Gulf Stream eddy characteristics in a high-resolution ocean model: Gulf Stream Eddy Characteristics

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Gulf Stream eddy characteristics in a high-resolution ocean model

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\([1]\) A detailed statistical study of the mesoscale eddy activity in the Gulf Stream (GS) region is performed based on a high-resolution multidecadal regional ocean model hindcast. An eddy detection and tracking method that can be used to capture eddy features from large datasets is presented. This method is applied to the 50 year model hindcast within a domain with the most energetic eddy activity along the GS. Detection results are then analyzed to investigate the kinematic properties and temporal variability of GS mesoscale eddies. The studied kinematic properties include the eddy size, duration, intensity, propagation, and spatial distribution. On average, cyclonic eddies are smaller in size but more energetic and remain coherent longer than anticyclonic ones. Cyclonic eddies generally travel further from the generation sites and have a strong tendency for westward propagation with a small equatorward deflection. Anticyclonic eddies remain near their generation locations and tend to propagate northward. The temporal evolution of eddy properties for long-lived eddies (lifetime $>90$ days) is also examined. For both cyclonic and anticyclonic eddies, the size increases rapidly to their maximum value within the first 20 days at which point they begin to slowly decay. In terms of intensity, cyclonic eddies show a quasi-linear decay while the anticyclonic ones reach a quasi-steady state after 3–4 months of a more rapid decay. Finally, the seasonal variability of the GS mesoscale eddies is explored. In autumn and winter, both types of eddies are more numerous and larger but less intense, while in spring they are more intense but less numerous and generally smaller. Several possible mechanisms, including the wind stress, thermal forcing, and topographic influence, are considered to explain the seasonal cycle of eddy variability.


1. Introduction

\([2]\) The Gulf Stream (GS) is a fast, intense western boundary current system in the Northwest Atlantic Ocean transferring heat and salt from the subtropical region in the southwest to the subpolar region in the northeast. It runs parallel to the U.S. east coast until reaching Cape Hatteras at $35^\circ$N where it separates from the coast and flows eastward into the open ocean. While flowing in deep water, the GS meanders can result in the formation and shedding of eddies when the path forms a closed loop. In particular, the intense, long-lived mesoscale eddies have come to be known as "rings" \([\text{Fuglister, 1972}]\). There are two types of rings shed off the GS: cold-core (cyclonic) and warm-core (anticyclonic) rings. Gulf Stream eddies play an important role in the transport of heat and nutrients. Cold-core rings spin off the southward bending meanders entraining cold productive coastal waters and generally travel to the south into the Sargasso Sea, while warm-core rings form off the northward bending meanders enclosing warm Sargasso Sea waters and are generally found north of the stream \([\text{Parker, 1971; Saunders, 1971}]\).

\([3]\) In the past few decades, a significant effort has been made to investigate the dynamics and kinematic behavior of GS rings \([\text{e.g., Richardson, 1980; Olson et al., 1985; Brown et al., 1986}]\) and their influence on oceanic distributions of physical, chemical, and biological properties \([\text{e.g., Backus et al., 1981; Ryan et al., 2001; Borkman and Smayda, 2009}]\). These previous studies are either through theoretical models or field and satellite observations. The most documented results are based on single surveys or time series from individual GS rings. Recently, \text{Chaudhuri et al. [2009]} used 22 years of satellite observations and 20 years of numerical simulations to explore the interannual variability of the GS warm-core rings and their correlation with the North Atlantic Oscillation (NAO). Despite these efforts on GS rings, there still lacks a thorough statistical assessment of the GS eddy activity using numerical models. In this study, we employ a 50 year (1958–2007) high-resolution ocean model hindcast to examine the kinematic
properties and temporal variations of mesoscale eddies along the GS in the Northwest Atlantic Ocean.

In order to study eddy activity from large data sets, it is crucial to have a suitable and competitive algorithm to automatically identify and track eddies. Several methods have been developed to detect eddies from satellite measurements [e.g., Isern-Fontanet et al., 2003; Morrow et al., 2004a; Chelton et al., 2007; Chaingneau et al., 2008] and numerical simulations [e.g., Penven and Echevin, 2005; Doglioli et al., 2007; Nencioli et al., 2010; Xiu et al., 2010]. These automatic eddy detection algorithms can be based either on physical properties or geometric characteristics of the flow field.

For physics-based eddy detection algorithms, a certain physical parameter is calculated and eddies are detected where the calculated values exceed a given threshold. The Okubo-Weiss (OW) parameter [Okubo, 1970; Weiss, 1991], which describes the relative dominance of deformation with respect to rotation of the flow, is one of the most widely used physical quantities to detect eddies [Isern-Fontanet et al., 2003; Penven and Echevin, 2005; Chelton et al., 2007; Xiu et al., 2010]. This method has shown considerable success, although it was reported to have a high number of false detections due to the noise in the field of the OW parameter [Sadarjoen and Post, 2000; Chaingneau et al., 2008]. In order to smooth the OW field, a spatial filter is often applied to the OW field itself before detections [Penven and Echevin, 2005; Souza et al., 2011]. Other physical properties, such as sea surface height (SSH) anomaly and relative vorticity (\(\omega\)), have also been employed to detect eddies [e.g., Morrow et al., 2004b; Chaingneau and Pizarro, 2005; Doglioli et al., 2007]. Doglioli et al. [2007] proposed a method to identify and track three-dimensional eddy structures based on the wavelet analysis of modeled relative vorticity. This method works well for a region with relatively homogeneous eddy activity. However, when applied to a large domain containing regions of very different sea level anomaly (SLA) variances, it may over-identify eddies in the less active parts of the domain [Souza et al., 2011].

