Three-Dimensional Model-Observation Comparison in the Loop Current Region

K. C. Rosburg^{a,1,*}, K. A. Donohue^a, E. P. Chassignet^b

^a*Graduate School of Oceanography, University of Rhode Island, Narragansett, Rhode Island, USA*

^b*Center for Ocean-Atmosphere Prediction Studies, Florida State University, Tallahassee, Florida, USA*

Abstract

Accurate high-resolution ocean models are required for hurricane and oil spill pathway predictions, and to enhance the dynamical understanding of circulation dynamics. Output from the 1/25° data-assimilating Gulf of Mexico HYbrid Coordinate Ocean Model (HYCOM31.0) is compared to daily full water column observations from a moored array, with a focus on Loop Current path variability and upper-deep layer coupling during eddy separation. Array-mean correlation was 0.93 for sea surface height, and 0.93, 0.63, and 0.75 in the thermocline for temperature, zonal, and meridional velocity, respectively. Peaks in modeled eddy kinetic energy were consistent with observations during Loop Current eddy separation, but with modeled deep eddy kinetic energy at half the observed amplitude. Modeled and observed LC meander phase speeds agreed within 8% and 2% of each other within the

*Corresponding author

Email addresses: `rosburk@my.uri.edu` (K. C. Rosburg), `kdonohue@uri.edu` (K. A. Donohue), `echassignet@fsu.edu` (E. P. Chassignet)

¹Now at: School of Ocean and Earth Science and Technology, University of Hawaii at Manoa, Honolulu, Hawaii, USA (`krosburg@hawaii.edu`)

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100–40 and 40–20 day bands, respectively. The model reproduced observed patterns indicative of baroclinic instability, that is, a vertical offset with deep stream function leading upper stream function in the along-stream direction. While modeled deep eddies differed slightly spatially and temporally, the joint development of an upper-ocean meander along the eastern side of the LC and the successive propagation of upper-deep cyclone/anticyclone pairs that preceded separation were contained within the model solution. Overall, model-observation comparison indicated that HYCOM31.0 could provide insight into processes within the 100–20 day band, offering a larger spatial and temporal window than observational arrays.

Keywords: Evaluation, Modelling, Ocean currents, Mesoscale eddies, Baroclinic instability, USA, Gulf of Mexico, Loop Current

1. Introduction

As part of the North Atlantic subtropical western boundary current system, the Loop Current (LC) enters the Gulf of Mexico (GOM) from the Caribbean Sea as the continuation of the Yucatán Current (YC), circulates anticyclonically within the Gulf forming a large loop, exits through the Florida Straits, and becomes the Florida Current after turning north along the eastern side of Florida. On irregular intervals, between 3–17 months, a large (200–400 km diameter) anticyclonic eddy, a LC Eddy (LCE), separates from the LC (Sturges and Leben, 2000; Dukhovskoy et al., 2015). The separation process, shown schematically in Figure 1, begins with the northward intrusion of the LC into the GOM, followed by the necking down of the LC and eventual pinching-off of a LCE. After separation, the LC re-

13 treats southward to the so-called port-to-port mode while the newly shed
 14 LCE propagates westward across the Gulf.

Figure 1: Maps of sea surface height depicting the three-stage Loop Current Eddy cycle: (a) northward intrusion/growth of the Loop Current (LC), (b) pinch-off of the anticyclonic ring, and (c) final separation and subsequent westward propagation of the eddy, and retreat of the LC to port-to-port mode. FC is Florida Current. YC is Yucatan Current. Sea surface height from the 1/25° Gulf of Mexico Hybrid Coordinate Ocean Model, GOM10.04 expt_31.0.

15 There is a strong need for predictive skill for LCE separation. For exam-
 16 ple, strong currents associated with the LC and LCEs, as well as the strong
 17 deep currents generated during LCE separation, are hazardous to deep-water
 18 oil drilling operations. The warm cores of LCEs are also known to modify
 19 the intensity of passing hurricanes (e.g. Cione and Uhlhorn 2003; Yablonsky
 20 and Ginis 2012; Lin et al. 2008). Deep circulation, especially along the steep
 21 escarpments of the Gulf’s continental slope play an important role in the
 22 rapid dispersal of contaminants (e.g. Paris et al. 2012; Nguyen et al. 2015).

23 Efforts have been made to predict and model LCE separation. Using
 24 an idealized vorticity model, Lugo-Fernández and Leben (2010) confirmed
 25 a linear relationship between the latitude of LC retreat and the length of
 26 time between LCE separations, a trend previously seen in satellite altime-

