

Impact of Bottom Drag on Gulf of Mexico Dynamics

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ABSTRACT

Bottom drag is a key oceanic energy sink, yet its influence on the Loop Current (LC) and Loop Current eddies (LCEs) in the Gulf of Mexico (GoM) has not been fully quantified. Using a suite of numerical simulations (1994–2015), we examine how removing bottom drag entirely or selectively in regions shallower or deeper than 1000 m modifies eddy kinetic energy (EKE), LC variability, and LCE properties. Removing deep bottom drag increases vertically integrated EKE in the eastern GoM by about 15% through enhanced barotropic conversions. However, it reduces LCE separation frequency, weakening the LC EKE tongue, a westward extension of elevated EKE across the central GoM. In contrast, removing shallow bottom drag substantially intensifies EKE within the upper 50 m of the eastern GoM, produces approximately 30% more energetic LCEs, and shifts both their trajectories and the LC EKE tongue southward. This near-surface intensification results from mesoscale activity generated in the shallow Caribbean Sea, particularly the Gulf of Batabanó. Across all configurations, time series of EKE and energy conversions show that LCE detachments coincide with pronounced EKE peaks. Baroclinic conversions contribute more strongly to EKE production during individual events at shorter running-mean windows, whereas barotropic conversions dominate over longer windows. Thus, bottom drag shapes GoM mesoscale dynamics not only through direct energy dissipation but also by regulating energy conversion pathways and the upstream Caribbean mesoscale variability. Accurate representation of bottom drag is therefore essential for realistic simulation and prediction of regional mesoscale circulation.

1. Introduction

Ocean dynamics in the Gulf of Mexico (GoM) are dominated primarily by the Loop Current (LC) system, which consists of the LC, the large anticyclonic eddies that detach from it, known as Loop Current eddies (LCEs), and cyclonic frontal eddies propagating around the LC boundary (Fig. 1). The LC connects the Caribbean Sea to the GoM via the Yucatan Channel and feeds into the Gulf Stream through the Straits of Florida (Hurlburt and Thompson, 1980; Oey et al., 2005). Its variability is characterized by retracted and extended phases (Leben, 2005; Garcia-Jove et al., 2016; Larrañaga et al., 2022). During the extended phase, the LC intrudes into the northern GoM, accumulates energy, and meanders until an LCE detaches, at which point the LC usually shifts abruptly back to a retracted phase. The GoM circulation can be viewed as a two-layer system, with most of the kinetic energy (KE) concentrated in the upper 1000 m (Hamilton et al., 2018; Furey et al., 2018). The LC and its associated mesoscale eddies dominate the upper layer (Hamilton, 2009; Sturges et al., 2013), whereas the lower layer is comparatively stable and largely driven by upper-layer momentum transfers (Maul, 1977; Bunge et al., 2002; Nedbor-Gross et al., 2014). This vertical stratification strongly shapes the regional distribution of energy and momentum.

LCEs are warm-core eddies that often exceed 200 km in diameter and represent the dominant source of mesoscale variability in the GoM (Donohue et al., 2016; Garcia-Jove et al., 2016; Yang et al., 2020; Larrañaga et al., 2022; Zhang et al., 2022). After detachment, an LCE can reattach to the LC or propagate westward across the basin before gradually dissipating along the GoM western boundary (Sturges and Leben, 2000; Leben, 2005). As they propagate westward ($\sim 5 \text{ km day}^{-1}$), LCEs strongly influence the GoM circulation and ecosystem dynamics by modifying the thermohaline structure, biological productivity, and air–sea interactions (Vukovich and Crissman, 1986; Chang and Oey, 2010; Androulidakis et al., 2019; Sosa-Gutiérrez et al., 2020; Meunier et al., 2020; Damien et al., 2021; Larrañaga et al., 2022; Lee-Sánchez et al., 2022). They influence the western GoM circulation by modulating the coastal currents and regional upwelling along the continental slope (Hamilton et al., 1999; Vukovich, 2007; Olvera-Prado et al., 2024).

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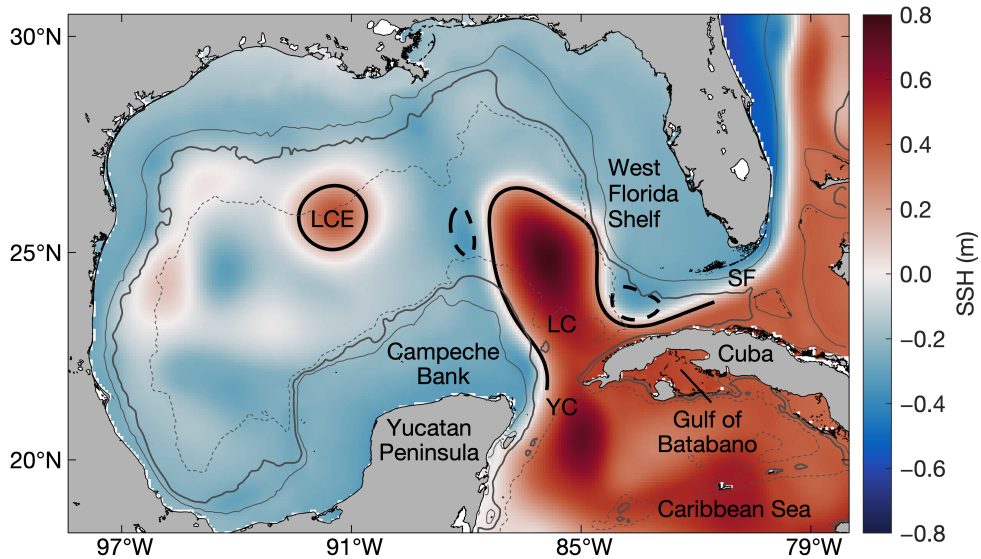


Figure 1: Schematic of circulation in the Gulf of Mexico based on a snapshot of modeled sea surface height (SSH; m). The thick solid black contours indicate the Loop Current (LC) and a Loop Current eddy (LCE). The dashed contours highlight the LC frontal eddies. The Yucatan Channel (YC) and the Straits of Florida (SF) are labeled, alongside key regional geographic features. The gray contours represent the 200 (thin), 1000 (thick), and 2500 m (dashed) isobaths.

LCEs can also contribute to rapid hurricane intensification, as their warm cores enhance the exchange of moisture and heat between the ocean and the atmosphere (Yablonsky and Ginis, 2012; Jaimes et al., 2016; Hong et al., 2000). Finally, the strong currents along the outer edges of LCEs pose significant risks to offshore oil and gas operations along the northern Gulf (Kantha, 2014).

Despite their profound impact on regional dynamics, predicting the timing and location of an LCE shedding event remains a persistent challenge due to the complex interplay of wind forcing, bathymetry, and mesoscale variability. Irregular shedding intervals, ranging from weeks to more than a year, further complicate predictability (Sturges and Leben, 2000; Lugo-Fernández, 2007; Vukovich, 2012; Dukhovskoy et al., 2015; Garcia-Jove et al., 2016; Larrañaga et al., 2022). To understand the essential role of LCEs in shaping GoM dynamics, the physical mechanisms governing their detachment have been extensively examined. Hurlburt and Thompson (1980) emphasized that differential planetary rotation ($\beta > 0$) is a necessary condition for eddy shedding, and Pichevin and Nof (1997) showed that LCE detachment can be triggered when a long Rossby wave outruns the growth of the LC. Leben (2005) and Lugo-Fernández and Leben (2010) stated that the latitude to which the LC retreats after a separation event is inversely related to the time until the next event; southern retreats ($< 25^\circ\text{N}$) lead to longer intervals, whereas northern retreats ($> 25^\circ\text{N}$) yield shorter ones. Current-driven instabilities also play a role in promoting LCE shedding. Chérubin et al. (2006) proposed a two-layer mechanism in which the deep-layer baroclinic energy conversions and the upper-layer barotropic energy conversions together promote LCE detachment. This mechanism is supported by analyses of mooring measurements around the LC that show increases in deep eddy kinetic energy (EKE) at the time of a detachment, driven by the conversion of available potential energy into EKE (Donohue et al., 2016; Hamilton et al., 2016). However, model diagnostics by Yang et al. (2023) and Chaichitehrani and He (2024) seem to imply that barotropic conversions from mean KE to EKE dominate the energy balance during LCE detachment events.

Mesoscale cyclonic eddies propagating along the LC boundary have also been known to play a critical role in LC dynamics and the timing of LCE detachment events (Chérubin et al., 2006; Le Hénaff et al., 2012; Rudnick et al., 2015; Jouanno et al., 2016; Hiron et al., 2020; Nickerson et al., 2022; Larrañaga et al., 2026). LC frontal eddies forming east of the Campeche Bank and Tortugas eddies located southeast of the LC have been shown to promote the detachment and separation of LCEs below 25°N (Fratantoni et al., 1998; Schmitz, 2005; Larrañaga et al., 2026). In addition, the Caribbean Sea acts as an energetic upstream boundary that continuously feeds the LC (Hurlburt and Thompson, 1980; Oey et al., 2005). Variations in transport and vorticity flux through the Yucatan Channel have been shown to influence LC variability (Candela et al., 2002; Oey, 2004; Moreles et al., 2021). Mesoscale eddies originating in the Caribbean

Sea, including cyclonic (Athié et al., 2012; Jouanno et al., 2016) and anticyclonic eddies (Ntaganou et al., 2023), can propagate through the Yucatan Channel and further modulate LCE separation events (Larrañaga et al., 2026). The importance of this upstream influence was highlighted by Le Hénaff et al. (2012) and Garcia-Jove et al. (2016) in idealized numerical simulations, which showed that artificial damping of mesoscale activity modulated LCE shedding frequency and increased LC penetration into the GoM.

Air–sea interactions also play an important role in modulating surface ocean dynamics. Thermal feedback, the influence of sea surface temperature on the atmosphere, has been shown to transfer up to $\sim 10\%$ of the potential energy of (sub)mesoscale eddies to the atmosphere (Putrasahan et al., 2017; Bishop et al., 2020; Renault et al., 2023). Current feedback, the influence of surface currents on the atmosphere, represents the interaction between surface currents and wind stress (Renault et al., 2016b, 2018), which can transfer up to 30% of ocean mesoscale KE to the atmosphere. In the GoM, current feedback also indirectly modifies barotropic and baroclinic energy conversions, influencing LC penetration, LCE shedding statistics, and LCE properties (Larrañaga et al., 2022).

In addition to these surface energy sinks, frictional dissipation within the bottom boundary layer, *i.e.*, bottom drag, is another important sink of KE in the ocean. Near the seabed and in the boundary layer, the velocity decreases toward the bottom, generating strong vertical shear that drives turbulence and extracts KE from the lower layer. Global energy budgets indicate that bottom drag represents a major sink of mechanical energy, dissipating approximately 0.1–0.8 TW of the geostrophic KE (Wunsch and Ferrari, 2004; Sen et al., 2008; Arbic et al., 2009; Wright et al., 2013; Huang and Xu, 2018). In numerical ocean models, resolving the near-bottom current shear directly is impractical, and bottom stress is therefore parameterized through bulk drag laws that relate stress to the near-bottom velocity. The most widely used formulation is a quadratic drag law, in which bottom stress scales with the square of the flow speed (Gill, 1982; Jayne and St. Laurent, 2001). It captures the enhanced dissipation over rough bathymetry and energetic boundary currents and produces realistic KE loss in eddy-resolving simulations (Howarth and Souza, 2005; Wright et al., 2013; Ruan et al., 2021). Because bottom drag regulates how mesoscale motions transfer energy to small-scale turbulence, it also influences the partitioning between barotropic and baroclinic energy pathways and alters eddy–mean flow interactions (Arbic and Flierl, 2004; Trossman et al., 2017).

In summary, the GoM circulation is highly energetic and reflects a delicate balance between energy sources and sinks. The primary energy input is associated with the upper-ocean inflow through the Yucatan Channel, and the main energy sinks are air–sea interactions and frictional dissipation in the bottom boundary layer (Sheinbaum et al., 2002; Oey et al., 2005; Maslo et al., 2020). However, despite the recognized importance of bottom friction in global energetics, the extent to which bottom drag controls LC variability, LCE shedding, and mesoscale energy distribution in the GoM has not yet been documented. The main objective of this study is therefore to quantify how bottom drag modulates LC and LCE dynamics in the GoM. This is achieved by systematically varying the bottom drag configurations in a suite of regional GoM numerical simulations. The paper is organized as follows. The model configuration and experimental design are described in Section 2. Section 3 first documents surface EKE, vertically integrated EKE, and their dependence on bottom drag, and then investigates how bottom drag modulates the vertical structure of EKE and the associated energy conversions. The impacts of bottom drag on the LC and LCE statistics are analyzed in Section 4. The results are summarized and discussed in Section 5.