For geometry-based eddy detection methods, the shape or curvature of the instantaneous streamlines is used. Two main techniques have been applied: the curvature center method [Leeuw and Post, 1995] locates eddies by finding the centers of curvature in the streamlines, while the winding-angle method [Sadarjoen and Post, 2000] identifies closed streamlines by measuring cumulative changes in streamline direction. Compared to the Okubo-Weiss-based methods, the winding-angle method has shown a more promising result in detecting mesoscale eddies; however, it is more computationally expensive [Sadarjoen and Post, 2000; Chaingneau et al., 2008; Souza et al., 2011]. Another eddy detection method proposed by Nencioli et al. [2010] is based entirely on the geometry of the velocity vectors. Four constraints characterizing the spatial distribution of the velocity vectors around eddy centers were derived and applied to all the grid points. This method shows a very low excess of detection rate (EDR), which is defined as the ratio of the number of incorrectly detected eddies relative to the number of true eddies [Chaingneau et al., 2008; Nencioli et al., 2010] (see Appendix A for details).

In this study, we present a new eddy detection method involving a combination of physical parameters and flow geometry. The basic concept of the detection is to first obtain a rough map of the eddy centers by examining the physical properties (OW and SSH), and then refine the detection by applying a series of geometrical constraints onto the velocity vectors. This method greatly reduces the false detection rate in the OW-based method. It also improves the detection accuracy compared to the purely physics-based or geometry-based methods (see Appendix A). We apply this eddy detection method to the 50 year model hindcast in order to investigate the Gulf Stream eddy characteristics. The rest of the paper is organized as follows: Section 2 gives a brief introduction to the numerical model setup in the Northwest Atlantic and evaluation. Section 3 describes the method used for eddy detection and tracking. The kinematic properties and temporal variability of the GS mesoscale eddies are investigated in sections 4 and 5, respectively. Finally, conclusions are summarized in section 6.

2. High-Resolution Ocean Simulation

2.1. Model Setup

The ocean circulation model used for this study is the Regional Ocean Modeling System (ROMS) [Shchepetkin and McWilliams, 2003, 2005]. ROMS solves the incompressible, hydrostatic, primitive equations with a nonlinear free surface. It employs orthogonal curvilinear coordinates in the horizontal and a generalized topography coordinate in the vertical. Split-explicit time stepping of the barotropic/baroclinic modes allows a substantial increase in the time-step size without affecting the accuracy.

The simulation domain covers the major path of the Gulf Stream in the northwest Atlantic (Figure 1). We employ a grid with a horizontal spacing of 7 km (720 × 360 grid points) and 40 vertical terrain-following levels stretched toward the surface so as to resolve the surface boundary layers. Following the ROMS convention (https://www.myroms.org/wiki/index.php/Vertical_S-coordinate), the vertical coordinate transformation and stretching functions are types 2 and 4, respectively, with the surface and bottom control parameters \(\theta_b = 7, \theta_s = 2\) and the stretching parameter \(T_{z_{line}} = 250\). The baroclinic and barotropic time steps are 180 and 30 s, respectively. Vertical mixing is determined by the K-profile Parameterization [Large et al., 1994]. A spatially variable bottom drag coefficient is used. The bottom bathymetry is derived from the 1 min resolution Shuttle Radar Topography Mission (SRTM) database [Farr et al., 2007], with a minimum depth set to 10 m. The reanalysis data of Simple Ocean Data Assimilation [Carton and Giese, 2008] version 2.1.6 (5 day averaged data with grid resolution of 0.5° × 0.5° × 40 layers) are used for initial and oceanic boundary forcing through the mixed radiation-nudging lateral boundary conditions. A sponge layer of 15 grid points wide is applied along the open boundaries to prevent reflections. Air temperature, sea level pressure, humidity, wind, solar radiation, and precipitation are extracted from the Coordinated Ocean-ice Reference Experiments (CORE.v2) data sets to compute air-sea fluxes using bulk formulae of Large and Yeager [2009]. River discharge is implemented as a fresh water flux using the global river flow and continental discharges estimated by Dai...
et al. [2009]. The model also includes 10 major tidal components extracted from TPXO data set [Egbert and Erofeeva, 2002]. We performed a 50 year (1958–2007) hindcast simulation with model outputs averaged and stored daily. Figure 1 demonstrates a typical model output of the daily mean sea surface temperature (SST) for a given day. The mesoscale meanders and eddies of the GS are clearly seen.