27 try (Leben, 2005). Maul (1977) hypothesized a linkage between the rate
 28 of change of LC volume and deep transport through the Yucatán Channel.
 29 This idea is supported by 7.5 months of YC mooring observations (Bunge
 30 et al., 2002) and the recent analysis of a 54-year free-running $1/25^\circ$ model
 31 (Nedbor-Gross et al., 2014). Chang and Oey (2011), on the other hand, sug-
 32 gest that mass exchange between the eastern and western basins, as well as
 33 exchange between the LC and deeper waters, play a significant role in the
 34 separation process. Evidence has been found for both seasonal (Leben et al.,
 35 2012; Chang and Oey, 2012) and inter-annual (Lugo-Fernández, 2007) trends
 36 in the length of the eddy separation period. Recent modeling studies suggest
 37 that seasonality in the trade winds may affect LCE separation (e.g. Chang
 38 and Oey 2013; Xu et al. 2013). Using an artificial neural network approach,
 39 Zeng et al. (2015) achieved reliable LCE shedding forecasts of up to four
 40 weeks in SSH. Numerical studies also point to the importance of instability
 41 processes, the coupling between upper and deep circulation, and the gen-
 42 eration of bursts of strong deep eddies during LCE separation. Examining
 43 instabilities exhibited in upper and deep pressure fields of a two-layer model,
 44 Hurlburt and Thompson (1980, 1982) found deep circulation driven by mixed
 45 baroclinic and barotropic instabilities. During LCE separation and detach-
 46 ment events, deep circulation is dominated by a field of intense deep eddies
 47 that propagate and couple with vortices of the upper-ocean LC (Sturges et al.,
 48 1993; Chérubin et al., 2005). Baroclinic instabilities near Campeche Bank
 49 and the West Florida Shelf have also been identified as a possible mechanism
 50 for the generation of deep eddies that facilitate LCE detachment (Chérubin
 51 et al., 2005; Oey, 2008). Finally, Le Hénaff et al. (2012) suggest that deep

eddy spin up as the LC moves off the Mississippi Fan. How well numerical models predict or simulate deep currents is not well documented owing to sparse observations of circulation below the surface and in particular below the thermocline.

Figure 2: Map of DynLoop mooring array, indicating locations of tall-moorings (gray filled stars), near-bottom moorings (triangles), and PIES (black filled circles), along with satellite altimeter exact repeat ground track coverage for OSTM/Jason-2 (solid) and Jason-1 tandem mission (dashed), as well as bathymetry (gray contours) at 500 m intervals.

In 2009, a comprehensive field study “Observations and Dynamics of the Loop Current” (DynLoop) was undertaken. Funded by the Bureau of Ocean Energy Management (BOEM), DynLoop aimed to investigate LC circulation dynamics, eddy-shedding mechanisms, and forcing of deep flow. The study utilized an *in situ* mapping array centered in the LC (Figure 2) that included nine full water column (tall) moorings, seven near-bottom moorings, and 25

62 pressure sensing inverted echo sounders (PIES). The array provides a unique
63 dataset for studying the LCE cycle: it was centered in the region of LCE for-
64 mation/separation and during its 30-month deployment observed four LCE
65 events with daily measurements throughout the water column at mesoscale
66 resolution. The dataset from this study provides critical deep-velocity infor-
67 mation required for a comprehensive 3D model-data comparison. Hamilton
68 et al. (2016) provides a review of the study.

69 Through advances in modeling, advanced assimilation techniques, and
70 increased computational power, modern predictive ocean models reproduce
71 surface currents to a high degree of accuracy. One example is the HYbrid
72 Coordinate Ocean Model (HYCOM). Because of the demonstrated applica-
73 tion of global- and basin-scale real time ocean predictions, the US Navy has
74 transitioned HYCOM into operational use at the Naval Oceanographic Office
75 (NAVOCEANO; Chassignet et al. 2009; Cummings and Smedstad 2013; Met-
76 zger et al. 2014). The high-resolution $1/25^\circ$ regional-scale data-assimilative
77 GOM HYCOM has undergone a number of improvements; the current ver-
78 sion (at the time of writing), GOMl0.04 expt_31.0 (hereafter HYCOM31.0)
79 is one of the highest resolution and most advanced data-assimilative nu-
80 merical models available for studies and predictions of GOM circulation.
81 HYCOM31.0 assimilates predominately surface measurements from remotely
82 sensed satellite altimetry and temperature, as well as temperature and salin-
83 ity profiles, but does not incorporate deep (> 2000 m) observations. Previous
84 validation of HYCOM includes comparison to other models, satellite SST,
85 SSS (salinity), SSH, and ocean color (Chassignet et al., 2005, 2007, 2009), to
86 satellite-tracked surface drifters (Liu and Weisberg, 2011; Liu et al., 2014),

87 and to airborne profiles of near-surface temperature and 20 °C isotherm depth
88 (Shay et al., 2011). Scott et al. (2010) did compare global HYCOM ocean
89 forecasting systems to a global current meter record dataset that included
90 observations below 2000 m depth, but comprehensive comparisons to deep
91 (> 2000 m depth) observations in the GOM are lacking. Other recent assim-
92 ilation efforts, Kantha et al. (2005); Yin and Oey (2007); Xu et al. (2013);
93 Gopalakrishnan et al. (2013b), have been made in the Gulf of Mexico. A
94 comprehensive review is beyond the scope of this study, which focuses upon
95 HYCOM31.0.

96 The overarching goal of this study is to assess the viability of HYCOM31.0
97 for use in studies of mesoscale LC processes. Here, we focus on two aspects:
98 LC path variability and vertical coupling between the upper and deep cir-
99 culation during LCE separation. The term LC Frontal Eddy (LCFE) has
100 been applied to describe variability along the LC path (see Le Hénaff et al.
101 2014 for a comprehensive review). LCFEs are thought to play a role in LCE
102 separation (e.g. Cochrane 1972; Chérubin et al. 2005; Schmitz 2005). In SST
103 and SSH, this variability appears as LC meanders and cyclonic eddy-like fea-
104 tures that propagate along the LC path (e.g. Walker et al. 2003). Here, we
105 choose to term variability along the LC path as “LC meanders” rather than
106 LC Frontal Eddies to reinforce the concept that the rich variability along the
107 LC path encompasses a wide range of spatial and temporal scales, and more
108 importantly that multiple dynamical processes are likely responsible for the
109 variability.