2. Model configuration and analysis

2.1. Numerical experiments

The numerical experiments are performed using the $1/12^\circ$ Gulf of Mexico configuration (GOMB0.08) of the HYbrid Coordinate Ocean Model (HYCOM version 2.3.01; Bleck, 2002; Chassignet et al., 2003; Halliwell, 2004). The model domain spans 263×193 grid points, covering 98° – 77° W and 18° – 32° N, with 41 hybrid σ_2 layers (Fig. 1). The bathymetry is derived from the high-resolution (~ 1 km) Gulf of Mexico dataset of Velissariou (2014). The model is integrated for 22 years (1994–2015) using initial conditions and daily lateral boundary conditions from the $1/12^\circ$ HYCOM reanalysis (GOFS3.1; <https://www.hycom.org/dataserver/gofs-3pt1/reanalysis>). The hourly atmospheric forcing is derived from the Climate Forecast System Reanalysis (CFSR; Saha et al., 2010, 2014), with turbulent fluxes and wind stress computed following the Kara et al. (2005) bulk formulation. The wind stress is computed using absolute winds to exclude the current feedback and isolate the impact of bottom drag. Freshwater forcing is applied as a surface mass flux, incorporating precipitation and river inflows to capture realistic sea-level and mass changes, with river discharge data sourced primarily from the U.S. Geological Survey (USGS). The horizontal viscosity is a combination of a Laplacian operator ($A_2 = \max[0.5 \times (\Delta x)^2 \times \text{Smagorinsky deformation tensor}, 20 \text{ m}^2 \text{ s}^{-1}]$) and

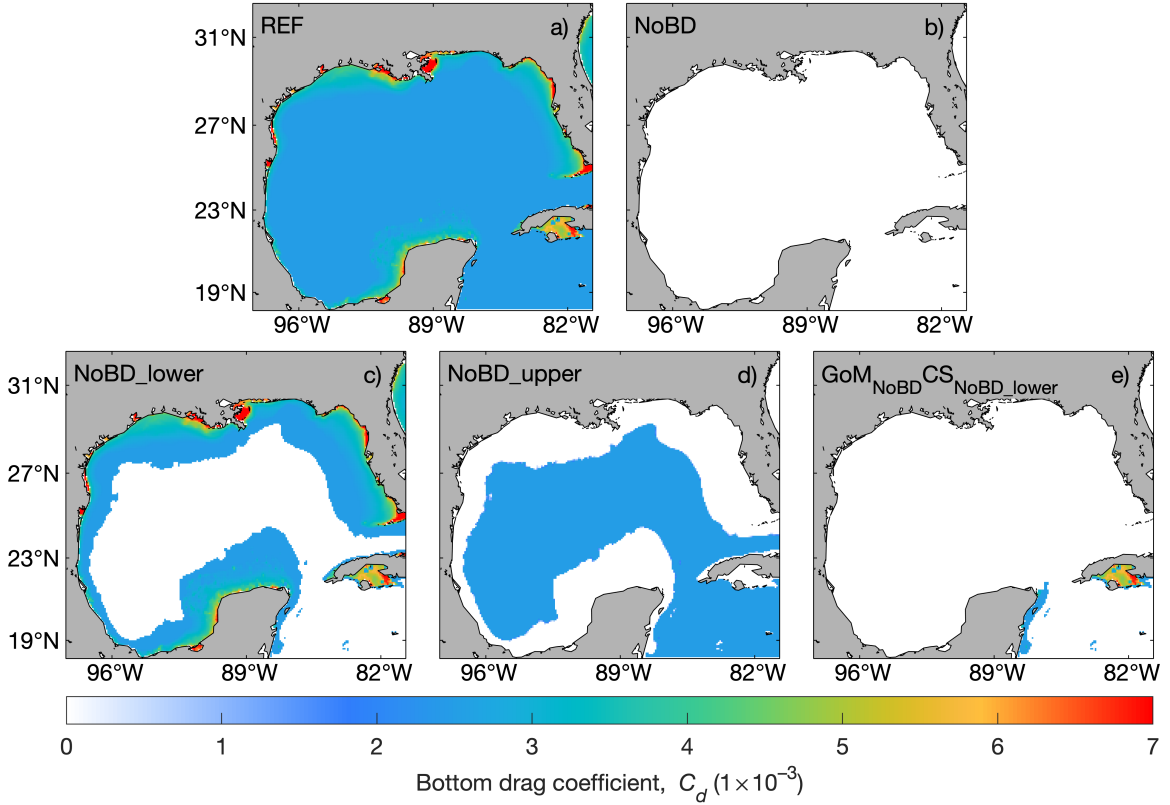


Figure 2: Spatial configurations of the bottom drag coefficient (C_d) for the numerical experiments. (a) REF, in which the bottom drag coefficient is spatially varied, ranging from 2.5×10^{-3} in the ocean interior to 7.5×10^{-3} near the coast, (b) NoBD, where bottom drag is removed within the model domain, (c) NoBD_lower, where bottom drag is removed in regions deeper than 1000 m, (d) NoBD_upper, where bottom drag is removed in regions shallower than 1000 m, and (e) GoM_{NoBD}CS_{NoBD_lower}, where bottom drag is restored only in the Caribbean Sea shallower than 1000 m.

a biharmonic operator ($A_4 = 0.02 \times (\Delta x)^3 \text{ m s}^{-1}$), where Δx is the grid spacing (Chassignet and Marshall, 2008). The K-profile parameterization (KPP) of Large et al. (1994) is used for vertical mixing in the surface mixed layer as well as in the ocean interior. The bottom stress (τ_b) is parameterized using a quadratic drag law:

$$\tau_b = \rho_0 C_d |\mathbf{u}_b| \mathbf{u}_b \quad (1)$$

where $\rho_0 = 1000 \text{ kg m}^{-3}$ is the reference seawater density, C_d is the bottom drag coefficient, and $\mathbf{u}_b$ is the bottom current velocity.

To systematically document the impact of bottom drag on the GoM circulation, its spatial distribution is modified relative to the reference experiment in a series of sensitivity experiments, with all other model parameters remaining unchanged (Fig. 2). In the reference experiment (REF), the bottom drag coefficient varies spatially, ranging from 2.5×10^{-3} in the ocean interior to 7.5×10^{-3} near the coast (Fig. 2a), reflecting stronger frictional damping in shallow coastal regions (Jayne and St. Laurent, 2001). In the NoBD (No Bottom Drag) experiment, the bottom drag coefficient is set to zero everywhere across the domain (Fig. 2b). In the NoBD_lower experiment, bottom drag is removed in regions deeper than 1000 m but retained in shallower regions to assess the impact of deep-ocean friction (Fig. 2c). Conversely, the NoBD_upper experiment isolates the effect of bottom drag in shallow regions by removing it only in regions shallower than 1000 m (Fig. 2d). The 1000-m isobath was selected as the spatial threshold because it dynamically partitions the upper layer, where most KE is concentrated, from the less energetic lower layer. The GoM_{NoBD}CS_{NoBD_lower} experiment is identical to NoBD, except for nonzero bottom drag in Caribbean Sea regions shallower than 1000 m (Fig. 2e).

2.2. Loop Current front and eddy tracking

In this study, we use the 0.17-m contour in the sea surface height (SSH) field to detect the LC front and identify LCE shedding events (Leben, 2005). Although alternative tracking methods have been proposed, such as Kalman filtering of SSH gradients (Dukhovskoy et al., 2015) or following the maximum velocity streamline connecting the Yucatan Channel to the Straits of Florida (Laxenaire et al., 2023), these approaches yield LC metrics that are in close agreement with the widely adopted 0.17-m SSH contour. Following Dukhovskoy et al. (2015), the spatially averaged SSH over the model domain is subtracted from the SSH fields to remove large-scale steric temporal variability.

Mesoscale eddies are tracked using an automated eddy tracking technique developed by Chaigneau et al. (2009) and adapted for the GoM by Sosa-Gutiérrez et al. (2020) and Larrañaga et al. (2022). Eddy centers are identified as local maxima and minima in daily sea level anomaly (SLA) fields, with their outer edges defined by the outermost closed SLA contour. To trace the trajectory of each eddy, we apply the algorithm proposed by Pegliasco et al. (2015), which links eddy centers across consecutive daily SLA maps, enabling continuous tracking of their movement over time. To ensure statistical robustness and filter out short-lived features, only eddies with a lifetime exceeding 10 days are considered in our analysis. For analyses of LCE statistics, only LCEs with a lifetime longer than 200 days are considered (Chelton et al., 2011; Sosa-Gutiérrez et al., 2020; Larrañaga et al., 2022).

3. Sensitivity of eddy kinetic energy to bottom drag

3.1. Surface eddy kinetic energy

Understanding the distribution of EKE in the GoM and its sensitivity to bottom drag is essential for assessing the fundamental role of mesoscale dynamics in the regional circulation and energy transfer. We first characterize the observed surface EKE derived from altimetry and then evaluate its representation in the reference numerical simulation (REF), which serves as the baseline for the subsequent sensitivity experiments. The observed surface EKE is computed using geostrophic currents derived from $1/4^\circ$ daily SLA fields (1994–2015) from the Copernicus Marine Environment Monitoring Service (CMEMS; <https://www.copernicus.eu>) and is then compared to HYCOM surface EKE over the same time period (Fig. 3). To facilitate the regional analysis, the GoM is divided into three regions, delimited by segmented lines in Fig. 3: the eastern GoM, dominated primarily by the LC and its interactions with frontal eddies; the central GoM, which serves as the transit path for westward-propagating LCEs; and the western GoM, often referred to as the “eddy graveyard,” where LCEs interact with topography and gradually decay (Biggs et al., 1996). Surface EKE is spatially integrated over each region where water depth exceeds 200 m (Fig. 3g), the approximate depth of the GoM shelf break, to suppress shelf-driven variability (*e.g.*, river plumes and local winds) and highlight the deep-basin mesoscale circulation.

The observations (Fig. 3a) depict intense EKE around the LC with maxima along the northeastern Campeche Bank and the southwestern Florida Shelf. These regions are associated with strong barotropic and baroclinic energy conversions (Donohue et al., 2016; Garcia-Jove et al., 2016; Hamilton et al., 2019; Yang et al., 2020; Larrañaga et al., 2022; Yang et al., 2023) and frequent frontal cyclonic eddies (Lee et al., 1995; Fratantoni et al., 1998; Zavala-Hidalgo et al., 2003; Rudnick et al., 2015; Jouanno et al., 2016; Hiron et al., 2020). In the central GoM, the EKE exhibits a pronounced westward-extending feature, the so-called “LC EKE tongue”. This climatological signature arises from the recurrent westward propagation of LCEs, which transport energy into the central basin and leave a persistent imprint on the long-term mean EKE field (Vukovich, 2007). In the western GoM, the surface EKE is lower than in the central basin (Fig. 3a,g), as LCEs decay and disperse while interacting with the topography (Biggs et al., 1996).

