temporal sampling errors (Keating et al. 2011). No inertial effects were included and the particles moved with flow passively and did not feedback on the surrounding fluid.

[10] ROMS includes routines for the integration of 3-D, isobaric and geopotential Lagrangian tracers. Particles move freely in the 3-D flow or are confined along the isobars and isobaths, respectively. We were interested in understanding the motion of surface floating tracers (regardless of source or sink) and a natural choice is to use isobaric particles. Model results, however, were independent on the kind of tracers adopted because at the ocean surface the vertical motion of neutrally buoyant and infinitesimally small particles is controlled uniquely by the surface boundary condition \( \frac{d\eta}{dt} = w \), where \( \eta \) is free surface elevation and \( w \) is vertical velocity. Neutrally buoyant particles deployed at the surface will keep following the surface isobaric plane. In our simulations, we deployed the Lagrangian tracers on a uniform grid in a shallow layer with the purpose of reducing the potential numerical error at the surface. In the idealized run, 65,536 particles were integrated starting 488 days after initialization and their trajectories were followed for 60 days.

[11] To validate the analysis of the idealized case, we also configured ROMS in the Gulf of Mexico and focused a high resolution run in the west part of the domain, where satellite images of Sargassum distributions are available (Gower et al. 2006). For this integration, we first spun-up the model at 5 km horizontal resolution and 30 vertical layers over the whole Gulf of Mexico. The atmospheric forcing fields were from NCEP/NCAR reanalysis (Kalnay et al. 1996) and included wind stress, surface net heat fluxes and water fluxes, and solar radiation fluxes. Climatological values over the period 1948-2001 were used. SODA (Simple
Ocean Data Assimilation, Carton and Giese 2008) reanalysis provided the nudging conditions at the open boundaries and the initial conditions. Again, monthly climatological values were employed. After ten years, the model output reached a statistically stationary state that was used as initial condition for a second experiment in which surface atmospheric forcings retain monthly variability over the period of 2002-2007. NCEP momentum fluxes were replaced by the monthly-averaged NCEP-QSCAT blended wind stresses from Colorado Research Associates (version 5.0) (Chin et al. 1998; Milliff et al. 2004). Finally, we concentrated on the western Gulf of Mexico with a nested domain at 1 km horizontal resolution. Nesting is a widely-used technique to run very high resolution simulations over limited areas, reducing the computational costs. It embeds a child grid of finer resolution into its parent coarse grid so as to be able to resolve the details of the mesoscale and submesoscale dynamical features over the region of interest. Here we adopted an offline one-way nesting method that has been recently developed by Mason et al. (2010). The child grid run was conducted after the parent grid integration (off-line); the parent grid conveyed initial and boundaries conditions to the child grid, but no information was passed back from the child to the parent grid (one-way). The forcing fields were analogous to the ones in the parent run except for using daily-varied data instead of monthly. The nested run was performed from April 1 to May 31, 2005. A set of 75,398 Lagrangian particles were uniformly deployed in the child grid in the surface layer on April 26 and their trajectories tracked for one month (the top layer in this case had variable depth due to the realistic bottom topography, but always less than 1 m and 0.2 m on average).