2.2. Model Evaluation

[10] ROMS is one of the most widely used ocean circulation models and has shown a high level of skill in simulating the coastal, regional, and basin-scale ocean dynamics and variability [Haidvogel et al., 2000; Curchitser et al., 2005; Warner et al., 2005; Hermann et al., 2009; Danielson et al., 2011]. As our configuration is a new implementation, we performed extensive model evaluations. Here we present two model metrics relevant to this work.

[11] We first evaluate the model performance in reproducing the mean GS path against satellite derived data. As seen from Figure 1, the Gulf Stream separates from the coast near Cape Hatteras and then flows eastward into the open ocean with prominent mesoscale activity along it. An accurate representation of the GS path is important for an investigation of the mesoscale eddy activity in the region. Drinkwater et al. [1994] derived the location of the GS north wall between Cape Hatteras (75°W) and Tail of the Grand Bank (50°W) based on 20 years (1973–1992) of sea surface satellite thermal imagery. We compare the modeled GS mean path downstream of Cape Hatteras with this satellite-derived product. Following Fuglister [1963] and Joyce et al. [2000], we estimate the GS north wall position with the 15°C isotherm at 200 m depth since it represents an isotherm in the center of the strong horizontal gradient of the GS and lies just to the north of the maximum GS flow at the surface. Figure 2 compares the modeled and satellite-derived annual mean paths of the GS. Although the modeled GS path between 285°E and 289°E lies slightly to the north when compared to observations, the satellite-derived GS path lies within one standard deviation of the modeled trajectory, which shows significant model skill in reproducing the mean GS pathway.

[12] For evaluating the model skill in representing the eddy characteristics, we further compare the modeled eddy kinetic energy (EKE) against satellite observations. The satellite product used here is the gridded delayed time maps of sea level anomalies product (1993–2007) produced by Ssalto/Duacs and distributed by Aviso (http://www.aviso.oceanobs.com/duacs/). This product employs the approach of Ducet et al. [2000] to combine TOPEX/Poseidon, ERS-1,2 and Jason altimetry data into the daily global sea surface height anomaly (SSHA) maps with a spatial resolution of ¼°.

Based on the SSHA maps, it also derives weekly global maps of geostrophic velocities anomalies with the same spatial resolution, which allow us to compute the satellite-derived EKE and compare our model result with it (Figure 3). In the figure, the ROMS result is interpolated onto the AVISO grid and the results outside the common regions of two grids are masked out. The average correlation coefficient between the model and satellite data is 0.71. The comparison highlights the model’s ability to reproduce both the large-scale and fine-scale features of the Gulf Stream. However, the mesoscale variability is underestimated approaching the northeastern open boundary of the model. This is an effect of the formulation along the open boundaries. Lavelle and Thacker [2008] have shown that the damping within the sponge layer has the side effect of also damping the region of interest. Therefore, in this study, we limit the eddy detection and analysis within a smaller region away from boundaries, which contains the most energetic eddy activity in both modeled and observed results. This study region is outlined by the small black box in Figure 3.

3. Eddy Detection and Tracking Algorithm

[13] The eddy detection method developed for this study is based on both physical properties and geometric characteristics of the flow field. It examines the Okubo-Weiss parameter, sea surface height, and velocity vectors that are obtained from the numerical simulations.

[14] The OW parameter [Okubo, 1970; Weiss, 1991] describes the relative dominance of deformation with respect to rotation of the flow and is defined as

\[
\text{OW} = s_x^2 + s_y^2 - \omega^2,
\]

where \(s_x = u_x - v_x\) and \(s_y = v_x + u_y\) are the normal and shear components of strain, respectively, and \(\omega = v_x - u_y\)
is the relative vorticity of the flow. The subscripts \( x \) and \( y \) represent the partial differentiations of the horizontal velocities \( u \) and \( v \) in the \( x \) and \( y \) directions, respectively. Given a positive threshold value of \( OW_0 \), the OW parameter allows partitioning of the flow into different regimes [Elhmaidi et al., 1993]: vortex cores \( (OW < -OW_0) \), organized structures surrounding vortex cores \( (OW > OW_0) \), and the background field \( (|OW| \leq OW_0) \). The threshold is typically taken as \( OW_0 = 0.2\sigma_{ow} \), where \( \sigma_{ow} \) is the spatial standard deviation of the OW parameter [Pasquero et al., 2001; Isern-Fontanet et al., 2003; Chaigneau et al., 2008; Nencioli et al., 2010]. Note that \( OW_0 \) changes with time as the OW parameter depends on time. As mentioned in section 1, the OW method has shown a high rate of false detections. We therefore combine it with the examination of the geometry of the velocity vectors. Moreover, another physical quantity, sea surface height (SSH), is also employed to determine the eddy rotation (cycloic or anticycloic) directions.