The reference simulation (REF) closely captures the main spatial characteristics of the observed surface EKE field, although regional differences exist (Fig. 3b). In the eastern GoM, REF reproduces EKE maxima along the LC pathway near the northeastern Campeche Bank and the southwestern Florida Shelf (Fig. 3b). However, the strong EKE north of 26°N seen in observations is weaker in REF, and the spatially integrated EKE in the eastern GoM is lower (Fig. 3g). Downstream in the central GoM, the LC EKE tongue associated with westward-propagating LCEs is narrower and more elongated than in observations. In REF, it remains confined south of 27°N , while altimetry shows an extension to 28°N . The EKE is about 25% weaker in the northwestern LC than in observations, primarily due to a reduced northward LC penetration in the model. In the western GoM, REF overestimates EKE compared to observations, with the largest differences in the northwestern basin (23° – 27°N ; Fig. 3b). These discrepancies can be attributed to differences in eddy properties and trajectories. Additional factors, such as coarse resolution, atmospheric forcing, and topography, can also contribute to deviations in the modeled surface EKE when compared to observations. Despite

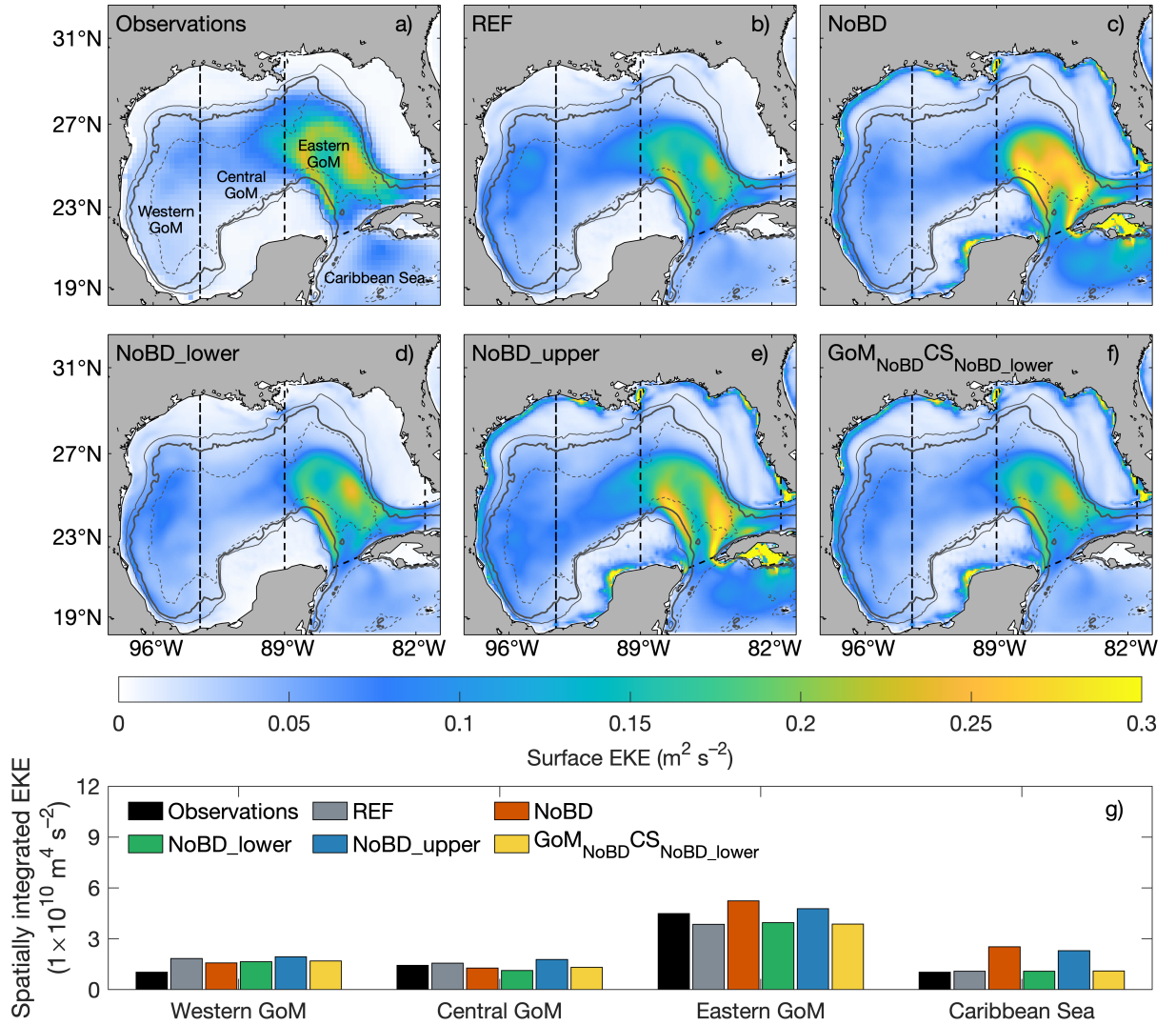


Figure 3: Surface eddy kinetic energy (EKE; $\text{m}^2 \text{s}^{-2}$) for (a) observations and (b–f) simulations. The black segmented lines delimit the western, central, and eastern GoM and the Caribbean Sea. The gray contours represent the 200 (thin), 1000 (thick), and 2500 m (dashed) isobaths. (g) Spatially integrated surface EKE ($\text{m}^4 \text{s}^{-2}$) over each region deeper than 200 m.

these regional differences, the reference experiment (REF) successfully reproduces the fundamental spatial structure of the climatological surface EKE field.

Now, to assess the impact of bottom drag on the modeled GoM circulation, we compare the EKE distribution and LCE dynamics in the three sensitivity experiments (NoBD, NoBD_lower, and NoBD_upper; Fig. 2b–d) to the reference experiment by highlighting regional changes in mesoscale activity across the GoM. Removing bottom drag across the domain (NoBD) results in a $\sim 40\%$ increase in surface EKE in the eastern GoM relative to REF (Fig. 3c,g). On the other hand, setting the bottom drag to zero only in regions deeper than 1000 m (NoBD_lower) does not substantially change the surface EKE pattern in the eastern GoM when compared to REF (Fig. 3d,g). However, removing bottom drag exclusively in regions shallower than 1000 m (NoBD_upper) produces a substantial $\sim 25\%$ increase in surface EKE, particularly along the eastern Campeche Bank and near the Dry Tortugas Islands (Fig. 3e,g). These contrasting responses indicate that surface EKE in the eastern GoM is more sensitive to shallow frictional damping than to deep-ocean bottom drag. Consistent with the eastern GoM, the central GoM EKE in NoBD_upper

is ~10% stronger than in REF (Fig. 3g) with an LC EKE tongue that extends further south (Fig. 3e). In contrast, the EKE tongue is significantly weaker in both NoBD and NoBD_lower (Fig. 3c,d), reducing EKE by 15–30% relative to REF in the central basin (Fig. 3g). Further west in the GoM, the surface EKE is enhanced in NoBD_upper by ~5% when compared to REF (Fig. 3g). However, while REF features an EKE maximum in the northwestern GoM (Fig. 3b), the EKE in NoBD_upper is more evenly distributed and shifted toward the southwestern basin (Fig. 3e). This shift is consistent with the southward extension of the LC EKE tongue in NoBD_upper, suggesting that the separated LCEs preferentially follow southern trajectories when there is no bottom drag on the shelf (see Section 4 for a discussion). While NoBD also exhibits an EKE tongue to the south, its overall magnitude is weaker, leaving the western GoM without a well-defined EKE core (Fig. 3c). In contrast, NoBD_lower reduces this southward extension, maintaining an EKE core in the northwestern GoM that closely resembles REF (Fig. 3d). The results indicate that the response of the eastern GoM to different bottom drag configurations leads to downstream effects that shape both the LC EKE tongue and the western distribution of surface EKE.

An unexpected result from this comparison is that changes in GoM surface EKE are also accompanied by drastic EKE anomalies upstream in the Caribbean Sea when bottom drag is removed in regions shallower than 1000 m. Specifically, one can note in Fig. 3 that the surface EKE is significantly elevated by 125–150% in the shallow Gulf of Batabanó (south of Cuba) in both NoBD and NoBD_upper (Fig. 3c,e,g). Because this intensification is absent in NoBD_lower, where shallow bottom drag is retained, it suggests that undamped mesoscale activity in the Caribbean Sea can propagate northward through the Yucatan Channel and act as an upstream source of mesoscale energy for the eastern GoM. To confirm and quantify the impact of this energy pathway, we designed an additional experiment where the bottom drag was selectively restored in the Caribbean Sea in areas shallower than 1000 m, including the Gulf of Batabanó (Fig. 2e). This experimental design isolates the effects of both the drag distribution as a function of depth (above vs. below 1000 m) and its geographical location (GoM vs. Caribbean Sea). We find that, in the absence of bottom drag in regions shallower than 1000 m, both the cyclonic and anticyclonic eddies in the Caribbean Sea are substantially strengthened in the upper 200 m. Anticyclones exhibit about 50% higher KE in NoBD and 25% higher in NoBD_upper than in REF, and cyclones are ~60% more energetic in both experiments. Restoring bottom drag in the shallow Caribbean Sea reduces surface EKE in the Gulf of Batabanó, weakens the intensity of Caribbean eddies, modifies the spatial distribution of surface EKE, and brings the EKE field closer to that in REF (Fig. 3f). This striking reduction indicates that the intensification of surface EKE in the eastern GoM in NoBD and NoBD_upper is largely a downstream consequence of mesoscale activity generated in the shallow Gulf of Batabanó in the absence of bottom drag.

3.2. Vertically integrated EKE

The surface EKE is only an indication of how the bottom drag affects the energy distribution throughout the water column. To quantify the latter, we now examine the vertically integrated EKE (Fig. 4). In all experiments, the vertically integrated EKE shows broadly similar patterns to those of the surface EKE in the central and western GoM, but differs in the eastern GoM (Fig. 3). The vertically integrated EKE is more spatially uniform, whereas the surface EKE exhibits maxima near the northeastern Campeche Bank and southwestern Florida Shelf. Not surprisingly, since bottom drag is a major energy sink, the EKE in the eastern GoM is 15–20% higher in NoBD and NoBD_lower than in REF (Fig. 4b,c,f). This increase is largest northeast of the LC and southeast of the Mississippi fan. These differences do not arise from the surface EKE maps alone and indicate a greater accumulation of energy at depth when the bottom drag is removed. Furthermore, when the bottom drag is restored in Caribbean Sea regions shallower than 1000 m, the spatial patterns and volume-integrated EKE remain nearly identical to those in NoBD (Fig. 4e). This comparison indicates that enhanced mesoscale activity generated in the shallow Gulf of Batabanó in the absence of bottom drag has a strong downstream surface EKE signature. However, it does not regulate the full-depth EKE in the GoM.

To further investigate the vertical distribution of mesoscale energy in the eastern GoM, Fig. 5 compares vertical profiles of spatially integrated EKE and volume-integrated values in discrete depth ranges (0–50 m, 50–1000 m, and 1000 m–bottom). Across all simulations, over 85% of the total EKE resides above 1000 m, consistent with the dominance of upper-ocean mesoscale variability (Hamilton et al., 2018; Furey et al., 2018). In NoBD, the EKE increases throughout the entire water column when compared to REF (Fig. 5b). NoBD_lower produces EKE similar to that in REF within the upper 0–50 m, but exhibits stronger EKE in the 50–1000 m and 1000 m–bottom layers, resulting in a total energy increase comparable to that of NoBD (Fig. 4f). In contrast, NoBD_upper shows an approximately 10% increase in EKE within the upper 50 m, while energy below this depth remains comparable to REF. The results associate this increase with enhanced mesoscale activity in the shallow Gulf of Batabanó when local bottom drag is

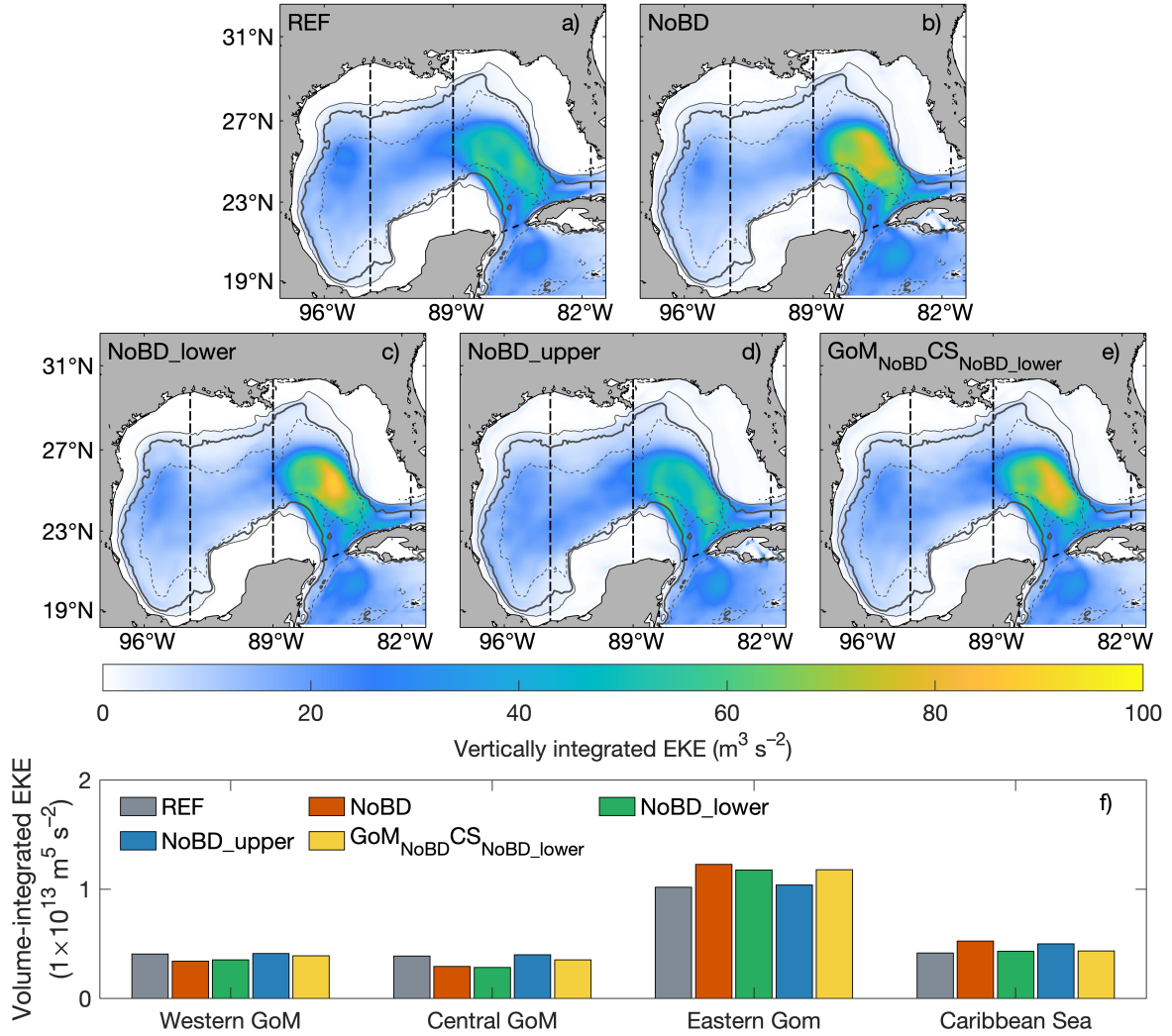


Figure 4: Vertically integrated eddy kinetic energy (EKE; $\text{m}^3 \text{s}^{-2}$). The black segmented lines delimit the western, central, and eastern GoM and the Caribbean Sea. The gray contours represent the 200 (thin), 1000 (thick), and 2500 m (dashed) isobaths. (f) Volume-integrated EKE ($\text{m}^5 \text{s}^{-2}$) over each region deeper than 200 m.