**Results**

*The idealized set-up*
The circulation in the idealized experiment is mainly horizontal and is characterized by a series of wind-induced coherent vortices embedded in the background turbulent field. Figure 2 shows a snapshot of vertical relative vorticity $\zeta$ scaled by planetary vorticity (i.e., by the Coriolis parameter $f$) in the surface and tenth layers, where particles are released, 60 days after their deployment. Relative vorticity, defined as $\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$, measures the local rotation rate in the horizontal plane. Positive values correspond to anticlockwise motion (cyclonic) and negative to clockwise (anticyclonic). The flow field has a pronounced dominance of long-lived anticyclonic vortices. At the surface, submesoscale vorticity filaments are also visible around the eddies, while they are absent at depth. The asymmetry between anticyclonic and cyclonic vortices results from the straining field exerted by vortex Rossby waves (VRWs, Montgomery and Kallenbach 1997), which induces a greater weakening of cyclones whenever the eddies have size comparable to the Rossby deformation radius, as in our simulation (Montgomery and Kallenbach 1997; Mcwilliams et al. 2003; Graves et al. 2006). The average horizontal diameter of mature vortices is about 20 km, the average eddy turnover time is 3.2 days, and the Lagrangian time scale, i.e. a measure of the persistence of the fluid motion or the time required for a water parcel to be displaced from one location to another by the eddy field, is of 2 days at the surface and 3 days around 106 m, in agreement with oceanographic observations in regions with an energetic mesoscale (Lumpkin et al. 2002). In the vertical, the modeled eddies have a bowl-shaped vorticity signature extending to over 100 m and reduced intensity. The ratio $\zeta/f$ provides an estimate of the local Rossby number. In the vortex cores, the absolute value of $\zeta$ is about 1.5 times larger than $f$ at the surface and only half the value of $f$ at 106 m depth (Fig. 2), implying that the surface
motion within the eddies is strongly ageostrophic, while circulation at depth is closer to geostrophic balance also inside the vortices. In contrast to anticyclones, cyclonic vortices in this idealized run typically have a smaller radius, are confined to the mixed layer and are stretched out into filaments more rapidly.

[13] To study the transport properties of this vortex-dominated flow, we deployed two sets of Lagrangian particles at different depths. The first set was released at about 1 m and represents a proxy of surface floating materials, whereas the second one released at 106 m, where the intensity of the vorticity field within the eddies is reduced to a third of the value at the surface. Each set contains 65,536 particles, initially sown uniformly over the entire domain. To highlight the role of vortices on particle distribution, we released a second set of 32,761 particles homogeneously distributed inside one vortex, both at the surface and at depth. The edge of the vortex was defined as the closed contour line surrounding its core where the Okubo-Weiss parameter ($OW$) (Okubo 1970; Weiss 1991) is zero. $OW$ measures the relative strength of local deformation and rotation. In 2-D and QG turbulence, this is defined as $OW = s^2 - \zeta^2$, where $s$ measures the horizontal deformation due to stretching and shearing ($s^2 = (\partial u / \partial x - \partial v / \partial y)^2 + (\partial u / \partial y + \partial v / \partial x)^2$), and $\zeta$ is relative vorticity as mentioned above. In 3-D turbulence, as in our case, the horizontal components of the velocity tensors are at least one order of magnitude larger than the vertical component ($10^{-4}$ vs. $10^{-5}$), and the 2-D definition of $OW$ still provides a clear horizontal separation between vortex-dominated and strain-dominated regions.

**Mechanism of pattern formation**

[14] Figure 2 shows, in the bottom panels, the particle distributions after 60 days at the
surface and deep levels. Tracers at the surface are no longer uniformly distributed but organized in *lines* and *rings*, with regions of very high and low concentrations, closely resembling the patterns seen in the satellite images of *Sargassum* (Fig. 1). Particles aggregated and organized themselves soon after they were released (in less than two days). This behavior differs significantly from the one of Lagrangian tracers in 2-D or QG turbulence. Particles advected at depth, on the other hand, still maintain an almost uniform distribution, as in 2-D flows.

[15] To understand the physical mechanism behind the different evolution of the particle trajectories, the velocity divergence/convergence fields should be considered. For an incompressible fluid, the continuity equation ensures that

\[
\text{div}(\mathbf{v}) = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0
\]

where \(u\), \(v\) and \(w\) are respectively the zonal, meridional and vertical component of the velocity \(\mathbf{v}\) field. In 2-D or QG flows, the vertical velocity is null or constant, and the horizontal velocity divergence \((\text{div}_H = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y})\) is therefore zero. A tracer initially distributed uniformly will maintain its uniformity with time. In 3-D, on the other hand, the horizontal divergence/convergence field is determined by the vertical gradient of \(w\). If such a gradient is small, i.e., in presence of vertical velocities that do not vary significantly with depth, then the particle behavior resembles the one of QG flows. This situation is recovered in our idealized simulations at any depth below 10 m from the surface, and follows from the ocean being a rapidly rotating, stably stratified flow. However, close to the surface, the velocity gradient is large. Vertical velocities are close to zero at the surface, while they are high immediately
below the surface due to the presence of Ekman pumping, wind-driven vortices, submesoscale fronts, and vorticity filaments. Koszalka et al. (2009) introduced a diagnostic equation for \( w \) based on the horizontal flow divergence and analyzed all contributions. Using such formulation, it is possible to identify the terms contributing to the vertical velocity at and below the ocean surface. At the surface, the only term which is non-zero is the free-surface (\( w = d\eta/dt; \) here \( \eta \) is free surface). Immediately below the ageostrophic term