110 The DynLoop analysis of LC meanders determined that within the mesoscale
111 band (3–100 day periods), wavelengths are between 230 km to 460 km with

112 phase speeds ranging between 8 to 50 km d⁻¹ (Donohue et al., 2015). More-
113 over, that study and Le Hénaff et al. (2014) demonstrate that variability
114 is strongest for periods between 40 and 100 days. Long-wavelength low-
115 frequency meanders were found to be restricted to east of LC, corroborating
116 the early findings of Vukovich (1988) and the recent analysis of Le Hénaff
117 et al. (2014). These long-wavelength meanders form along the eastern edge
118 of the LC prior to eddy separation. Development of the upper meander is ac-
119 companied by elevated deep eddy kinetic energy and the formation of a deep
120 cyclone (anticyclone), which leads the upper-ocean meander trough (crest)
121 by roughly a quarter wavelength in a pattern consistent with baroclinic in-
122 stability (Donohue et al., 2016).

123 Observational studies are inherently limited both spatially and tempo-
124 rally, and numerical simulations provide the larger space and time window
125 required for a deeper dynamical understanding. For example, we ultimately
126 seek to determine what triggers the growth of long-wavelength low-frequency
127 meanders, the role of topography in stabilizing or destabilizing the LC, and
128 how topography dictates the pathways of the deep energy generated during
129 LCE formation. This preliminary 3D comparison is a necessary first step in
130 order to use the model for dynamical interpretation.

131 A detailed description of the observations, HYCOM31.0, and methodolo-
132 gies used in this study is provided in Section 2. Section 3 outlines the findings
133 of our time series and point-to-point statistical comparisons, followed by the
134 results of broad-scale spatial comparisons (SSH variance and EKE distribu-
135 tions) in Section 4. In Section 5, we present a phenomenological comparison
136 of a subset of the processes involved in the LCE cycle. The results of this

137 study are discussed in the broader context of the literature in Section 6.

138 **2. Data & Methods**

139 *2.1. Observations*

140 Observations derive from the comprehensive DynLoop field study in the
141 GOM, which included a large mooring array centered near 26°N 87°W (Fig-
142 ure 2). This array produced a unique dataset: the array, deployed for nom-
143 inally 30 months from April 2009 to November 2011, captured three LCE
144 separations and the initial detachment of a fourth LCE; the instrumentation
145 provided full water column observations; and the instrument spacing resolved
146 the mesoscale circulation. Details regarding the full suite of instrumentation
147 and processing are provided in Hamilton et al. (2014).

148 Nine tall moorings sampled the full water column. Point current meters
149 recorded velocities at 600, 900, 1300, and 2000 meters depth, with addi-
150 tional current meters located 100 meters above bottom (mab). Near-surface
151 currents were profiled by an upward-looking 75 kHz ADCP situated at 450
152 meters depth. Temperature sensors were located at 75, 150, 250, 350, 525,
153 600, 750, 900, 1100, 1300, 1500, and 2000 meters depth, as well as 100 mab.
154 Seven additional near-bottom moorings had a single current meter 100 mab.
155 Twenty-five PIES were deployed with a horizontal resolution of ~ 53 km.
156 PIES, moored at the ocean floor, record bottom pressure and the round trip
157 travel time, τ , of emitted 12 kHz sound pulses. Mooring velocity, tempera-
158 ture, τ , and bottom pressure were filtered with a 72-hour 4th order low-pass
159 Butterworth filter and subsampled at 24-hour intervals. A subset of PIES
160 and tall moorings were aligned along altimeter ground tracks (Figure 2).

161 Vertical profiles of temperature, salinity, and specific volume anomaly
 162 were calculated from τ using look-up tables (e.g., gravest empirical mode,
 163 GEM; Meinen and Watts 2000) constructed from historical hydrography.
 164 Donohue et al. (2015) reviews this methodology as applied to the GOM. The
 165 GEM tables extended from the surface to 3000 dbar. Geopotential at each
 166 PIES site was determined by integrating specific volume anomaly. Through
 167 optimal interpolation (OI; Bretherton et al. 1976), horizontal gradients of
 168 specific volume anomaly yielded mapped geostrophic velocity referenced to
 169 zero at the ocean bottom, nominally 3000 dbar. We term this field baroclinic
 170 referenced to the bottom or *bcb*.

171 As described in Donohue et al. (2010), the near-bottom pressure records
 172 were detided, dedrifted, and leveled. Here, leveled bottom pressures means
 173 bottom pressures that have been adjusted to the same absolute geopotential
 174 surface, nominally 3000 dbar. Simultaneous OI mapping of deep currents
 175 and pressure were used to provide a 3000 dbar reference velocity for the *bcb*
 176 geostrophic velocities. We term the deep 3000 dbar field reference or *ref*.

177 Absolute SSH was determined with PIES by combining a reference level
 178 sea surface height (SSH_{ref}), leveled 3000-dbar pressures converted to height
 179 (pressure divided by gravity and density), with baroclinic SSH referenced to
 180 the bottom (SSH_{bcb}), surface geopotentials referenced to 3000 dbar converted
 181 to height (geopotential divided by gravity). This methodology is well estab-
 182 lished (e.g., Baker-Yeboah et al. 2009; Park et al. 2012). Estimated PIES
 183 SSH error is 5.7 cm (Donohue et al., 2015). In this work, we use absolute
 184 SSH for the model comparisons. While the SSH_{ref} has important dynamic
 185 contributions, for the DynLoop PIES sites the variance of the SSH signal is

dominated by SSH_{bcb} : 98% of the total SSH variance and 96% of mesoscale band (100–3 day) SSH variance is due to variance in SSH_{bcb} .