absent. Restoring bottom drag in Caribbean Sea regions shallower than 1000 m (GoM_{NoBD}CS_{NoBD_lower}) eliminates the upper-ocean EKE maximum observed in NoBD (Fig. 5a). However, changes below 50 m are negligible, indicating that the increase in EKE is largely confined to the near-surface layer. Taken together, these EKE vertical profiles emphasize the distinct decoupling between deep and shallow processes. Removing bottom drag in regions deeper than 1000 m allows mesoscale energy to persist and accumulate throughout the water column, highlighting its role in energy dissipation. The following subsection explores the underlying energy conversion terms to provide some mechanistic insights into the changes in EKE across simulations.

3.2.1. Energy budget

To evaluate how the bottom drag modulates EKE in the eastern GoM, we use the simplified EKE budget of Marchesiello et al. (2003) and focus exclusively on energy conversions and bottom drag dissipation. These terms represent the relevant physical mechanisms through which mesoscale energy is generated and dissipated (Marchesiello et al., 2003; Contreras et al., 2023b). The remaining terms, *i.e.*, the advective transport and pressure work terms (Maslo et al., 2020), primarily redistribute the energy components spatially rather than generate or dissipate EKE. Using a

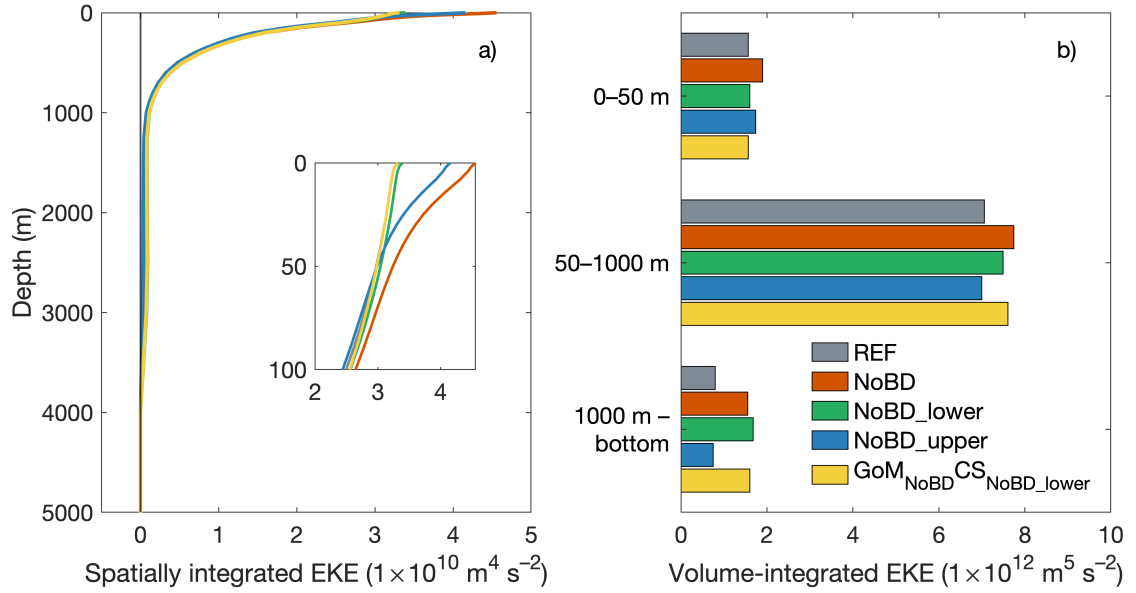


Figure 5: (a) Vertical profile of spatially integrated EKE (m⁴ s⁻²) over the eastern GoM. The inset highlights the upper 100 m. (b) Volume-integrated EKE (m⁵ s⁻²) over the eastern GoM, partitioned by depth range: 0–50 m, 50–1000 m, and 1000 m–bottom.

long-term temporal Reynolds decomposition in which u' and v' are the deviations of the zonal and meridional velocity components, respectively, from the long-term means (1994–2015), the key terms in the energy budget are separated into mean and eddy components:

$$F_{\text{eb}} K_{\text{eb}} = \frac{1}{\rho_0} \left(\overline{u'_b \tau'_x} + \overline{v'_b \tau'_y} \right), \quad (2)$$

$$K_{\text{m}} K_{\text{e}} = - \int_z \left(\overline{u' u'} \frac{\partial \bar{u}}{\partial x} + \overline{u' v'} \frac{\partial \bar{u}}{\partial y} + \overline{u' w'} \frac{\partial \bar{u}}{\partial z} + \overline{v' u'} \frac{\partial \bar{v}}{\partial x} + \overline{v' v'} \frac{\partial \bar{v}}{\partial y} + \overline{v' w'} \frac{\partial \bar{v}}{\partial z} \right) dz, \quad (3)$$

$$P_{\text{e}} K_{\text{e}} = - \int_z \frac{g}{\rho_0} \overline{\rho' w'} dz, \quad (4)$$

where ρ is the potential density with a reference value $\rho_0 = 1000 \text{ kg m}^{-3}$ and g is the gravitational acceleration. The overbar ($\bar{}$) denotes the time mean (1994–2015), and the prime ($'$) indicates deviations from the mean (*i.e.*, eddy components). The bottom dissipation term ($F_{\text{eb}} K_{\text{eb}}$) measures KE dissipation on the seafloor through the covariance of bottom velocity and stress anomalies, indicating how friction removes kinetic energy from the flow. The barotropic conversion term ($K_{\text{m}} K_{\text{e}}$) represents KE transfer between the mean flow and eddy motions through Reynolds stresses acting on mean-flow shear. The baroclinic conversion term ($P_{\text{e}} K_{\text{e}}$) measures the transformation of available eddy potential energy to EKE through vertical velocity shear in the stratified flow. Positive $K_{\text{m}} K_{\text{e}}$ indicates a transfer of mean KE to EKE, whereas negative values indicate the transfer from EKE to the mean KE. In contrast, positive $P_{\text{e}} K_{\text{e}}$ indicates a conversion of eddy available potential energy to EKE, whereas negative values indicate a conversion of EKE back to eddy available potential energy (Oey, 2008; Wang and Pierini, 2020; Yang et al., 2021).

In all simulations, barotropic conversions are the primary energy source, contributing roughly 85% of EKE production in the eastern GoM (Fig. 6a), in agreement with earlier studies (Garcia-Jove et al., 2016; Larrañaga et al., 2022; Chaichitehrani and He, 2024). The vertical distribution of both barotropic and baroclinic energy conversions in the eastern GoM shows that most EKE production occurs above 1000 m (Fig. 6b–e). The vertical profile of barotropic

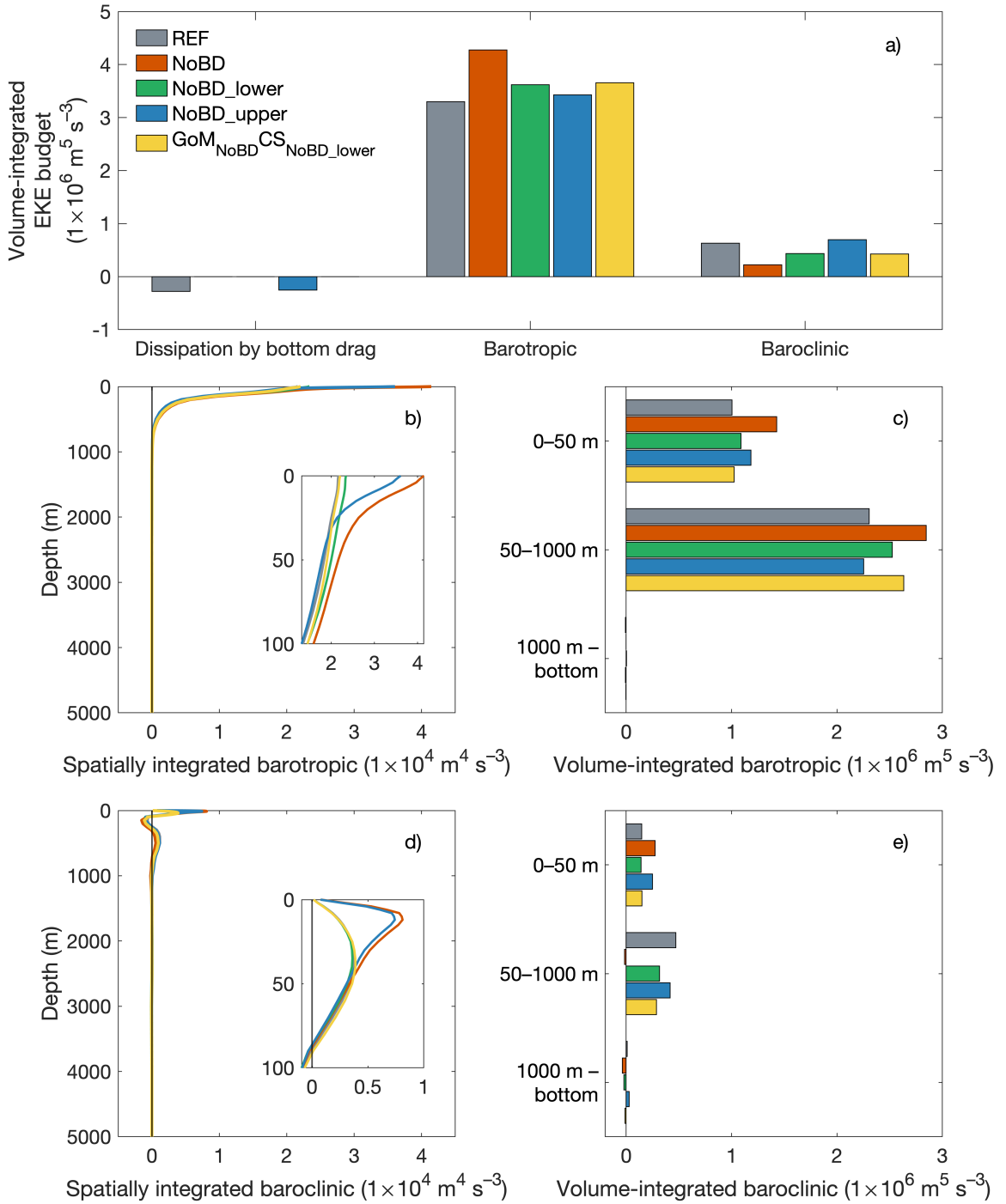


Figure 6: (a) Volume-integrated energy budget terms ($\text{m}^5 \text{ s}^{-3}$) over the eastern GoM. Vertical profiles of spatially integrated (b) barotropic ($K_m K_e$) and (d) baroclinic ($P_e K_e$) energy conversions ($\text{m}^4 \text{ s}^{-3}$), with insets highlighting the upper 100 m. Volume-integrated (c) barotropic and (e) baroclinic conversions ($\text{m}^5 \text{ s}^{-3}$) over the eastern GoM, partitioned by depth: 0–50 m, 50–1000 m, and 1000 m–bottom.

conversions (Fig. 6b) closely follows that of EKE (Fig. 5a) and intensifies near the surface. When divided by depth ranges, barotropic conversions in the 50–1000 m layer are nearly twice those in the upper 50 m (Fig. 6c). In contrast, baroclinic conversions are more evenly distributed between these layers (Fig. 6e). Barotropic conversions provide the main contribution to EKE production in the eastern GoM, and the overall vertical profiles of barotropic and baroclinic conversions are in good agreement with previous studies (Chérubin et al., 2006; Maslo et al., 2020).