\[
-\int_a^b \alpha_1 \left[ \frac{\partial \zeta_1}{\partial t} + u \frac{\partial \zeta_1}{\partial x} + v \frac{\partial \zeta_1}{\partial y} + w \frac{\partial \zeta_1}{\partial z} \right] dz - \int_a^b \alpha_2 \left[ \frac{\partial \zeta_2}{\partial t} + u \frac{\partial \zeta_2}{\partial x} + v \frac{\partial \zeta_2}{\partial y} + w \frac{\partial \zeta_2}{\partial z} \right] dz,
\]

where \( \zeta_1 = \partial v / \partial x, \zeta_2 = - \partial u / \partial y, \alpha_1 = (f + \zeta_1)^{-1}, \alpha_2 = (f + \zeta_2)^{-1} \), stretching

\[
-\int_a^b \alpha_1 \left[ \chi_1 \zeta_1 \right] dz - \int_a^b \alpha_2 \left[ \chi_2 \zeta_2 \right] dz,
\]

where \( \chi_1 = \partial u / \partial x, \chi_2 = \partial v / \partial y \) and tilting

\[
-\int_a^b \alpha_1 \left[ \frac{\partial w \partial v}{\partial x \partial z} \right] dz + \int_a^b \alpha_2 \left[ \frac{\partial w \partial u}{\partial y \partial z} \right] dz
\]

dominate the vertical velocity field and all display their maxima values, creating a very strong gradient.

[16] Figure 3 shows the particle distribution overlying the local horizontal velocity divergence/convergence field at the surface with magnification of vortex \( V \), whereas Figure 5 focuses on the tracers at 106 m. The horizontal velocity divergence/convergence balances the vertical velocity gradient and quantifies the rate of local fluid particle dispersion/accumulation; if negative, it represents convergence, while positive value indicates divergence. At the surface, most particles aggregate where convergence occurs (blue regions), while the divergence zones (in red) are left with a smaller number of tracers or almost vacant. Areas of strong divergence/convergence are found in and around the most intense vortices and submesoscale fronts, but do not correlate directly with the vorticity field. The divergence field, likewise \( w \) (Koszalka et al. 2009), is very complex, and results from the interplay of advection,
stretching, and instantaneous vertical velocity changes. At the surface the horizontal convergence field is so strong that surface particles show aggregation patterns about a day after deployment and tend to form circular rings at the vortices peripheries. The aggregation time scale at the surface is therefore shorter that the mean Lagrangian time scale (2 days) at which particles are advected by the eddy field. As a result, more particles are found in the blue regions at all times (Fig. 4). Differences in the distributions shown in Figure 4 are related to the intrinsic variability of the eddy field given that eddies and vorticity filaments are continuously formed and destroyed in the idealized run in a chaotic manner. During the 60 days during which the Lagrangian tracers are advected around, the total area occupied by eddies and strong filaments varies between 11% and 15.5%, and it is 12.5% on average. At 106 m, the intensity of the convergence/divergence field is reduced to ¼ or less of the surface values (the absolute value of the horizontal convergence averaged over the whole domain is five to six times smaller at 106 m than at the surface at all times), and the time required for aggregation is longer that the Lagrangian time scale (Fig. 5). The horizontal convergence field at depth is therefore insufficient to drive particle aggregation, except for very few areas corresponding to the cores of the most intense and deep-reaching vortices. At depth, for the first few weeks after the tracer deployment it is noticeable the twisting motion transmitted by the eddies to the particles in their cores, due to the smaller horizontal velocities of the background flow around the vortices. This may cause the impression of particle aggregation for time scales shorter than the advective time of the background flow. As shown in Figure 6, particles are equally distributed in (weakly) positive and negative convergence regions at all times.
The idealized integration provides a theoretical framework to investigate particle distributions in a wind-driven, eddy-dominated ocean. It suggests that at the surface floating tracers aggregate following the horizontal divergence field which is unusually strong due to the large gradients in the vertical velocity field due to submesoscale ageostrophic processes in correspondence to vortices, vorticity filaments, submesoscale fronts.