The DynLoop array provides daily maps of temperature, density, sea surface height, and geostrophic velocity at mesoscale resolution. A thorough validation of the PIES methodology is provided in Hamilton et al. (2014) and Donohue et al. (2015). Here, we note that within the thermocline, the PIES captured more than 95% of the temperature variance, and RMS differences were small relative to signal size. Velocity comparisons within the thermocline revealed RMS differences less than 0.10 m s^{-1} .

The Colorado Center for Astrodynamics Research (CCAR) objectively mapped historical mesoscale altimeter data reanalysis product (Leben et al., 2002) was used to determine the position of the LC in the Gulf. The satellite altimeter data available for the historical reanalysis during the observational program included Jason-1, Envisat, and OSTM/Jason-2 satellite altimeters. Jason-1 tandem mission was operating during the program. Envisat transitioned from its nominal 35-day repeat orbit to a 30-day repeat orbit on 22 October 2010. A detailed description of the processing of the GOM SSH dataset can be found in Hamilton et al. (2014). Separation of LCEs from the LC was identified by the breaking of the 17-cm SSH contour in the CCAR GOM historical SSH data product. In this product, the 17-cm SSH contour closely tracks the LC (Leben, 2005).

2.2. Model

This study evaluates outputs from the data-assimilative GOM HYCOM expt_31.0. This particular model has $\sim 4 \text{ km}$ horizontal grid spacing at the latitude of the GOM ($1/25^\circ$) and uses 20 vertical coordinate surfaces. The

211 model uses a hybrid vertical layering system, employing isopycnal layers in
 212 the stratified open ocean, bottom-following σ -coordinates in coastal areas,
 213 and fixed pressure-coordinates in the mixed layer (Bleck, 2002). Interface
 214 depths change at each time step to reflect thermohaline variability, and lay-
 215 ers are more closely spaced in the upper ocean. Outputs are interpolated
 216 to a nominal latitude-longitude-depth grid and archived in NetCDF format.
 217 The model is run in near real time at the NAVOCEANO Major Shared Re-
 218 source Center to produce seven-day forecasts and four-day hindcasts. Here,
 219 analysis is performed on archived hindcast data spanning 15 May 2009 to 23
 220 October 2011. This range was chosen to encompass available model output
 221 during a unified period of high data return from mooring instruments. Hourly
 222 hindcast data are publicly available on the HYCOM consoritum data server
 223 (<http://hycom.org/dataserver>). For a detailed description of the model and
 224 its outputs, the reader is referred to [http://hycom.org/data/goml0pt04/expt-](http://hycom.org/data/goml0pt04/expt-31pt0)
 225 31pt0. For a detailed description of HYCOM, the reader is referred to Bleck
 226 (2002), Chassignet et al. (2003), and Chassignet et al. (2006).

227 HYCOM31.0 uses the 3D-VAR Navy Coupled Ocean Data Assimila-
 228 tion (NCODA) system (Cummings, 2005; Cummings and Smedstad, 2013).
 229 NCODA assimilates all available observations. These include surface infor-
 230 mation from satellites (SST and SSH), plus *in situ* temperature and salinity
 231 profiles from XBTs (expendable bathythermographs), CTDs (conductivity-
 232 temperature-depth), gliders, and Argo floats (Chassignet et al., 2007, 2009;
 233 Cummings and Smedstad, 2013; Metzger et al., 2014). Satellite altimetry
 234 for NCODA comes from the NAVOCEANO Altimeter Data Fusion Center,
 235 which combines SSH from Jason-1, OSTM/Jason-2, Geosat, and Envisat.

236 Vertical projection of the surface observations is achieved via generation of
237 synthetic profiles using the Modular Ocean Data Analysis System (MODAS;
238 Fox et al. 2002).

239 Midnight snapshots were used for this study: 00z model hindcasts were
240 compiled into time series and low-passed with a 72-hour 4th order low-pass
241 Butterworth filter. This filtering paralleled the treatment of the DynLoop
242 observations. Modeled temperature and velocity at the grid points closest to
243 mooring sites were used in site-to-site comparisons of temperature and ve-
244 locity between tall moorings and HYCOM31.0. Differences between mooring
245 locations and nearest model grid point were less than 2.2 km. Tall moor-
246 ings experienced “blow-down” or “draw-down” during time periods of strong
247 currents. This drew instrumentation below its nominal depth. Therefore,
248 measurement depth $p(t)$ varied with time. For point comparisons, model
249 temperature and velocity were also vertically interpolated to $p(t)$ for each
250 moored sensor. If a companion pressure measurement did not exist for a
251 current meter or temperature sensor, $p(t)$ was constructed by linear interpo-
252 lation of pressure records above and below the sensor.

253 Following Dukhovskoy et al. (2015) and Leben (2005), the position of the
254 modeled LC is also tracked using the 17-cm contour in the demeaned SSH
255 fields. Note that in this work, the SSH contours are used qualitatively to
256 place statistical quantities, such as eddy kinetic energy and SSH variance,
257 into the context of the LC position.