### 3.1. Detecting Eddy Centers

The basic concept of our eddy center detection is to first obtain a rough map of the eddy centers by examining the physical properties, and then refine the detection by applying a series of geometrical constraints. The detailed procedure is described as follows:

1. Compute the OW parameter and the threshold \( OW_0 \) from the sea surface velocity field.
2. Detect the local minima and maxima of SSH, as well as the local minima of the OW parameter and sea surface velocity magnitude.
3. Refine the local minima of the surface velocity magnitude with geometric constraints (1)–(4).
4. Combine the minima of the OW parameter, velocity field, and the minima (maxima) of SSH to determine cyclonic (anticyclonic) eddy centers.

In step 3, a series of geometric constraints are applied to each local minimum point of the surface velocity magnitude in order to refine the detection:

1. Along the east-west section across the eddy center, \( v \) has to reverse in sign.
2. Along the north-south section across the eddy center, \( u \) has to reverse in sign.
3. The sign of vorticity cannot change around the eddy center.
4. Around the eddy center, the ratio of maximum and minimum of the velocity magnitude has to be smaller than a given threshold.

The first three constraints are similar to those used by Nencioli et al. [2010]. They are based on the definition of an eddy as a region where the velocity vectors rotate clockwise or counter-clockwise around a center [Okubo, 1970; Weiss, 1991]. A parameter \( n_e \) is used to define how many grid points away from the eddy center where constraints (1)–(3) are checked. A threshold parameter \( r_u \) is also needed as a constraint (4). An appropriate choice of \( r_u \) allows the inclusion of elliptical eddies, such as may present in highly sheared flows. For our high-resolution ROMS simulation, we use \( n_e = 8 \) and \( r_u = 5 \), which we derived from sensitivity tests (see Appendix A for details).

### 3.2. Determining Eddy Sizes and Intensities

Once eddy centers are identified, the eddy boundaries are estimated by the closed contour of \( OW_0 = -0.2\sigma_{ow} \) following Chaigneau et al. [2008]. The eddy radius is defined as

\[
R_e = \sqrt{\frac{A}{\pi}} \tag{2}
\]

where \( R_e \) is the equivalent radius of a circle with the same area \( A \) enclosed by the eddy boundary. Note that the OW criterion may underestimate the area of an eddy as it is restricted to the core of the vortices [Basdevant and Philippovitch, 1994; Doglioli et al., 2007]. The equivalent eddy intensity is estimated as

\[
I_e = \frac{\omega}{f_0} \tag{3}
\]

where \( \omega \) is the absolute value of the mean relative vorticity within the eddy boundary and \( f_0 \) is the Coriolis frequency at the eddy location. For simplicity, we choose a constant value of \( f_0 = 7.3 \times 10^{-5} \) because the distribution of detected eddies does not have a big range in latitudes.
The shape and size of an eddy generally evolve during its life cycle. The shape is not necessarily a regular circle. Our detection algorithm is capable of capturing the irregular shapes of an eddy during its evolution. For example, an anticyclonic eddy is detected in mid-July of 2000 with an initial near circular shape (Figure 4a). It elongates along the east-west direction and becomes more elliptical (Figure 4b) in time. Its major axis also rotates clockwise as it changes shape and size. When the eddy becomes too narrow to maintain coherence, it breaks, as we observe, in mid-August. The top left portion of the eddy forms a smaller elliptical anticyclonic eddy (Figure 4c). It continues to evolve and its major axis rotates clockwise (Figure 4d). Finally, this elliptical shape deforms again and the eddy dissipates in early September.

### 3.3. Tracking Eddy Paths

After eddy centers are detected for the entire length of the simulation, we track eddy paths by comparing their centers at successive time frames (daily here). For a given eddy at time step $t$, an area of $10 \times 10$ grid points centered around it is searched at time step $t + 1$. If an eddy center of the same type (cyclonic/anticyclonic) is found, the location of the new eddy center is recorded at time step $t + 1$. If no eddy center of the same rotation type is found, the location of the eddy center at time step $t$ is recorded and flagged at time step $t + 1$. Then at time step $t + 2$, if the search for the flagged eddy center is successful, the location of the new eddy center is recorded and the flag is removed. If the search fails, the eddy is considered to have dissipated and therefore the record of this particular path stops here. This tracking procedure is analogous to that used by Doglioli et al. [2007] and Nencioli et al. [2010]. It is designed to avoid the error due to missed eddy detection at a single time step. However, as the information for each eddy is kept only two steps after dissipation, an eddy leaving the detection domain and subsequently coming back two steps later is recorded as a new eddy. To estimate the impact on the tracking accuracy of reentering eddies, we compute the ratios of the numbers of eddies that disappear and appear near the boundaries (defined as a region of 15 grid cells wide along the detection boundaries) relative to the total numbers of eddies that dissipate and are generated within the detection domain. In our case, these two ratios are 6.01% and 1.13%, respectively, which implies minor impact on the overall accuracy of the eddy tracking and the resulting analysis. Furthermore, as this study focuses on longer-lived mesoscale eddies, those with lifetime durations shorter than 15 days are discarded.