In REF, the bottom drag dissipates EKE at a rate of $0.28 \times 10^6 \text{ m}^5 \text{ s}^{-3}$ (Fig. 6a). As already stated, barotropic conversions are the main driver of EKE generation, contributing $3.30 \times 10^6 \text{ m}^5 \text{ s}^{-3}$, while baroclinic conversions account for $0.63 \times 10^6 \text{ m}^5 \text{ s}^{-3}$. In NoBD, the EKE budget differs, and barotropic conversions increase by $\sim 30\%$ in the eastern GoM, driven by stronger horizontal shear resulting from the removal of bottom drag (Fig. 6a). Baroclinic conversions are also 86% higher in the upper 50 m but turn negative near 100 m (Fig. 6d). As a result, baroclinic conversions in the 50–1000 m layer decrease to near zero (Fig. 6e), yielding a 64% reduction in total baroclinic conversions relative to REF (Fig. 6a). In NoBD_lower, the energy conversions in the upper 50 m remain similar to those in REF. Between 50 and 1000 m, barotropic conversions increase, while baroclinic conversions decrease relative to REF (Fig. 6c,e), resulting in total barotropic and baroclinic conversions that are about 10% higher and 30% lower, respectively (Fig. 6a). Taken together, configurations without bottom drag in regions deeper than 1000 m (NoBD and NoBD_lower) have higher velocities near the bottom, which overall strengthen horizontal shear while weakening vertical shear. This dynamically suppresses baroclinic conversions but enhances barotropic conversions, ultimately driving the observed EKE increase in the eastern GoM (Fig. 4b,c). On the other hand, in NoBD_upper, barotropic and baroclinic conversions in the upper 50 m increase by 18% and 70%, respectively, while changes in the 50–1000 m layer are negligible when compared to REF (Fig. 6c,e). These near-surface enhancements result in about a 10% increase in EKE in the upper 50 m (Fig. 5b). Because bottom drag is retained in regions deeper than 1000 m, these intensified energy conversions and EKE are confined near the surface (0–50 m), and the vertically integrated EKE remains close to REF (Fig. 4d,f). Consistent with the EKE profiles, the energy budget exhibits a similar vertical response to changes in Caribbean bottom drag. When bottom drag is restored in the Caribbean Sea, both barotropic and baroclinic conversions weaken significantly in the upper 50 m (Fig. 6). The robust reduction in both conversion terms further indicates that shallow Caribbean bottom drag acts primarily as a near-surface damping mechanism in the eastern GoM. These results confirm that bottom drag in the shallow Caribbean Sea plays a significant role in regulating the mesoscale variability entering through the Yucatan Channel.

In summary, in line with previous studies (Garcia-Jove et al., 2016; Larrañaga et al., 2022), barotropic conversions are the dominant source of EKE in the eastern GoM. The bottom drag contributes both as a local sink and as an indirect control on instability pathways by modifying horizontal and vertical shear. Removing bottom drag in regions deeper than 1000 m weakens baroclinic conversions and strengthens barotropic conversions, increasing EKE in the eastern GoM. In contrast, when bottom drag is removed only in regions shallower than 1000 m, barotropic and baroclinic conversions intensify within the upper 50 m, but total EKE production remains close to that of the reference experiment. These results indicate that bottom drag in the deep GoM basin modulates the vertical accumulation of energy, whereas drag in the shallow Caribbean Sea mainly regulates near-surface EKE production.

3.2.2. Temporal variability of EKE and energy conversion terms

While the preceding long-term EKE budget characterizes the net contribution of individual energy-conversion pathways, it can mask enhanced EKE production during episodic events such as LCE detachment and separation. To examine the temporal evolution of the LC system energetics, we analyze the time series of volume-integrated EKE together with the associated barotropic and baroclinic energy conversion terms. Following previous studies (Jouanno et al., 2012; Yang et al., 2020; Chaichitehrani and He, 2024), the current velocity is decomposed into $\mathbf{u} = \mathbf{u}_{lp} + \mathbf{u}'$, where $\mathbf{u}_{lp}$ is the low-pass component obtained from a running mean and $\mathbf{u}'$ is the corresponding anomaly. We first evaluate the sensitivity of barotropic and baroclinic energy conversions to different running-mean windows in the reference simulation (Fig. 7a,b). In the upper 50 m, baroclinic energy conversions are negligible (Fig. 7b) and barotropic energy conversions increase with the window length and stabilize around 180 days (Fig. 7a). In the 50–1000 m layer, both energy conversion terms are largest. Barotropic conversions exhibit a similar dependence on the running-mean window to that observed in the upper 50 m (*i.e.*, increase with the window length and stabilize after 180 days), while baroclinic conversions peak at 60 days. Both barotropic and baroclinic energy conversions below 1000 m remain negligible across all running-mean windows.

Because the 60- and 180-day decompositions define the mean flow differently and therefore yield different conversion estimates, we then examine the temporal evolution of the volume-integrated EKE and associated energy

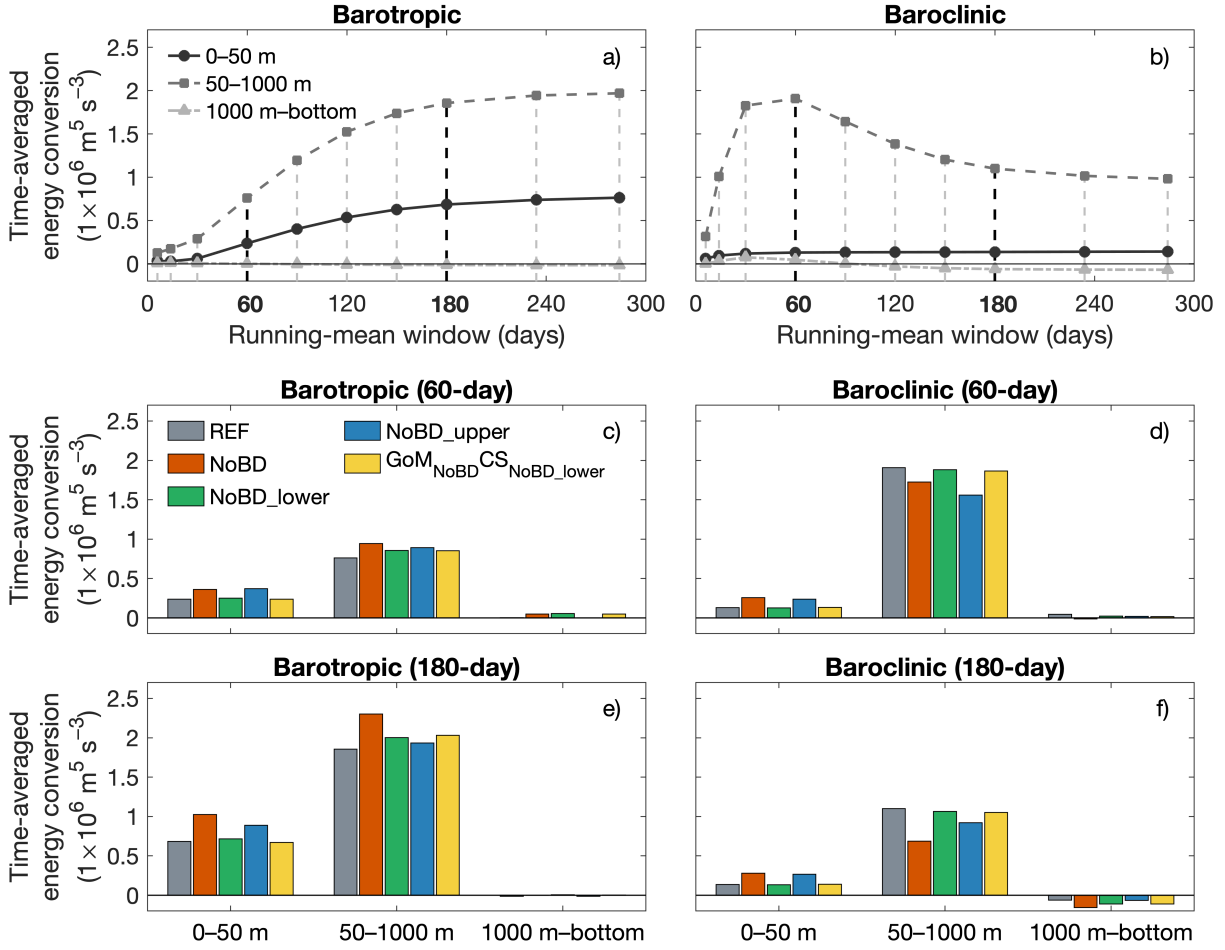


Figure 7: Sensitivity of barotropic ($K_m K_e$) and baroclinic ($P_e K_e$) energy conversions to the running-mean window and bottom drag configuration. Time-averaged (a) barotropic and (b) baroclinic energy conversions in the reference simulation (REF) as a function of the running-mean window (days), partitioned by depth: 0–50 m, 50–1000 m, and 1000 m–bottom. The black dashed vertical lines indicate the 60- and 180-day windows selected for further analysis. Time-averaged energy conversions for all numerical experiments using the (c,d) 60-day and (e,f) 180-day running-mean decompositions.

conversions for these two windows (Fig. 8). At all depths and for both time windows, the EKE displays recurrent peaks that closely correspond to LCE detachment events, regardless of whether the LCE reattaches to the LC or fully separates (Chérubin et al., 2006; Donohue et al., 2016; Hamilton et al., 2016). This correspondence is consistent with the release of available potential energy during an eddy shedding event (Chassignet and Boudra, 1988; Fig. 8e,f). For barotropic energy conversions associated with the 60-day window, the values are predominantly positive during isolated detachment events in the upper 1000 m (Fig. 8a,b). However, in cases of consecutive detachment events, such as those in 1999, 2005, and 2013, negative values are also observed. These indicate a transfer from the EKE to the mean KE defined by the 60-day window. For the 180-day window, barotropic energy conversions are mostly positive, indicating EKE production even during consecutive detachment events (Fig. 8a,b).

When using a 60-day window, baroclinic energy conversions in the upper 1000 m are stronger and more closely associated with EKE production than when using the 180-day window. Although the 180-day window retains some of the conversions captured at 60 days, the reduced net baroclinic conversions in the 180-day window (Fig. 7b) reflect a change in the volume-integrated energy conversion balance between the eddy potential energy and the EKE in the upper 1000 m. This is due to the fact that the available potential energy depends strongly on the background state, which is representative of either the state before an event (a window shorter than 60 days) or the longer-term mean (a