258 *2.3. Methodology*

259 Taylor diagrams display the simultaneous comparison of multiple time
260 series (Taylor, 2001). In the Taylor diagram representation, comparisons are

made to a “reference” time series. Here, the reference time series are the observations: to compare 900 m temperature at mooring a1, for example, the observational time series at this location and depth is used as a reference for comparison with the modeled equivalent. A comparison at one depth and location yields a single point on the Taylor diagram indicating correlation coefficient and root-mean-squared difference (RMSD) between the modeled and observed time series, as well as the ratio of their standard deviations ($\sigma_{hyc}/\sigma_{obs}$). Hence, the ideal comparison has a correlation of 1.0, zero RMSD, and $\sigma_{hyc}/\sigma_{obs} = 1.0$. Note that RMSD is normalized by the standard deviation of the reference series, and that this normalized value will be referred to herein simply as RMSD. Because the RMSD is normalized by standard deviation, its inverse is a proxy for signal-to-noise ratio. The ratio $\sigma_{hyc}/\sigma_{obs}$ evaluates the relative magnitude of variance of a modeled time series compared to the corresponding observation (Taylor, 2001).

Array-mean model-to-observation coherence was calculated by averaging cross- and auto-spectral density functions over all PIES sites. If P_{xy} is the cross-spectral density between HYCOM31.0 and PIES, and P_{xx} and P_{yy} are the power spectral densities of HYCOM31.0 and PIES measurements, respectively, at a single site, then the array-mean coherence is given by

$$C_{avg} = \frac{|\langle P_{xy} \rangle|^2}{\langle P_{xx} \rangle \langle P_{yy} \rangle}, \quad (1)$$

where $\langle \rangle$ indicates the average over all sites. For this study, P_{xx} , P_{yy} , and P_{xy} were calculated using Welch’s method with a 128-day Hanning window and 50% overlap (see Bendat and Piersol 2000). Error is estimated by the 95% confidence limit following Harris (1978) and Thompson (1979).

284 Complex Empirical Orthogonal Functions (CEOF) of mapped PIES and
 285 HYCOM31.0 SSH fields were generated for each eddy event and for four
 286 frequency bands to quantify meander propagation. Here we followed the
 287 methodology of Barnett (1983), where the cross-covariance matrix for the
 288 EOF is derived from the scalar band-passed SSH fields and their Hilbert
 289 transform. The CEOF method yields a spatial amplitude and phase, as well
 290 as a temporal amplitude and phase. This differs from a complex vector EOF
 291 where, for example, the cross-covariance matrix for the EOF comes from the
 292 complex input time series $U = u + iv$ where u and v are zonal and meridional
 293 velocities. A review of EOF methods can be found in Hannachi et al. (2007).
 294 Following comparable analysis in Donohue et al. (2015), for each CEOF the
 295 spatial phase gradient, $\delta\phi/\delta s$ where ϕ is phase and s is distance, is calculated
 296 for regions where the corresponding normalized CEOF spatial amplitude is
 297 greater than 0.5. Note that $\delta\phi/\delta s$ is the magnitude of the wavenumber.
 298 Propagation phase speed is then determined from

$$c_p = \frac{\omega}{\delta\phi/\delta s} \quad (2)$$

299 where ω is the central frequency of a given frequency band.

300 Model mapped stream function fields were generated by optimally inter-
 301 polating HYCOM31.0 velocity fields using a process adapted from Bretherton
 302 et al. (1976), detailed in Watts et al. (1989, 2001), and applied to the Dyn-
 303 Loop observations (Hamilton et al., 2014; Donohue et al., 2015, 2016). A
 304 correlation length scale of 50 km was used.

Figure 3: Time series of observed (black) and modeled (gray) (a) temperature, (b) zonal, and (c) meridional velocity. Nominal depths are noted along the right side of each panel. Correlation coefficients between observed and modeled velocity time series are given in the lower left corners. Temperature correlation coefficients are 0.88 for the 500 and 900 m depths and 0.82 for the 900 m record. Note that y -axis limits vary.

3. Time-Series Point Comparisons

Figure 3 shows time series of temperature, zonal, and meridional velocity at mooring a1. Visually, modeled upper ocean temperatures and 150 m velocity time series closely track their corresponding observed time series (Figure 3). Temperatures below the thermocline (~ 900 m depth) are quite uniform,

310 therefore temperature comparisons were restricted to the upper 900 m of the
 311 water column. Correlation coefficients cited in Figure 3 provide qualitative
 312 assessment of how well the two time series co-vary in time. Statistical sig-
 313 nificance of correlation coefficients are discussed below when presented in
 314 Taylor diagrams. Multiple time scales are evident in the temperature and
 315 150 m velocity time series. There was a low-frequency (> 300 day) signal
 316 associated with the intrusion and retreat of the LC. In the mesoscale (3–100
 317 day) band, relatively high-frequency oscillations, ~ 3 –20 day, tended to occur
 318 in this record as the LC entered the array, for example in May/June 2010,
 319 followed by lower-frequency variability between 40–100 day. At depth, model
 320 and velocity time series do not consistently co-vary with one another (Figure
 321 3b,c: 900 m and 2900 m). Both model and observations showed increased
 322 deep variability during LCE separation events, however, this enhancement
 323 was more dramatic in the observations; for example, the strong pulses in Oc-
 324 tober 2009 during Eddy Ekman’s separation and August 2011 during Eddy
 325 Hadal’s separation.