### 4. Eddy Kinematic Properties

The eddy activity becomes more prominent as the Gulf Stream separates from the coast at Cape Hatteras and flows eastward into the open ocean. As discussed in section 2.2, we choose a study region with the most energetic eddy activity that is away from boundary effects (small black box in Figure 3) and apply the eddy detection and tracking
algorithm to it. In this section, we investigate the statistics of the eddy kinematic properties (size, duration, intensity, propagation, and spatial distribution) using our detection algorithm described above for all eddies in our study domain over the entire simulation period (1958–2007). The annual mean and maximum values of the eddy properties are listed in Tables 1 and 2, respectively. Figure 5 shows the probability distributions of the various eddy properties. The relations between each pair of the properties for cyclonic and anticyclonic eddies are shown in Figures 6 and 7.

### Table 1. Annual Mean Eddy Properties (With Standard Deviations) for all Eddies in the Study Domain Over the Entire Simulation Period (1958–2007)

<table>
<thead>
<tr>
<th>Eddy Type</th>
<th>Number</th>
<th>Radius (km)</th>
<th>Duration (days)</th>
<th>Intensity (normalized by $f_0$)</th>
<th>Distance (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cyclonic</td>
<td>24 (4.1)</td>
<td>58 (13.2)</td>
<td>47 (37.0)</td>
<td>0.22 (0.08)</td>
<td>133 (124.4)</td>
</tr>
<tr>
<td>Anticyclonic</td>
<td>15 (3.7)</td>
<td>64 (21.5)</td>
<td>31 (19.6)</td>
<td>0.14 (0.03)</td>
<td>87 (61.0)</td>
</tr>
</tbody>
</table>

### Table 2. Annual Maximum Eddy Properties for all Eddies in the Study Domain Over the Entire Simulation Period (1958–2007)

<table>
<thead>
<tr>
<th>Eddy Type</th>
<th>Number</th>
<th>Radius (km)</th>
<th>Duration (days)</th>
<th>Intensity (normalized by $f_0$)</th>
<th>Distance (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cyclonic</td>
<td>37</td>
<td>106</td>
<td>382</td>
<td>0.62</td>
<td>962</td>
</tr>
<tr>
<td>Anticyclonic</td>
<td>25</td>
<td>119</td>
<td>165</td>
<td>0.27</td>
<td>472</td>
</tr>
</tbody>
</table>

### Figure 5. Probability distributions of (a) lifetime averaged radius $<R_\text{e}>$, (b) duration, (c) lifetime averaged intensity $<I_\text{e}>$, and (d) propagation distance for all cyclonic (circles) and anticyclonic (stars) eddies in the study domain over the entire simulation period (1958–2007).

#### 4.1. Eddy Size

[36] The radius of an eddy in a given day is estimated by equation (2). However, a mesoscale eddy generally lasts from several weeks to several months. Therefore, the eddy radius (or size) here is referred to as the average value of the daily eddy radius during its lifetime, given by $<R_\text{e}>$. The probability distributions of eddy size for both cyclonic and anticyclonic eddies are shown in Figure 5a. The anticyclonic eddies have a slightly larger mean value of eddy size compared to cyclonic eddies (Table 1). The standard deviation of the eddy size for anticyclonic eddies (21.5 km)
almost doubles that for cyclonic eddies (13.2 km). For both type of eddies, approximately 3/4 of the eddies have radii larger than 50 km. The maximum radius of cyclonic eddies in our study domain is 106 km, while that of anticyclonic eddies is 119 km (Table 2). Note that, as mentioned in section 3.2, the area of an eddy may be underestimated using the OW criterion.

### 4.2. Eddy Duration

The probability distribution of eddy duration (or lifetime) shows a strong asymmetric distribution for either cyclonic or anticyclonic eddies (Figure 5b). In our study domain, the mean lifetimes of cyclonic and anticyclonic eddies are roughly 1.5 and 1 months, respectively (Table 1). The maximum lifetimes of cyclonic and anticyclonic eddies are 106 km and 119 km, respectively.

![Figure 6](image1.png)  ![Figure 6](image2.png)  ![Figure 6](image3.png)

**Figure 6.** Relations between the eddy kinetic properties (lifetime averaged radius $\langle R_e \rangle$, duration, lifetime averaged intensity $\langle I_e \rangle$, and propagation distance) for all cyclonic eddies in the study domain over the entire simulation period (1958–2007).

![Figure 7](image4.png)  ![Figure 7](image5.png)  ![Figure 7](image6.png)

**Figure 7.** Same as Figure 6, but for anticyclonic eddies.
eddy intensity is the average value of the daily eddy intensity during its lifetime. The intensity of cyclonic eddies has a larger mean value and standard deviation compared to that of anticyclonic eddies. Roughly 80% of the cyclonic eddies are with intensity between 0.1 and 0.3, while about 90% of the anticyclonic eddies are with intensity between 0.09 and 0.2. The maximum intensities of cyclonic and anticyclonic eddies are 0.62 \( f_0 \) and 0.27 \( f_0 \), respectively (Table 2). The relation between eddy size and lifetime is similar to that between eddy intensity and lifetime, which implies that the long-lived eddies do not last as long. This may be because that small eddies are generally weak in intensity while very large eddies are unstable to remain their coherent structure very long.