Figure 7 shows a snapshot of the surface relative vorticity field in the western Gulf of Mexico (scaled by Coriolis parameter $f$, counterpart of Fig. 2) on May 1, 2005 in the nested area where the horizontal resolution is 1 km. The run reproduces extremely well the sea surface height variability observed in satellite data and vorticity values are realistic. The vorticity field displays more complicated but overall slightly weaker patterns than in the idealized integration. Many eddies of different sizes as well as numerous filaments and submesoscale fronts can be identified in this vorticity map. Four areas with pronounced mesoscale/submesoscale features are labeled with $A$, $B$, $C$ and $D$ and they represent all the different patterns that can be found in the chosen domain. The three largest eddies seen in our domain ($B$, $C$ and $D$) are originated from the Loop Current that enters the Gulf of Mexico between Cuba and the Yucatán peninsula, loops west and south and then exits through the Florida Straits. Its instability generates eddies at irregular intervals of 9-14 months that have typical diameters of 150-300 km and extends up to 1000 m in depth (Vukovich 1995; Sturges and Kenyon 2008). Newly formed eddies move westwards with a speed of about 2-5 km/day and can exist for months to years until they decay through the interaction with the continental shelf (Elliott 1982; Vukovich and Crissman 1986). When the Loop Eddies reach the Mexican
continental shelf, they generally break up into smaller eddies generating dipole or, more rarely, tripole structures. Numerous small coherent vortices, generated independently of the Loop Current, are also resolved in our nested experiment. Their formation is by baroclinic instability associated with lateral boundary friction and bottom topography (see, for example, the train of eddies along the continental shelf on the west of the domain), or by baroclinic instability of the density fronts surrounding the Loop Eddies (see region A in Fig. 7). The intensity of both the vorticity and horizontal divergence/convergence fields varies within the different dynamical structures. The two regions with highest values are found in correspondence of A and B. A includes submesoscale eddies and filaments, whereas B is an anticyclonic eddy with diameter of approximately 200 km. The remaining two Loop Eddies (C and D) show a more homogeneous structure at their interior and lower intensity in the divergence/convergence field.

[19] To investigate the transport properties of this flow field, we deployed a set of 75,398 passive Lagrangian particles in the surface model layer over the whole domain on April 26, 2005. The tracers are initially distributed homogeneously. The distance between two neighboring particles is 30 km. Figure 8 displays the particle distribution five days after deployment superimposed to the convergence field. As in the idealized case, particles rapidly aggregate to form patterns resembling curls and lines. Curls are particularly evident in regions A and B and are associated with comparatively strong vorticity and convergence/divergence fields. This kind of pattern is also captured by the satellite image shown in the right panel of Figure 1. Particle aggregation along lines mostly occurs in the enhanced submesoscale frontal regions, as for example on the edge of the weak eddies C and D. Those lines may extend to
several hundred kilometers. The left panel in Figure 1 shows a satellite image of analogous ‘lines’ formed by \textit{Sargassum} in the same year of our simulation. Snapshots of regions \textit{A} and \textit{B} confirms that most of the particles concentrate in the convergence zones (blue shading) forming the characteristic patterns due to the strong vertical velocity gradient of the flow at the ocean surface (Fig. 9).

[20] At 100 m depth, the vorticity field is highly correlated with the surface one, as expected based on the vertical extension of the Loop Eddies (Fig. 10). Submesoscale vorticity patterns are also recognizable. However, the intensity of the vorticity scaled by the Coriolis parameter is comparable to the idealized simulation and everywhere less than unity, indicating that the dynamics is not dominated by ageostrophic processes, and the convergence field is too weak to generate any particle accumulation.