326 Figure 4a summarizes the point-to-point temperature statistics. Standard
 327 deviation ratios above the thermocline (~ 600 –900 m depth) were clustered
 328 near 1.0, indicative of comparable variance between model and observations,
 329 and ranged between 0.62 and 1.27. There is a tendency for model records
 330 deeper than ~ 600 m depth to have reduced variance relative to observations.
 331 The majority of normalized RMSDs were below 0.5. This corresponds to
 332 signal-to-noise ratios above 1.0 for these points. Dimensional RMSD (Fig-
 333 ure 5a) decreased with depth, with values near 1.5°C , 1.25°C , 0.7°C , and
 334 0.4°C , at 75 m, 250 m, 600 m, and 900 m, respectively. Correlation coeffi-

Figure 4: Taylor diagram of observation-to-model correlation (blue labeled axis), normalized RMS difference (red labeled axis), and standard-deviation ratio (black labeled axis) for (a) temperature, (b) SSH, (c) zonal velocity, and (d) meridional velocity comparisons. Time series depths are denoted by color scaling: a key is provided below panels c,d for velocity and below panel a for temperature. The black dot in each panel indicates the reference point. Green (purple) filled circles in panel d indicate PIES sites co-located (not co-located) with OSTM/Jason-2 altimeter tracks. Red and blue lines in panels a,b indicate 95% statistical significance for the correlations above and below 900 m, respectively, and blue line in c shows 95% significance for all temperature correlations.

Figure 5: Dimensional RMSD versus depth for all (a) temperature and (b) zonal (gray circles) and meridional (black points) velocity. Note that y -axis scales differ between panels.

335 cients ranged from 0.75 to 0.98 for all moorings and depths, with an array-
 336 averaged correlation of 0.92, indicating that modeled and observed temper-
 337 atures had a similar pattern of variability. These correlations, interpreted
 338 in the context of a linear relationship, show that on average 85% of the
 339 common variance is explained by a linear fit. All correlations were statis-
 340 tically different from zero at the 95% confidence level. Degrees of freedom
 341 (DOF) were determined from autocorrelations of the measurements following
 342 the methodology discussed in Bendat and Piersol (2000). Average DOF for
 343 the temperature time series was near 15. All temperature correlations were
 344 greater than 0.482, the criteria for 95% statistical significance.

345 Similar to the upper-ocean temperature comparisons, model and PIES
 346 SSH agree well with one another in that standard deviation ratios were near
 347 one, the majority of the normalized RMSD were less than 0.5, and corre-
 348 lation coefficients were above 0.84 (Figure 4b). Standard deviation ratios
 349 ranged from 0.82 to 1.26 with a mean of 1.03. All comparisons resulted in

350 normalized RMSD lower than 0.58 with a minimum of 0.26, corresponding
 351 to a dimensional RMSD range of 7–14 cm. Correlation coefficients ranged
 352 between 0.84 and 0.97, with mean value of 0.93. DOF for SSH were near
 353 15. Hence, all SSH comparisons were statistically significant at the 95% level
 354 ($r > 0.482$). Modeled SSH explains nearly 87% of the observed signal. No
 355 distinction in statistics were found for sites on or off the OSTM/Jason-2 al-
 356 timeter ground tracks. Our interpretation of this result is that, in general,
 357 high correlation coefficients occurred at points with high variance, and most
 358 of the variance derived from low-frequency variability associated with the LC
 359 intrusion and retreat cycles.

Figure 6: Site-to-site (thin lines) and array-mean (thick line) mean-squared SSH coher-
 ence between HYCOM31.0 and PIES. 95% confidence limits for individual sites (horizontal
 dashed line) and array-mean (horizontal dash-dot line) give estimates of significance. Co-
 herence drops around $1/20 \text{ days}^{-1}$ (vertical dashed line), near the Nyquist frequency of the
 Jason-1 and OSTM/Jason-2 altimetry satellites. PIES co-located (not co-located) with
 OSTM/Jason-2 altimeter ground tracks are denoted by black (gray) thin lines.

360 Comparisons between observed and modeled velocity showed mixed re-

361 sults with a marked distinction between upper and deep levels in both RMSD
 362 and correlations (Figure 4c,d). Model-to-observation standard deviation ra-
 363 tios, $\sigma_{hyc}/\sigma_{ref}$, were below 1.0 for 79% of all velocity comparisons (81% for
 364 zonal, 76% for meridional) indicating lower velocity variance in the model
 365 than observations. This was especially so for depths greater than 900 m:
 366 88% of comparisons yielded ratios below 1.0. On average, modeled variance
 367 was 77% that of observations (65% below 900 m). Normalized RMSD were
 368 between 0.5 and 1.0 for depths less than 450 m. Signal to noise ratio decreased
 369 with depth, as evidenced by the increase in RMSD to values greater than 1.0
 370 for the majority of velocity comparisons below 600 m. Dimensional RMSD
 371 are shown in Figure 5b: RMSD was greatest in the upper water column with
 372 a maximum of 0.33 m s^{-1} at 80 m, and decreased with depth to below 0.14
 373 m s^{-1} deeper than 900 m and to $\sim 0.1 \text{ m s}^{-1}$ around 3000 m depth. Above
 374 900 m, mean correlations ranged between 0.62 and 0.74. Average DOF for
 375 velocity time series varied with depth, reflecting the larger contribution of
 376 low-frequency variability in the upper-ocean spectra. In the upper-ocean,
 377 for depths above 900 m, DOF were near 25, hence, correlations greater than
 378 0.381 were significant at the 95% level. We note that although correlations
 379 in the upper 900 m were statistically significant, the variance explained is
 380 low, ranging from 38–55%. Below 900 m depth, mean correlations were low:
 381 0.30 and 0.12 for zonal and meridional velocity, respectively. At 900 m and
 382 below, DOF were near 60, with 0.250 as the criteria for 95% statistical signif-
 383 icance. Again, while a handful of sites had correlations statistically different
 384 than zero, the explained variance is low. Curiously, there were differences
 385 between zonal and meridional comparison statistics. For depths greater than

386 500 m, zonal velocity correlation coefficients, RMSD, standard deviation ra-
387 tios indicated better overall agreement with observations than for meridional
388 velocities. The reasons for this are not well understood at this time.