### 4.3. Eddy Intensity

[32] The intensity of an eddy in a given day is estimated by equation (3). Similar to the definition of \( <R_e> \), the eddy intensity here is referred to as the average value of the daily eddy intensity during its lifetime \( <I_e> \). The intensity of cyclonic eddies has a larger mean value and standard deviation compared to that of anticyclonic eddies (Figure 5c). Roughly 80% of the cyclonic eddies are with intensity between 0.1 and 0.3 \( f_0 \), while about 90% of the anticyclonic eddies are with intensity between 0.09 and 0.2 \( f_0 \). The maximum intensities of cyclonic and anticyclonic eddies are 0.62 \( f_0 \) and 0.27 \( f_0 \), respectively (Table 2). The relation between eddy intensity and lifetime is similar to that between eddy size and lifetime, which implies that the long-lived eddies are neither particularly weak nor strong (Figures 6b and 7b).

[33] A tendency for mean eddy intensity to increase with the mean eddy size can be seen for smaller eddies in Figures 6d and 7d. However, this tendency seems to reverse for the larger eddies. Note that the eddy radius (or intensity) here is the average value over the life of the eddy. In section 5.1, we will demonstrate how the eddy size and intensity evolves in time.

### 4.4. Eddy Propagation

[34] The propagation distance of an eddy is defined as the distance between its generation and dissipation locations. The probability distribution of eddy propagation distance (Figure 5d) is quite similar to that of eddy lifetime (Figure 5b). For cyclonic eddies, the relations between eddy propagation distance and the other two eddy properties (Figures 6e and 6f) are similar to those between eddy lifetime and the two properties (Figures 6b and 6a, respectively).

[35] We also investigate the eddy propagation directions by putting all the eddy trajectories together with the same starting point (0, 0) (Figure 8). Cyclonic eddies generally travel far away from the generation sites and have a strong tendency for westward propagation with a small equatorward deflection (Figure 8a). Anticyclonic eddies tend to remain near the generation sites and have a tendency for poleward propagation (Figure 8b). Recent studies of Chelton et al. [2007] and Chelton et al. [2011] examined the global propagation characteristics of long-lived eddies based on satellite altimetry data. They observed that on the global scale the cyclonic and anticyclonic eddies both have a strong westward propagation tendency, but with distinct preferences for poleward and equatorward deflections, respectively. Our results for the GS region are different from these global features due to the mean-current advection and the obstruction of the topographic features to the north.

[36] Moreover, we calculate the mean eddy propagation speed as roughly 2.8 km/day (or 3 cm/s) for cyclonic eddies. Considering the southward deflection, the westward propagation speed for cyclonic eddies is around 2.5 cm/s. This is comparable to the westward phase speeds of classical Rossby waves at latitude of 35°N based on the propagation speed of nondispersive baroclinic Rossby waves shown in Figure 4 of Chelton et al. [2007].

### 4.5. Eddy Spatial Distribution

[37] Figures 9 and 10 depict the spatial distributions of the birthplace and occurrence for all eddies in our study domain throughout the model run. Note that no eddies are detected on the inner eight grid points along the study domain because of constraints (1)–(3) in our algorithm. The eddy generation is distributed over the whole study domain with a concentration located along the mean path of the Gulf Stream. About 65% of the cyclonic and 72% of
the anticyclonic eddies are generated within three standard deviations of the climatological GS pathway. This can be used to roughly estimate the percentage of GS rings with respect to all types of eddies in our study domain. The spatial distribution of eddy occurrence shows that the cyclonic (cold core) eddies are mostly located south of the mean GS path. When traveling to the west, they tend to have a southward deflection due to the obstruction by the Gulf Stream to the north (Figure 10a). On the other hand, the anticyclonic (warm core) eddies are mostly constrained between the GS and the continental shelf (Figure 10b). No strong tendency of westward propagation is observed for anticyclonic eddies due to the obstruction by topographic features. Moreover, the cyclonic eddies generally travel longer distances and a large number of them continue to propagate when approaching the southwestern boundary of our study domain. A similar situation is also observed for some of the anticyclonic eddies when reaching the northwestern boundary.

5. Eddy Temporal Variability

In section 4, we presented a picture of the mean eddy behavior in our study domain. In this section, the temporal evolution and variability of the mesoscale eddies are investigated for different time scales.