[21] Whereas \textit{Sargassum} blooms in the Gulf of Mexico are predominately observed in spring and early summer (March to July, Gower and King 2008), the horizontal divergence/converge field contributes to pattern formation at the ocean surface all year around. In Figure 11, we present the outcome of a separate experiment run with the same set-up but 2 km horizontal resolution and over a domain covering the whole western Gulf of Mexico. 40,955 Lagrangian floats were deployed on January 27, 2004 and their trajectories were integrated for one month. A long-lived large Loop eddy, with a turnover time of about 7 days, dominates the mesoscale field in this simulation and displays a complex horizontal divergence field that causes the particles to aggregate along the convergence at its interior, again with patterns in agreement with the satellite observations.

\textbf{Discussion}
In this work, we investigated how passive tracers are advected at the ocean surface using a regional ocean model run at high spatial resolution. Results have been qualitatively compared to satellite images capturing the *Sargassum* distribution at the ocean mesoscales. Both idealized and realistic model outputs strongly suggest that the highly heterogeneous patterns seen in *Sargassum* distribution at scales of kilometers are due to horizontal convergence/divergence patterns associated with submesoscale ageostrophic processes (McWilliams et al. 2009a; 2009b). The horizontal convergence field is particularly intense at the ocean surface to balance the strong vertical gradient of $w$, and it weakens rapidly below it. As a result, particle distributions at depth do not display any systematic heterogeneity. The stronger the surface horizontal convergence field is, the faster the particles aggregate. For weak convergence/divergence values, the time required to accumulate the particles is longer than the Lagrangian time scale, and particle distribution resemble those found in 2-D turbulence studies. According to our numerical simulations, patterns become recognizable when the model horizontal resolution is greater than 2 km and the temporal resolution of the momentum forcing is at least daily. Such a threshold may be model dependent as the parameterizations used for subgrid vertical and horizontal mixing may have a direct impact on the representation of the convergence/divergence field. The high horizontal resolution is required to properly resolve the structure of the submesoscale circulation. The wind field must contain energy at the inertial frequency (or higher) to excite quasi- and near-inertial motions in the vertical velocity, and therefore high values of $w$ below the surface (Cardona and Bracco 2011).
The Lagrangian tracers adopted in this work are neutrally buoyant and infinitesimally small. *Sargassum*, however, is slightly buoyant, has a finite size and will eventually sink due to epiphytic ballasting and buildup of vegetative (non-buoyant) growth. Particles with buoyancy that differs from the surrounding fluid and with finite size are subject to a complex suite of forces (see Maxey and Riley 1983 for the full equation describing their motion) and their behavior may differ dramatically from the one of neutral tracers. If the particle size is less than the mean free path in the ambient fluid, the two dominant forces influencing the particle motion are the Stokes' drag (i.e., the friction that the fluid exerts on a particle moving with a velocity that is different from the fluid one), and the force that the fluid would exert on a fluid particle placed at the particle position, weighted by the relative inertia (i.e., the ratio of the fluid density to the density of the individual particle) (Crisanti et al. 1992; Tanga and Provenzale 1994; Bracco et al., 1999). To verify that our conclusions are relevant to *Sargassum*, we modified the tracer equation in ROMS to include the two forces above following Tanga and Provenzale (1994) and we considered small spherical particles up to 10% lighter than the surrounding fluid. The resulting equation is a 2nd order ordinary differential equations and its numerical convergence is achieved only with a very small time-step whenever in presence of an energetic advective field. Since the computational time required to advect buoyant tracers is at least one order of magnitude greater than the time required to advect neutral particles, we tested the robustness of our results only in the idealized configuration. We did not find any significant difference when comparing the patterns of neutral versus slightly buoyant tracers, and we verified that the forces acting on the particles do not contribute to the pattern formation.
Significance to Aquatic Environments