389 To investigate the agreement between model and observations as a func-
390 tion of frequency, mean-squared coherence between HYCOM31.0 and PIES
391 SSH was calculated. At all PIES sites, coherence decreased as frequency in-
392 creased (Figure 6). Many of the individual site-to-site coherences fell below
393 the 95% confidence limit near a frequency of $1/20 \text{ days}^{-1}$. Array-mean coher-
394 ence also fell sharply at this frequency, which corresponds to the Nyquist fre-
395 quency of the Jason-1 and OSTM/Jason-2 altimetry missions that provided
396 data assimilated by HYCOM31.0. Note that the variability for frequencies
397 higher than $1/20 \text{ days}^{-1}$ represented a small fraction, $< 2\%$, of the total
398 variability, and only $\sim 8\%$ of the variance for mesoscale frequencies (100–3
399 day). While there was a sharp decrease in coherence below $1/20 \text{ d}^{-1}$, statis-
400 tically significant coherence did exist at some sites for the high frequencies.
401 We explore the spatial distribution of SSH variance further in section 4.1.

402 Point-to-point comparisons are demanding: a model may correctly sim-
403 ulate circulation features, but a spatial or temporal offset from observations
404 could spoil the point-to-point comparison. Moreover, point-to-point compar-
405 isons offer limited insight into how well a model simulates a specific oceanic
406 process. Taking this into consideration, the remainder of this paper focuses
407 on broad-scale and feature-based comparison.

Figure 7: Time-averaged (a) observed and (b) modeled eddy kinetic energy (EKE; shading) at 200 m depth, with time-mean velocity vectors superimposed. PIES (circles), tall-mooring (diamonds), and near-bottom mooring (triangles) locations are plotted along with bathymetry contoured at 1000, 2000, 3000 m depth (thin contours) and mean Loop Current position (thick black curve). (c) Time series of array-mean observed (black) and HYCOM31.0 (solid gray) 200 m EKE averaged over the same region, and LC area (dashed) from the CCAR SSH product.

4. Broad-Scale Spatial Patterns

In the upper ocean, observed and modeled EKE exhibit similar spatial structure and strength. Observed and modeled EKE at 200 m depth is shown in Figure 7. Both fields exhibit bands of high EKE along the mean path of the LC. Amplitudes of array-averaged 200 m EKE from mapped PIES and HYCOM31.0 were comparable, with time-mean values of $\sim 580 \text{ cm}^2 \text{s}^{-2}$ and $\sim 600 \text{ cm}^2 \text{s}^{-2}$, respectively. Time series of observed and modeled array-averaged EKE matched well (Figure 7c): peaks occurred together during time periods when the LC is positioned within the array; the correlation

Figure 8: Same as Figure 7, but for 2500 m depth.

between the series is 0.72.

A time series of modeled array-averaged deep (2500 m) EKE shows peaks consistent with observations prior to and during eddy separations, but with roughly half (53%) the observed amplitude (Figure 8c). Correlation between the two array-averaged time series was 0.68. Spatial patterns of EKE agree in the sense that both model and observations showed enhanced deep eddy variability in the eastern portion of the array, but these maps showed again that modeled deep EKE was approximately half that of observations (Figure 8a,b). Note that the mean fields both showed deep mean anticyclonic circulation in the northwestern array, and a deep cyclone in the northeast corner. The model, however, showed features that were not present in observations: a deep northern flow just offshore of the West Florida Shelf, and a deep anticyclone in the southern array.

Figure 9: Standard deviation of PIES (top row) and HYCOM31.0 (middle row) band-passed SSH, with increasing band frequency from left to right. Black dots show PIES locations. Similar magnitudes and patterns of variance are seen between datasets. Bottom panels map the correlation coefficient between the two series. Satellite altimeter tracks are also plotted on each map: OSTM/Jason-2 (green), Jason-1 Tandem Mission (red), and ERS (blue). Black contours in the bottom row indicate statistical significance at the 95% confidence level for each band.

4.1. Sea Surface Height Variance in Frequency Space

SSH variance was dominated by the intrusion and retreat of the LC associated with the LCE cycle (Figure 9a,f). Periodicities longer than 100 days accounted for $\sim 80\%$ of the SSH variance. Liu and Weisberg (2012) determined the peak-to-peak amplitude of the seasonal steric signal to be near 12 cm, which in terms of standard deviation is 4.2 cm. Therefore, a small portion, between 2–5% ($4.2^2/30^2 - 4.2^2/20^2$) of this variance is due to the seasonal steric signal. Shorter-period mesoscale (100–3 day) meanders play an important role in LCE dynamics. To investigate spatial patterns as a

439 function of frequency, SSH fields were band-passed into four frequency bands.
 440 Cut-off frequencies for the bands followed Donohue et al. (2015) and were
 441 based upon peaks in array-measured SSH spectra near $1/60$, $1/30$, and $1/15$
 442 d^{-1} . The four bands include two low-frequency bands corresponding to pe-
 443 riods of 100–40 and 40–20 days, and two high-frequency bands with periods
 444 of 20–10 and 10–3 days. The mesoscale band, 100–3 days, represented 12%
 445 and 13% of modeled and observed total SSH variance, respectively, within
 446 the mapping array. In the mesoscale band and within the mapping array,
 447 modeled variance was distributed as follows: 64%, 22%, 9%, and 5% of vari-
 448 ance in the 100–40, 40–20, 20–10, and 10–3 day bands, respectively. This is
 449 compared to 70%, 21%, 6%, and 2% for observations. Note that Donohue
 450 et al. (2015) assessed bottom-referenced baroclinic SSH (SSH_{bc}), rather than
 451 total SSH, hence percent variance cited here differ slightly for observations.