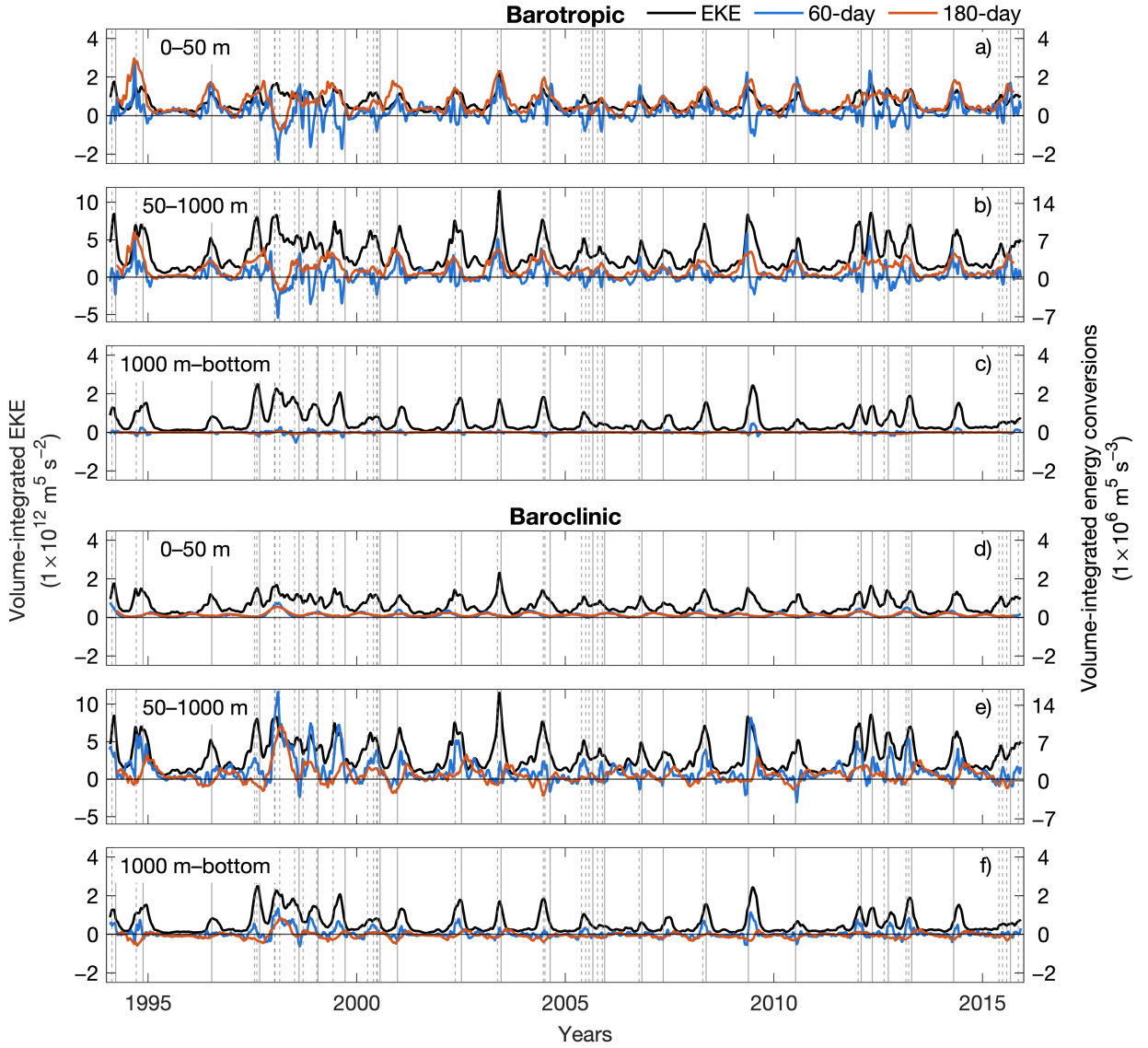


Figure 8: Temporal evolution of volume-integrated EKE and (a)–(c) barotropic ($K_m K_e$) and (d)–(f) baroclinic ($P_e K_e$) energy conversions over the eastern GoM in the reference simulation (REF) using the 60- and 180-day running-mean decompositions, partitioned into the (a,d) 0–50 m, (b,e) 50–1000 m, and (c,f) 1000 m–bottom layers. The black line denotes EKE calculated using the 60-day decomposition, similar to that derived from the 180-day window (not shown). The blue and orange lines represent energy conversions derived from the 60- and 180-day running-mean windows, respectively. The vertical light gray solid and dashed lines indicate LCE separation and reattachment events, respectively.

333 window longer than 180 days). When comparing the long-term EKE budget to the budgets derived from the various
 334 running-mean windows, we find that barotropic and baroclinic energy conversions computed with the 180-day window
 335 (Fig. 7e,f) closely match the long-term EKE budget and the energetics of the LC system (Fig. 6). Together, these
 336 results show that LCE shedding is linked to pronounced EKE peaks for both windows within the upper 1000 m,
 337 with baroclinic energy conversions contributing more to EKE production at shorter timescales and barotropic energy
 338 conversions at longer timescales. This behavior is consistent with previous studies that demonstrate that LCE shedding
 339 is governed by a combination of barotropic and baroclinic energy conversions/instabilities (Hurlburt, 1986; Chérubin
 340 et al., 2006). Although the instantaneous amplitude of baroclinic conversions exceeds that of barotropic conversions

during detachment events, barotropic conversions dominate the production of EKE over a long period of time (Garcia-Jove et al., 2016; Larrañaga et al., 2022).

It is also worth noting that there are distinct small peaks in baroclinic energy conversions below 1000 m during LCE detachment events for the 60-day window (Fig. 8f), whereas barotropic energy conversions remain negligible (Fig. 8c). EKE generation in the lower layer of the GoM is mainly driven by the pressure work that transfers energy from the upper 1000 m, as reported by Maslo et al. (2020) and Yang et al. (2020). With respect to the sensitivity of these results to bottom drag, the response to the choice of the window is similar among experiments, confirming that removing the deep bottom drag enhances barotropic conversions and suppresses baroclinic conversions, while the shallow Caribbean drag primarily reduces both conversion terms near the surface and regulates the KE flux entering the eastern GoM (Figs. 6c,e and 7c–f).

4. Impact of bottom drag on Loop Current penetration and eddy statistics

Having established how bottom drag alters the regional energetics, we now assess how these changes affect the physical behavior of the LC and its separated eddies. We first quantify LC penetration and the latitude of eddy separation and then analyze LCE properties, trajectories, and spatial occurrence. These diagnostics are then linked to the spatial distribution of surface EKE in the central and western GoM.

4.1. LC penetration and LCE separation latitude

Statistics of LC penetration are obtained from the spatial distribution of the LC outermost position, as in Larrañaga et al. (2022). In this paper, it is defined as the point along the 0.17-m LC contour with the greatest perpendicular distance to a transect connecting the Yucatan Channel (86.1°W, 21.6°N) and the Straits of Florida (81.5°W, 23.7°N). Using this metric, the LC is classified into three discrete states: retracted, canonical, and extended (Leben, 2005; Garcia-Jove et al., 2016; Larrañaga et al., 2022). The LC states are identified using longitude thresholds at 86°W and 89°W: retracted east of 86°W, canonical between 86°W and 89°W, and extended west of 89°W. Fig. 9 shows the probability density function (PDF) of the LC outermost position and the number of LCE separations binned by latitude. The histograms to the right of and below each panel summarize the latitudinal and longitudinal distributions of LC penetration, respectively. Detachment events are visually identified when the 0.17-m contour breaks into two independent contours, one corresponding to the LC and the other to the detached LCE, and detachment locations are defined by the LC coordinate closest to the periphery of the detached LCE (Larrañaga et al., 2022, 2026). A detachment event is classified as a separation event if the LCE does not reattach to the LC and subsequently propagates westward.

In the reference experiment (REF), the retracted form dominates (60.1%) with a relative maximum near the Dry Tortugas Islands, followed by the canonical (37.4%) and extended (2.5%) forms (Fig. 9a). The LC extends north of 27°N only 6.3% of the time. The LCE separation latitudes are mostly south of 26°N, with no events above 26°N (Fig. 9f). In NoBD, the LC spends more time in the retracted form than in REF and almost never reaches the extended form (0.3%; Fig. 9b). This is consistent with the number of separations, which increases from 8 to 12 below 24°N and decreases from 7 to 2 above 25°N (Fig. 9f). A similar but weaker response is observed in NoBD_lower (Fig. 9c). When compared to REF, the number of separations below 24°N increases by only 25% (instead of 50% in NoBD), whereas the number above 25°N decrease by about 40% (instead of 70% in NoBD; Fig. 9f). In NoBD_upper, LCE separations are even more concentrated at southern latitudes, with 68% occurring south of 24°N (Fig. 9f). Accordingly, the penetration PDF shows a roughly 5% increase in the retracted form (Fig. 9d). Adding back bottom drag to NoBD in the Caribbean Sea regions shallower than 1000 m ($\text{GoM}_{\text{NoBD}}\text{CS}_{\text{NoBD_lower}}$) drastically increases the number of separations north of 25°N and brings it back to the REF value (Fig. 9f). This further illustrates how upstream conditions can regulate the LC dynamics and energetics (see Section 3).

Overall, these results illustrate how bottom drag affects both the latitude of LCE separations and the extent of LC penetration. We find that removing the bottom drag in regions shallower than 1000 m, especially in the Caribbean Sea, has the strongest impact on the LC, with more frequent low-latitude separations and more time spent in the retracted phase. In the following subsection, we evaluate the impact of these changes on LCE properties and trajectories.

4.2. LCE properties, trajectories, and their relation to the LC EKE tongue

Following detachment, the EKE in the central GoM is shaped by the properties and trajectories of westward-propagating LCEs (Oey, 1996; Hamilton et al., 2019; Larrañaga et al., 2022). The intensity and spatial extent of the LC EKE tongue depend not only on how often LCEs separate but also on their KE and propagation pathways. To

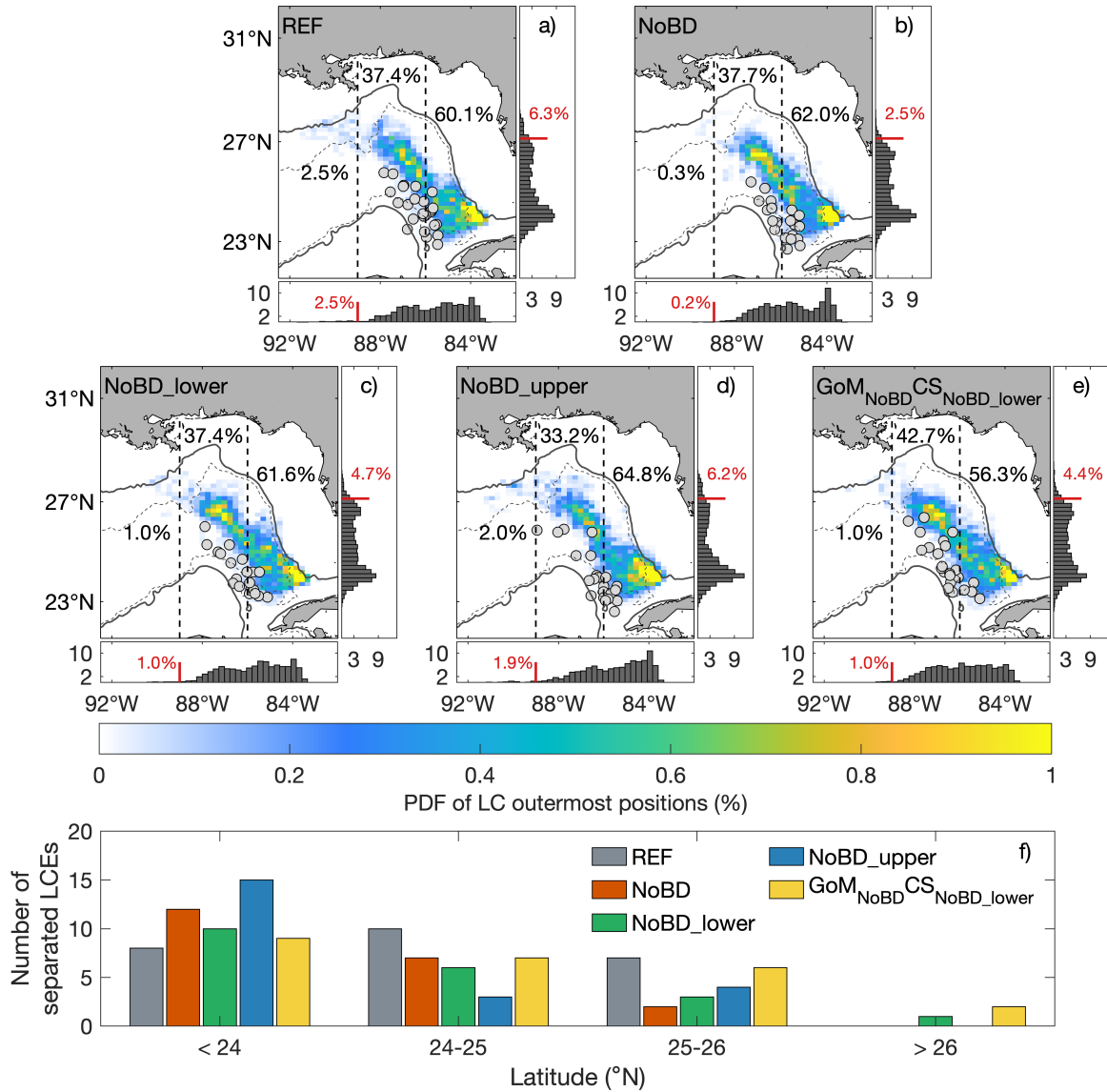


Figure 9: Probability density functions of the LC outermost positions (%). The gray circles indicate the locations of separation events. The black vertical segmented lines at 86°W and 89°W partition the LC into retracted, canonical, and extended forms. The gray contours represent the 1000 (continuous) and 2500 m (dashed) isobaths. Histograms on the right and bottom of each panel summarize the latitudinal and longitudinal distributions of LC penetration. The red labels denote the fractions beyond 27°N and 89°W. (f) Number of separated LCEs binned by separation latitude.

assess how bottom drag modulates these characteristics, we analyze the LCE separation frequency, intensity, size, and trajectories for all simulations (Figs. 10 and 11). As in previous studies, trajectories are depicted for eddies with a lifetime longer than 200 days (Chelton et al., 2011; Sosa-Gutiérrez et al., 2020; Larrañaga et al., 2022).