As proxy for Lagrangian particles, Sargassum provides a readily visible tracer for the aggregation process. By extension, any surface associated particle would be concentrated by this mechanism as well. This would include fish eggs, larval forms, surface-dwelling zooplankton such as pontellid copepods, siphonophores such Physalia and Vellela, floating anthropogenic debris, and various components of the sea surface microlayer. Whereas the model does not incorporate growth, grazing or migration into/out of this region, and only describes the behavior of a collection of small individuals – large Sargassum patches probably act as a unit – it provides a first-order, easily testable mechanism for focusing biological interactions. This would include such processes as larval recruitment to Sargassum, and concentration of planktonic predator and prey. Large-scale patterns seen in Trichodesmium blooms also exhibit aggregation along curls and lines similar to those seen in the model (Subramaniam et al. 2002; Sarangi et al. 2004). For positively buoyant particles, this organization at scales of tens of kilometers and persistent from weeks to months – depending on the eddy lifetime – would be superimposed on top of Langmuir cell concentration at scales of tens of meters. When coupled to behavioral modes and/or life history cycles that include upward movement, our finding implies that at the ocean surface horizontal patchiness as described in this work will quickly evolve even in the event of uniform upward supply of the surface layer.

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Figure legend

Figure 1: Ocean color satellite images of *Sargassum*. Left panel: *Sargassum* lines in a MERIS MCI (Maximum Chlorophyll Index) image on June 2, 2005 (also in Gower et al. 2006). Right panel: A patch observed by the same satellite in the North Atlantic on September 4, 2008 (http://www.esa.int/esaEO/SEMH61AWYNF_index_2.html). The center of the patch is located at 35°45'N and 66°21'W and its diameter is approximately 45 km.

Figure 2: Relative vorticity scaled by the Coriolis parameter ($\zeta/f$) at the surface (1 m, top left panel) and at 106 m (top right panel) in the idealized integration 60 days after particle deployment, and (bottom) the corresponding particle distribution. Axes are in km.

Figure 3: Distribution of Lagrangian particles deployed at 1 m depth after two days (top left), after two weeks (top right) and after 60 days (bottom), with zooms on distributions of particles deployed inside one vortex overlying the horizontal divergence field (in color). The vortex is destroyed during a merging event after 40 days. Units of divergence field are $10^{-5}$ s$^{-1}$.

Figure 4: Probability of surface particles of falling into areas with different divergence values (aka PDF) at different time during the integration. The x-axis is divergence. (Unit: $10^{-5}$ s$^{-1}$)

Figure 5: Same as in Figure 3 but for particles and divergence field at 106 m depth. The divergence field is plotted using the same color scale in Figure 3.

Figure 6: Same as in Figure 4 but for particles at 106 m depth.

Figure 7: A snapshot of surface relative vorticity scaled by the Coriolis parameter ($\zeta/f$) in the western Gulf of Mexico in May. The four regions described in the text are marked $A$ to $D$. Pattern formation at the ocean surface
Figure 8: Distribution of surface Lagrangian particles overlying the horizontal divergence field (in color, Unit: $10^{-5}$s$^{-1}$) in the western Gulf of Mexico on May 1, 2005, with zooms on Region $A$ and $B$.

Figure 9: Probability of surface particles of falling into areas with different divergence values in the western Gulf of Mexico on May 1, 2005. The x-axis is divergence. (Unit: $10^{-5}$s$^{-1}$)

Figure 10: Relative vorticity scaled by Coriolis parameter (left) and divergence field (right) at 100 m depth in the Gulf of Mexico. (The unit for the divergence is $10^{-5}$ s$^{-1}$)

Figure 11: Distribution of surface Lagrangian particles in the western Gulf of Mexico on February 26, 2005, with a zoom over the large Loop Eddy overlying the horizontal divergence field (in color; Unit: $10^{-5}$s$^{-1}$). The horizontal resolution of this run is 2 km.
Figure 1

Pattern formation at the ocean surface
Pattern formation at the ocean surface
Figure 3

Pattern formation at the ocean surface
Figure 4

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Figure 6
Pattern formation at the ocean surface
Figure 8

Pattern formation at the ocean surface
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Figure 9
Figure 10

Pattern formation at the ocean surface
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