5.1. Eddy Evolution

We first examine the mean temporal evolution of eddy size and intensity for long-lived eddies in our study domain (Figure 11). This analysis is based on 143 cyclonic and 17 anticyclonic eddies with a lifetime longer than 90 days. For long-lived cyclonic eddies, size increases rapidly to the maximum values within the first 20 days and then starts to decrease slowly afterward. The radius growth and decay rates are roughly 15 and \(-3\) km per month, respectively. While for long-lived anticyclonic eddies, the radius grows and decays rapidly during the first 2–3 months (at rates of roughly 18 and \(-7.5\) per month), but tend to remain constant afterward. Note that the standard error of eddy size for anticyclonic eddies is large due to the small sample size (Figure 11a). The intensity of long-lived cyclonic eddies decays quasi-linearly by roughly 30% in 5 months, whereas the anticyclonic one decays about 35% in the first 3 months. The anticyclonic intensity shows an increase during the last month, but its standard error is much bigger than in previous months (Figure 11b). Chaigneau et al. [2008] have observed the similar temporal evolution...
characteristics for long-lived mesoscale eddies in the eastern South Pacific along the Peruvian coast.

### 5.2. Seasonal Variability

[40] The monthly variability of the eddy number, size, and intensity is presented in Figure 12. During autumn and winter, both cyclonic and anticyclonic eddies are more numerous, slightly bigger, but less intense. In spring, they are more intense but less numerous and smaller in size. In the summer season, the anticyclonic eddies are both big and intense.

[41] The annual cycle of the mesoscale variability in the western boundary current systems has been explored in many previous studies. However, the dynamics governing the seasonal variability is not fully understood. Here we discuss several possible mechanisms to explain the seasonal variability of the GS eddies. However, a thorough investigation of the dynamics behind the GS eddy properties and variability is beyond the scope of the current paper.

[42] Wind was indicated as a major force driving the upper layer circulation and influencing the eddy activity through baroclinic instability [Gill et al., 1974; Garnier and Schopp, 1999]. The wind stress in the GS region has been observed to have an annual cycle with the maximum in winter and minimum in late summer/early autumn [Fu et al., 1987; Brachet et al., 2004; White and Heywood, 1995]. The EKE in the GS region has been found to follow the seasonal fluctuation of the wind forcing by a couple of months [White and Heywood, 1995; Garnier and Schopp, 1999; Zhai et al., 2008]. In our study domain, the EKE exhibits a seasonal cycle with the maximum in spring/early summer and the minimum in autumn/winter (Figure 13), which is consistent with the seasonal variations of eddy intensity (Figures 12e and 12f).

[43] Besides the wind forcing, the thermal forcing was also suggested to influence the seasonal fluctuation of eddy activity [Qiu, 1999; Chaigneau et al., 2008]. Local heating/cooling can change the density structure in the top few hundred meters and, hence, affect the stability properties of the ocean [Gill et al., 1974]. Figure 14 depicts the modeled monthly mean vertical thermal structures across the GS along 295°E. In summer and autumn, the upper ocean is well stratified due to strong surface heating and weak wind-induced vertical mixing (June–October). As the cooling-induced convection and wind-induced vertical mixing are enhanced, the flat upper thermocline becomes steeper in winter (December–February) and reaches the maximum in early spring (March). Such annual cycle of the vertical thermal structure was also observed by Zhai et al. [2008] along 300°E across the GS based on the World Ocean Atlas. Baroclinic instability is strongest in early spring (March) when the upper thermocline tilt reaches its maximum, which leads the EKE maximum by 2 months. One possible connection between them is that in late spring (May), the available potential energy built up in winter/early spring is released by eddies through baroclinic instability and the thermocline is subsequently flatter in summer (June).

[44] Moreover, the GS position may also be related to the eddy temporal variability [Chaudhuri et al., 2009]. Figure 15 compares the modeled mean GS paths in different seasons. The GS tends to shift southward and flows with reduced meanders during spring, while in autumn and winter it meanders more and tends to shift northward. During the summer season, the GS path has significant curvature but its mean position is in the middle. The overall seasonal variance of the GS pathway is not significant except over the New England Seamount Chain (NESC) region. The GS interaction with the NESC has been shown to affect the GS meandering and eddy formations [Spall and Robinson, 1990; Teague and Hallock, 1990]. The seasonal variability in the GS position may influence the GS-NESC interaction and, hence, affect the GS eddy occurrence.

[45] Overall, different mechanisms appear to work together to produce the observed seasonal variability of the GS eddies. A more thorough investigation will be carried out in following work.

### 6. Summary

[46] We have performed a detailed statistical study of the mesoscale eddy activity along the Gulf Stream (GS) based on a high-resolution regional ocean model hindcast from 1958 to 2007. We analyzed the GS mean path and eddy kinetic energy as a measure of the model’s variability...
and found significant correlations with satellite observations for these fields.

[47] In order to capture the eddy features from large data sets, we first presented an automatic eddy detection and tracking method. This method is based on both physical properties and geometric characteristics of the flow field. It examines the Okubo-Weiss parameter, sea surface height, and velocity vectors obtained from the numerical simulations. We applied this method to the 50 year model hindcast within a domain with the most energetic eddy activity along the GS, and then analyzed the detection results to investigate the kinematic properties and temporal variations of the GS mesoscale eddies.