In the reference experiment (REF), 25 separation events occur with a mean KE of $57 \text{ m}^3 \text{ s}^{-2}$ (Fig. 10). After detachment, LCEs propagate westward between 23° and 26°N, and the crossings of 92°W are broadly distributed within this latitude range (Fig. 11a,f). This latitudinal spread contributes to a zonally extended LC EKE tongue across the central GoM (Fig. 11a). In NoBD, the number of separations decreases by 16% (21 events), while the separated eddies are 17% more energetic (Fig. 10). The eddy crossings north of 25°N drop from 7 in REF to 3, while the total crossings below 25°N remain comparable to REF (Fig. 11f). This decrease in high-latitude tracks shifts the LC

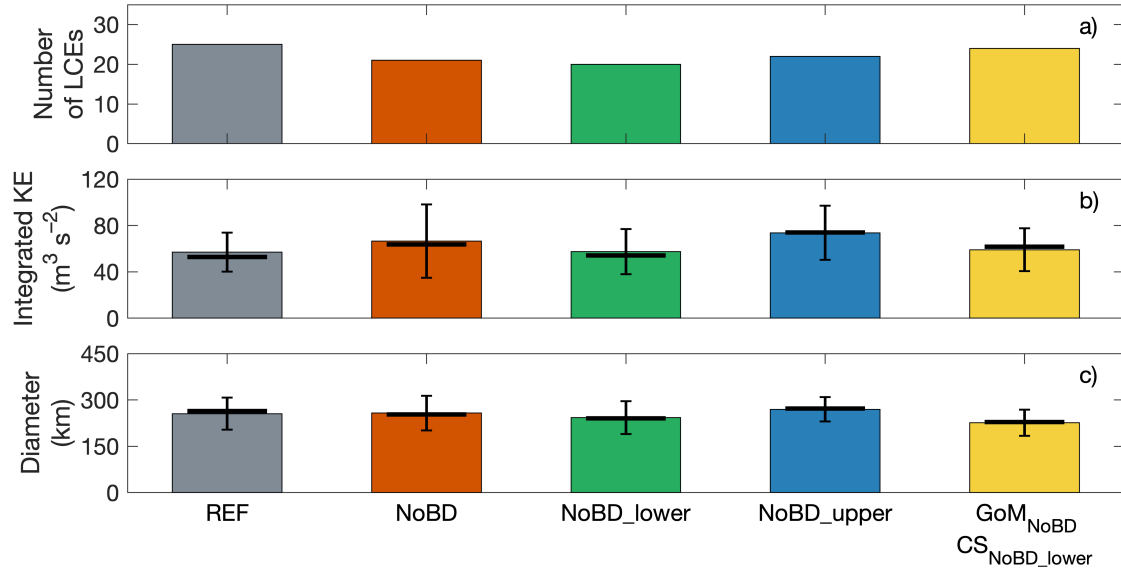


Figure 10: LCE properties at separation. (a) Number of separation events, (b) 0–200 m depth-integrated kinetic energy (KE; $\text{m}^3 \text{s}^{-2}$), and (c) diameter (km). Bars show the mean, error bars denote the standard deviation, and horizontal ticks mark the median.

EKE tongue slightly southward (Fig. 11b). However, the reduced separation frequency outweighs the increase in eddy intensity, leading to a $\sim 20\%$ decrease in surface EKE in the central GoM (Fig. 3g). In NoBD_lower, eddy crossings are nearly identical to those in REF across all latitude ranges (Fig. 11f). The number of separated eddies decreases by 20%, but their KE remains comparable to REF (Fig. 10). With similar eddy energetics but fewer separations, the LC EKE tongue weakens (Fig. 11c), and surface EKE in the central GoM is reduced by $\sim 30\%$ compared to REF (Fig. 3g). In contrast, NoBD_upper increases crossings south of 24°N by 22% (from 9 in REF to 11; Fig. 11f), consistent with the concentration of separations below 24°N (Fig. 9f). Combined with LCEs that are about 30% more energetic (Fig. 10b), NoBD_upper generates a stronger southward-displaced LC EKE tongue (Fig. 11d), resulting in a 13% increase in surface EKE in the central GoM with respect to REF (Fig. 3g). To determine whether these changes in LCE trajectories and intensity are driven by local GoM dynamics or upstream forcing, we examine the impact of restoring shallow bottom drag in the Caribbean. When bottom drag is restored in the Caribbean Sea in NoBD, both the mean KE and diameter of separated LCEs are reduced by $\sim 10\%$ (Fig. 10b,c). Restoring upstream drag also modulates the trajectories: the southward shift of trajectories is less pronounced, with the number of eddy crossings decreasing south of 24°N and increasing north of 25°N (Fig. 11f). Together, these changes realign the westward LCE pathways, leading to a less southward-tilted and weaker LC EKE tongue (Fig. 11e).

In summary, across configurations with bottom drag removed throughout the domain or in regions shallower or deeper than 1000 m, the number of separations decreases relative to the reference case, while LCE size remains largely unchanged. Removing bottom drag only in regions deeper than 1000 m yields modest changes in LCE intensity and pathways, whereas configurations with bottom drag absent in regions shallower than 1000 m produce more energetic LCEs and southward-shifted trajectories. However, these upper-layer responses remain highly sensitive to upstream frictional damping in the Caribbean Sea. Consequently, bottom drag modulates the structure of the LC EKE tongue: its magnitude is mainly governed by LCE separation frequency, and its meridional position is controlled by the LCE propagation pathways.

4.3. Impact of LCE trajectories and properties on western GoM circulation

LCEs play a central role in shaping mesoscale circulation in the western GoM, redistributing energy as they propagate westward from the LC region (Hamilton et al., 1999; Vukovich, 2007; Olvera-Prado et al., 2024). We now assess how the changes in LCE properties and trajectories due to bottom drag project into the western Gulf circulation. To diagnose how bottom drag modulates the western GoM energy field, we analyze the spatial distribution of LCEs

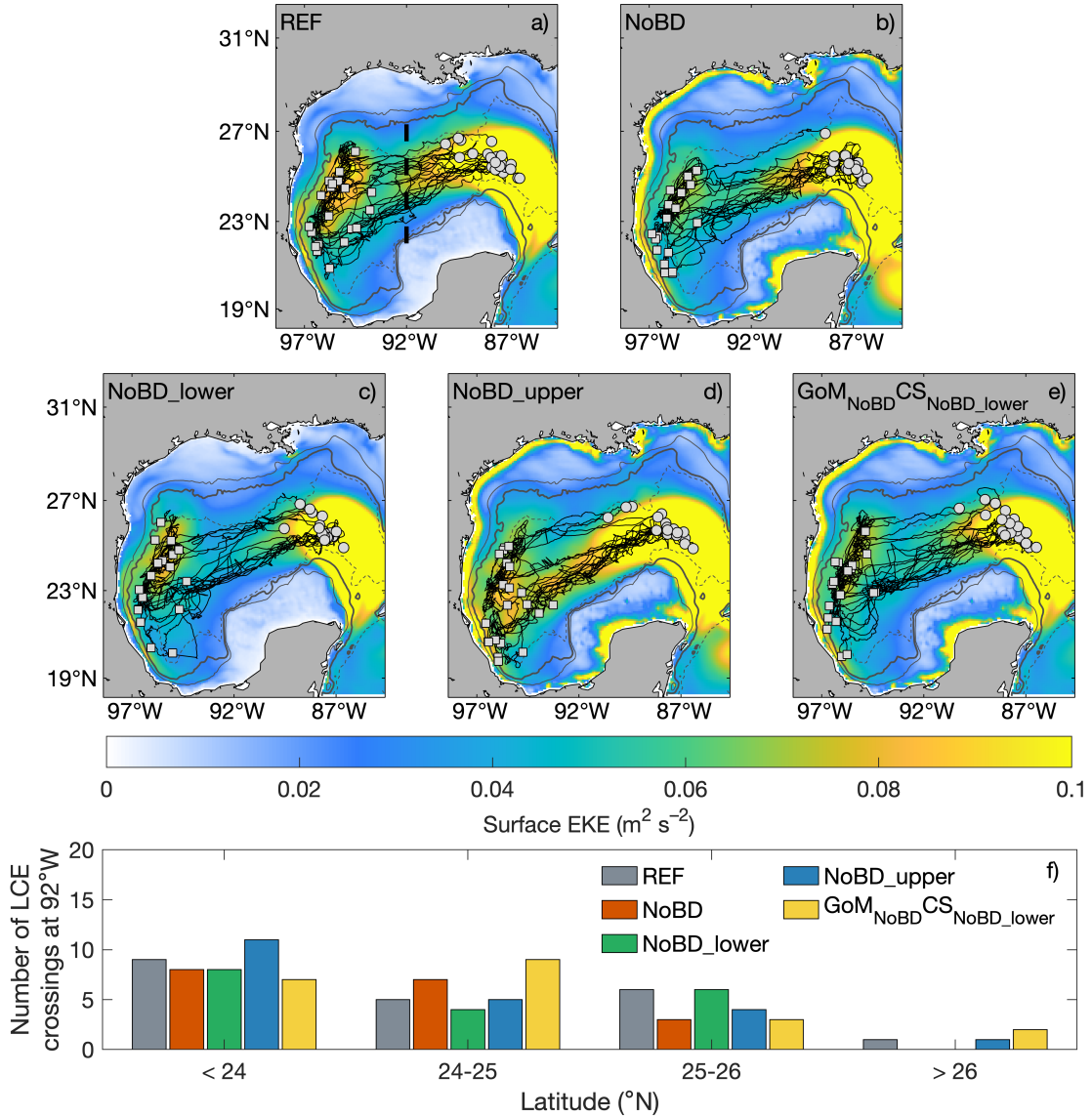


Figure 11: Trajectories of LCEs with a lifetime longer than 200 days. The gray squares denote the centers of the LCEs upon separation, and gray circles indicate their centers when they are no longer detected. The color map shows the surface EKE. To better highlight differences in surface EKE over the central and western GoM, the color scale is limited to 0–0.1 $\text{m}^2 \text{s}^{-2}$, with higher values shown as saturated colors. The gray contours represent the 200 (thin), 1000 (thick), and 2500 m (dashed) isobaths. (f) Number of LCEs crossing the 92°W transect shown in (a), using 1° latitude bins ($\pm 0.5^\circ$ about the bin center).

with a lifetime longer than 200 days. In Fig. 12, the 10% and 20% occurrence contours are displayed over the surface EKE for all experiments. Across all simulations, the 10% and 20% LCE occurrence contours generally coincide with regions of stronger surface EKE; however, occurrence frequency alone does not account for the differences in EKE between experiments in the western GoM. We therefore examine three LCE properties within the western GoM: the number crossing the 93°W transect, intensity, and residence time (Fig. 13). Intensity and residence time are calculated from the first crossing of 93°W until the LCE is no longer detected.

In the reference experiment (REF), the 10% LCE occurrence contour extends westward from the eastern GoM across the central GoM (Fig. 12a), consistent with eddy trajectories (Fig. 11a). The occurrences are concentrated

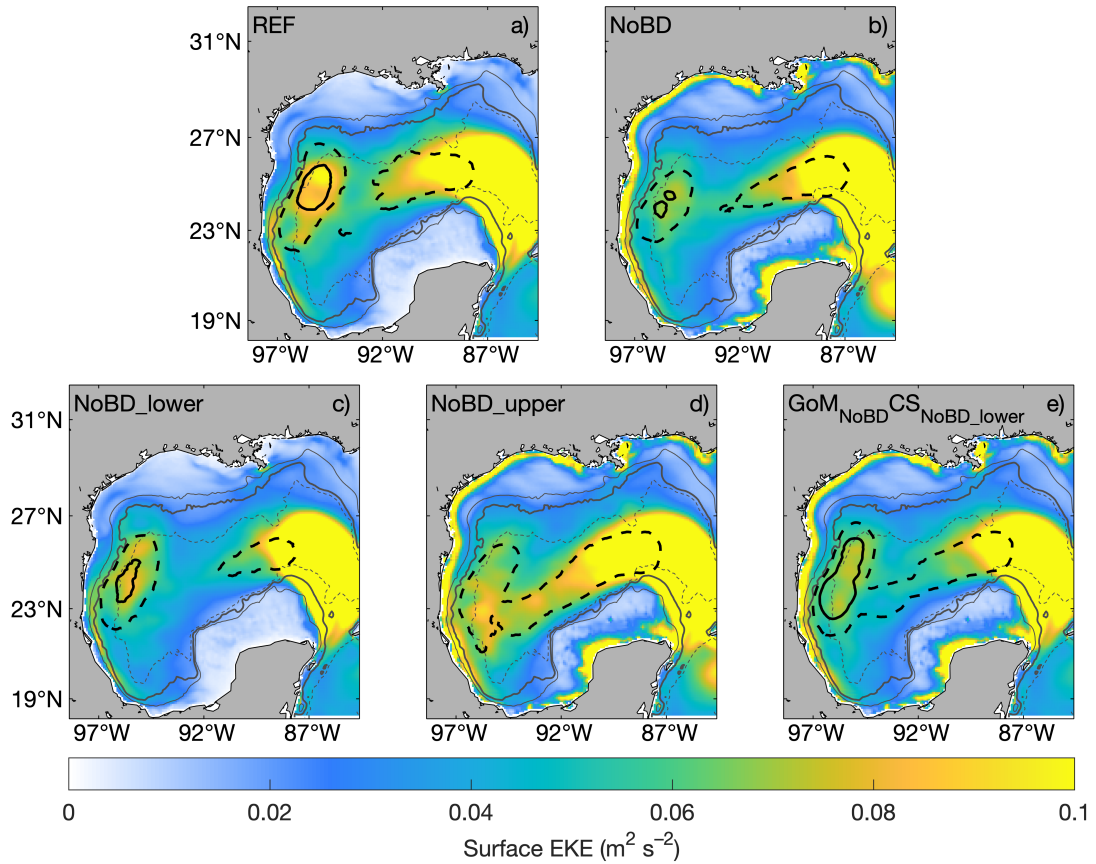


Figure 12: Spatial distribution of LCEs with a lifetime longer than 200 days. The black contours represent the probability density function of LCE occurrences. The black dashed and solid contours indicate the 10% and 20% occurrence thresholds, respectively. The color map shows the surface EKE. To better highlight differences in surface EKE over the central and western GoM, the color scale is limited to 0–0.1 $\text{m}^2 \text{s}^{-2}$, with higher values shown as saturated colors. The gray contours represent the 200 (thin), 1000 (thick), and 2500 m (dashed) isobaths.