[48] The studied kinematic properties include the eddy size, lifetime, intensity, propagation, and spatial distribution. We found that, on average, cyclonic eddies are smaller in size but more energetic and of longer duration than anticyclonic eddies. Cyclonic eddies generally travel longer distances from the generation sites and have a strong tendency for westward propagation with a small equatorward deflection. Anticyclonic eddies do not travel far and tend to propagate northward. Previous studies of Chelton et al. [2007] and Chelton et al. [2011] showed that, on the global scale, the cyclonic and anticyclonic eddies both have a strong westward propagation tendency, but with distinct preferences for poleward and equatorward deflections, respectively. Our results for the GS region are different from the global features due to the mean-current advection and topographic obstruction.

Figure 12. Monthly variations of the eddy (a and b) number, (c and d) size, and (e and f) intensity for all cyclonic (left) and anticyclonic (right) eddies in the study domain over the entire simulation period (1958–2007). Associated standard errors are shown with shading.

Figure 13. Monthly variation of the modeled surface eddy kinetic energy averaged over the study domain for entire simulation period (1958–2007). Associated standard error is shown with shading.
We also examined the mean temporal evolution of eddy properties for long-lived eddies (lifetime > 90 days). For both types of eddies, the size increases rapidly to the maximum value within the first 20 days and then decreases slowly afterward. In terms of intensity, cyclonic eddies show a quasi-linear decay while the anticyclonic ones seem to reach a quasi-steady state after 3–4 months of rapid decay.

Finally, we explored the seasonal variability of the GS mesoscale eddies. In autumn and winter, both types of eddies are more numerous, larger, but less intense, while in spring they are more intense but less numerous and smaller. We discussed several possible mechanisms to explain this variability. The wind stress, thermal forcing, and topographic influence seem to work together in producing the season cycle of eddy variability. Further investigations are required to better understand how these mechanisms interact to produce the observed structure of the GS region.

Recent studies have suggested that the long-term variability of mesoscale eddies is related to large-scale climate oscillations [Chaigneau et al., 2008; Chaudhuri et al., 2009]. Future work with our hindcast simulation and

Figure 14. Modeled monthly mean vertical thermal structures along 295°E averaged over the entire simulation period (1958–2007).

Figure 15. Modeled seasonal mean paths of the Gulf Stream averaged over the entire simulation period (1958–2007). Black dashed line represents the vertical section along 295°E discussed in Figure 14. Bathymetry contours are spaced at 1000, 2000, 2500, 3000, 3500, 4000, and 4500 m.
analysis techniques will include an investigation of the interannual and decadal variability of the GS mesoscale features and their relation to the climate of the North Atlantic Ocean.

Eddies are important not only to ocean dynamics but also to biological productivity through their impact on the distribution of heat, salt, momentum, and biogeochemical properties [Backus et al., 1981; Ryan et al., 2001]. In future work, we will explore the links between ecosystems and mesoscale eddies in the GS region with a recently developed coupled circulation-biogeochemistry model for the region.

Appendix A: Evaluation of the Eddy Detection Algorithm

We evaluate the performance of the eddy detection algorithm and examine the sensitivity of detection results on two parameters $n_u$ and $r_u$. Following Chaigneau et al. [2008] and Nencioli et al. [2010], two different detection rates are used to evaluate the algorithm efficiency: the success of detection rate (SDR) and the excess of detection rate (EDR). They are defined as

$$SDF = \frac{N_e}{N_t},$$

(A1)

$$EDF = \frac{N_e}{N_t},$$

(A2)

where $N_e$ is the total number of true eddies, $N_t$ is the number of eddies detected by the algorithm that are true eddies, and $N_d$ is the number of eddies detected by the algorithm that are not true eddies. As in Nencioli et al. [2010], the “true eddies” are manually identified by the authors through examining the OW parameter, SSH and sea surface velocity field.

The evaluation procedure is similar to the one used in Nencioli et al. [2010]. The eddy fields detected by the algorithm with different values of $n_u$ and $r_u$ are compared to the true eddy fields for 10 days randomly selected among the simulation outputs. The detection rates SDR and EDR are computed for different combinations of the two parameters $n_u$ and $r_u$ (Figure 16). Maximum values of the SDR occur for $n_u = 8$ and $r_u$ ranging from 4 to 8. Within the same range, the EDR is below 10% with a minimum value of 4.78% occurring for $n_u = 8$ and $r_u = 5$. Therefore, $n_u = 8$ and $r_u = 5$ is chosen as the optimal combination of the parameters for the eddy detection. The average SDR for our detection algorithm is 93.2%, which is higher than the SDR observed for the geometry-based algorithm (92.9%) and the Okubo-Weiss method (86.8%) in Chaigneau et al. [2008]. It is also slightly higher than the SDR observed for the Okubo-Weiss method (86.8%) in Chaigneau et al. [2008]. Note that a determination of the statistical significance of these differences in SDRs is not possible because the algorithms have been applied to different data sets in these studies. The average EDR for our algorithm (4.78%) is much lower than those for the winding-angle method (18.7%) and the Okubo-Weiss method (63.3%) observed in Chaigneau et al. [2008], but it is slightly higher than the EDR for the method (2.9%) by Nencioli et al. [2010].

References