between 24° and 26°N, with the 20% contour enclosing the EKE maximum. A total of 21 LCEs cross 93°W into the western GoM, with a mean KE of $32.7 \text{ m}^3 \text{s}^{-2}$ and a mean residence time of 309 days (Fig. 13). From REF to NoBD, the 10% and 20% occurrence contours contract, indicating a distinctly reduced LCE footprint (Fig. 12b). The number of LCEs entering the western GoM in NoBD decreases from 21 in REF to 18, and both the intensity and residence time decrease by 17% (Fig. 13). Collectively, fewer, less energetic, and shorter-lived LCEs weaken the cumulative eddy imprint, reducing the western GoM surface EKE by ~15% relative to REF (Fig. 3g). In NoBD_lower, the 10% occurrence contour is spatially similar to that in REF. However, the area enclosed by the 20% contour is smaller (Fig. 12c), reflecting about 14% fewer LCEs with 12% lower KE than REF in the western basin (Fig. 13). As a result, surface EKE in the western GoM decreases by 10% (Fig. 3g). In NoBD_upper, overall LCE properties in the western GoM are similar to those in REF (Fig. 13). However, the 20% occurrence contour is absent in the western GoM, while the 10% contour remains continuous from the eastern GoM into the western basin, reaching as far south as 21°N (Fig. 12d). This pattern arises because LCEs preferentially follow southward-shifted trajectories (Fig. 11d), which in turn concentrates surface EKE in the southwestern GoM. The spatial distribution of surface EKE in the western GoM is also sensitive to the Caribbean bottom drag configuration. From NoBD to GoM_NoBD_CS_NoBD_lower, the 10% and 20% occurrence contours expand in the northwestern GoM (Fig. 12e), as the number of eddies and their residence time in the western GoM increase by about 17% and 34%, respectively, relative to NoBD (Fig. 13a,c). However, the surface EKE within the 20% contour is still weaker than in REF, as the eddies reaching the western basin are still 27% less energetic (Fig. 13b).

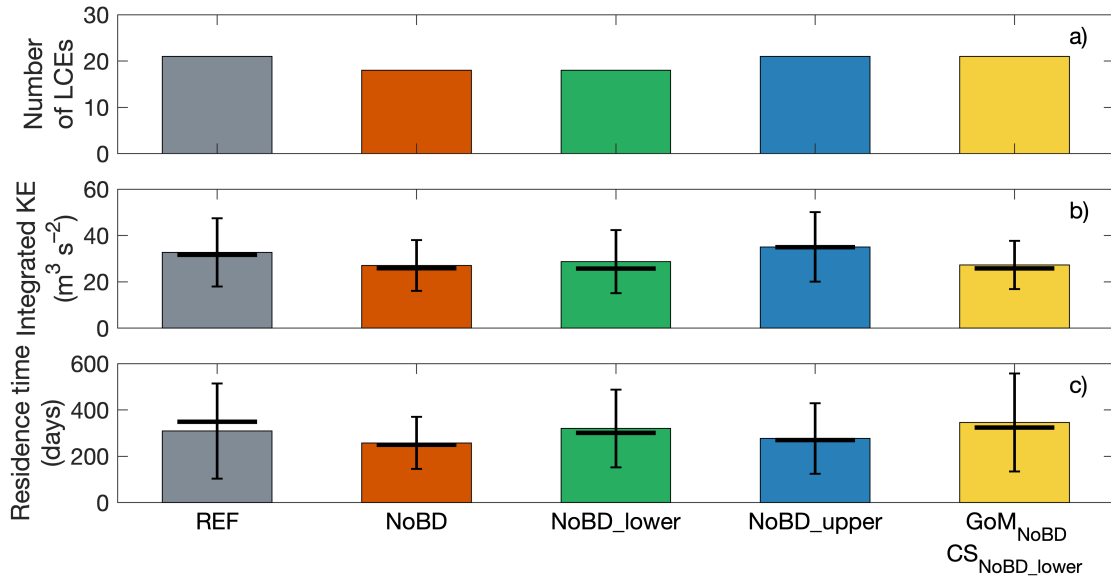


Figure 13: Properties of LCEs with a lifetime longer than 200 days in the western GoM. (a) Number of LCEs crossing 93°W , (b) 0–200 m depth-integrated KE ($\text{m}^3 \text{s}^{-2}$), and (c) residence time (days). KE and residence time are computed from the first crossing of 93°W until each LCE is no longer detected. Bars show the mean, error bars denote the standard deviation, and horizontal ticks mark the median.

The surface EKE in the western GoM reflects three interacting factors: how many LCEs enter, how energetic they are, and how long they persist. When bottom drag is removed in the domain or only in regions deeper than 1000 m, fewer LCEs enter the western basin, and their intensity is lower than in the reference case, resulting in a less extensive 20% occurrence contour and a reduced surface EKE in the western GoM. In contrast, removing bottom drag only in regions shallower than 1000 m produces LCE properties close to the reference, but displaces the EKE maximum southward, consistent with preferentially south-shifted trajectories. Additionally, the Caribbean bottom drag experiment demonstrates that upstream damping plays a critical role in regulating the energy and trajectory of LCEs, dictating how and where they dissipate in the western basin.

5. Discussion and conclusions

This study quantifies how bottom drag regulates EKE and GoM dynamics using a suite of HYCOM numerical simulations. Bottom drag acts not only as a direct sink of kinetic energy but also as an indirect regulator of instability pathways by modifying horizontal and vertical shear. These dual roles modulate the regional energy budgets, LC variability, and LCE statistics across the eastern, central, and western GoM. Bottom drag therefore emerges as an important control on mesoscale dynamics in the GoM, linking boundary-layer friction to basin-scale circulation structure. The GoM circulation can be viewed as a two-layer system in which most oceanic energy is concentrated within the upper 1000 m (Hamilton et al., 2018; Furey et al., 2018). Accordingly, the dynamical role of bottom drag differs fundamentally between regions shallower and deeper than 1000 m.

An unexpected result is the large increase in surface EKE in the eastern GoM when bottom drag is set to zero throughout the domain or only in regions shallower than 1000 m. This response is associated with enhanced mesoscale activity in the shallow Gulf of Batabanó, south of Cuba, that propagates into the GoM. Adding back bottom drag only in the shallow Caribbean Sea indeed removes the near-surface intensification over the eastern GoM, returning the surface EKE toward the reference levels.

Long-term EKE budgets provide a framework for identifying the dominant energy conversion pathways, and we find that, for all experiments, barotropic energy conversions act as the main EKE source in agreement with previous studies (Garcia-Jove et al., 2016; Larrañaga et al., 2022). When the bottom drag is removed in regions deeper than 1000 m, the lower-layer currents intensify, weakening the vertical shear but also strengthening the horizontal shear. This leads to an increase in barotropic energy conversions while simultaneously suppressing baroclinic energy

conversions, resulting in a ~15% increase in EKE in the eastern GoM. While the long-term EKE budget characterizes the net energy conversion pathways, it does not capture their temporal variability during episodic events such as LCE detachments. We therefore examine the energetics associated with these events using running-mean decompositions of the energy budget. Overall, the response is similar in all experiments. In the upper 1000 m, baroclinic conversions contribute substantially more than barotropic conversions to EKE production when using shorter time windows (less than 90 days), whereas barotropic conversions become dominant at windows longer than 180 days. When using a 180-day window, the time-averaged conversions more closely resemble those obtained from the long-term Reynolds decomposition, indicating that a 180-day running average is representative of the long-term mean. The differences in energy conversions between the short and long time windows are particularly evident during LCE detachment events since the variability captured by each time window is strongly dependent on its respective mean state. In the lower layer, barotropic energy conversions are negligible for all time windows, while small baroclinic energy conversions can be observed during LCE detachment events. On average, both conversions are small, and most lower-layer EKE is a consequence of vertical pressure work variability as reported by Maslo et al. (2020) and Yang et al. (2020).

The sensitivity of EKE to bottom drag is not uniform across the basin. For example, removing bottom drag in the domain leads to an increase in EKE over the eastern GoM, but a decrease in the central and western regions. It also favors a more retracted LC state due to an increase in the number of southern LCE separations. The latter is directly tied to the enhanced mesoscale activity south of Cuba. This is in agreement with Le Hénaff et al. (2012) and Garcia-Jove et al. (2016), who showed that reduced mesoscale variability in the Caribbean Sea is associated with increased LC penetration into the GoM. The strength of the LCEs is strongly dependent on whether bottom drag is removed in shallow or deep regions. When the bottom drag is set to zero in regions shallower than 1000 m, the LCEs are about 30% more energetic than in the reference simulation, consistent with the more energetic inflow from the Caribbean Sea. The LCE trajectories are also displaced southward. Restoring the bottom drag in the shallow Caribbean Sea reduces the upstream energetic intensification and brings LCE intensity and trajectories closer to the reference state. When bottom drag is removed in regions deeper than 1000 m, the number of LCE separations decreases relative to the reference simulation, resulting in a weaker LC EKE tongue signature, the latter being representative of the cumulative imprint of LCE separation frequency, intensity, and pathway. In all experiments, the EKE response in the western GoM reflects the cumulative impacts of the LCEs that reach the western region.

While the primary objective of this study was to evaluate the role of bottom drag in modulating GoM dynamics, the analysis also identified several aspects where our numerical simulations need improvement. Specifically, in our reference simulation, we find that the LC remains retracted around 60% of the time, compared with approximately 30% in the 24-year altimetry record (Larrañaga et al., 2022). It also exhibits more LCE separations than observed south of 25°N and fewer between 25° and 26°N. These systematic differences suggest that missing physical processes or limitations in model forcing and resolution affect the representation of the LC system. One possible source of this discrepancy is the omission of surface-current feedback to the atmosphere. Larrañaga et al. (2022) showed that including current feedback partially reduces biases in LC-state occurrence and LCE separation statistics, although the extended LC state remains underestimated. Another source may be the model grid spacing, since a 1/12° configuration does not allow for an explicit representation of submesoscale processes and may not be energetic enough (Chassignet and Xu, 2017). Future work should therefore investigate the combined roles of bottom drag, ocean–atmosphere feedback, and model resolution in controlling LC and LCE variability. Such experiments would help distinguish the effects of bottom-boundary dissipation from those associated with air–sea coupling and unresolved fine-scale dynamics, providing a more complete understanding of the GoM energy budget.

Overall, this study demonstrates that bottom drag functions both as a dissipation mechanism and a key regulator of mesoscale variability in the GoM. Therefore, an accurate representation of bottom drag in numerical models is essential for realistically simulating the large-scale circulation and energy budget and, consequently, improving operational ocean forecasts. Efforts are underway to derive bottom drag parameterizations that vary regionally and are a function of the observed bottom roughness (Radko et al., 2026).

CRediT authorship contribution statement

Ilkyeong Ma: Writing – original draft, Writing – review & editing, Conceptualization, Methodology, Software, Visualization, Validation, Investigation, Formal analysis, Data curation. **Marco Larrañaga:** Writing – review & editing, Conceptualization, Methodology, Software, Validation, Investigation. **Eric P. Chassignet:** Writing – review & editing, Supervision, Conceptualization, Methodology, Investigation, Funding acquisition. **Alexandra Bozec:** Writing

– review & editing, Methodology, Software, Data curation. **Takaya Uchida:** Writing – review & editing, Funding acquisition.

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.

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Data availability

All data presented in this study are available on Figshare at <https://doi.org/10.6084/m9.figshare.33342390>.

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