452 Maps of standard deviation of band-passed HYCOM31.0 SSH fields (Fig-
 453 ure 9g–j) revealed similar spatial distributions of variance to those found
 454 by Donohue et al. (2015) (Figure 9b–e). In the two low-frequency bands,
 455 variance was highest along the eastern and southeastern sides of the array,
 456 while in contrast, the two high-frequency bands had elevated variance along
 457 the north-northwest portion of the array. In the mesoscale band, meanders
 458 along the LC path, including adjacent frontal eddies, were responsible for
 459 the variance distribution. The CEOF analysis of Donohue et al. (2015) was
 460 repeated using modeled and observed SSH fields to document wavelengths
 461 and phase speeds associated with these spatial patterns (see Section 5.1).

462 The bottom panels of Figure 9 show correlation between observed and
 463 HYCOM31.0 SSH. As expected from Figure 6, correlations decreased as fre-

quency increased, with marginally significant correlations for the highest frequency band. In the full band and 40–20 day band, correlations at 100% of the points were significant at the 95% confidence level; 93%, 67%, and 22% of points in the 100–40, 20–10, and 10–3 day bands, respectively, had significant correlation. For reference, correlations greater than 0.482, 0.468, 0.330, 0.236, and 0.140 were significant for the full band, 100–40, 40–20, 20–10, and 10–3 day bands, respectively. In each frequency band, regions of high variance and high correlations were co-located. No obvious relationship between satellite tracks and correlation was found.

5. Phenomenological Comparisons

5.1. *LC Meander Characteristics*

To investigate the propagation characteristics of LC meanders, CEOFS were determined from observed and modeled SSH for four time periods when the LC was positioned within the DynLoop array and for the four frequency bands used to partition the mesoscale variance in Figure 9. We term the time periods by the LCE event: Ekman May 1 – September 1, 2009; Franklin February 1 – September 1, 2010; Hadal May 1 – August 1, 2011; and Icarus September 1 – October 23, 2011. For these CEOFs, we considered only the first CEOF mode. Variance explained by the first mode exceeded twice the variance explained by the second mode, with one exception for the observations: Ekman 20–10 day band; and four exceptions for the model: Ekman, Franklin, Hadal 20–10 day band, and Icarus 10–3 day band (Table 1). Spatial amplitude and phase are shown in Figures 10 through 13. Note that this was a slightly different analysis than Donohue et al. (2015), where bottom-

Table 1: Percentage of total CEOF variance explained by the first and second mode for each eddy event and frequency band from CEOFs of PIES and HYCOM sea surface height fields.

	Band (days)	PIES (Mode Variance) Mode-1 / Mode-2	HYCOM31.0 (Mode Variance) Mode-1 / Mode-2
Ekman <i>04 May 2009 – 01 Sep. 2009</i>	100–40	89.2% / 9.60%	87.6% / 9.6%
	40–20	62.6% / 30.0%	63.5% / 30.0%
	20–10	48.8% / 28.9%	56.3% / 28.9%
	10–3	69.0% / 12.8%	41.5% / 12.8%
Franklin <i>01 Feb. 2010 – 01 Sep. 2010</i>	100–40	79.6% / 14.6%	70.0% / 14.6%
	40–20	57.2% / 21.4%	57.1% / 21.4%
	20–10	53.5% / 21.1%	32.9% / 21.1%
	10–3	54.9% / 14.5%	39.6% / 14.5%
Hadal <i>01 Mar. 2011 – 01 Aug. 2011</i>	100–40	85.3% / 12.5%	83.6% / 12.5%
	40–20	65.9% / 21.3%	72.7% / 21.3%
	20–10	50.7% / 22.0%	32.9% / 21.1%
	10–3	35.7% / 16.9%	34.5% / 16.9%
Icarus <i>01 Sep. 2011 – 23 Oct. 2011</i>	20–10	77.2% / 13.0%	52.1% / 13.0%
	10–3	56.3% / 23.1%	36.2% / 23.1%

reference baroclinic SSH was used rather than total SSH. Nevertheless, the overall patterns and phase speeds were similar: phase speeds from Donohue et al. (2015) ranged from 8 to 50 km day⁻¹ and those presented here using total SSH spanned a range of 8 to 51 km day⁻¹.

Modeled and observed CEOF spatial patterns in the low frequency bands (100–40 and 40–20 day) shared the following characteristics. In the 100–40 day band (Figures 10–12; panels a–d), spatial amplitudes were high along the eastern side of the LC; propagation was clockwise. In the 40–20 day band (Figures 10–12; panels e–h), modeled and observed CEOF spatial peaks appear in similar regions of the array, with clockwise propagation along the LC.

498 For these low frequency bands, the DOF are low. For this reason, we do not
 499 show Icarus 100–40 or 40–20 day band CEOFs. Note that, while the degrees
 500 of freedom are limited within each time period, the wavenumber/phase speed
 501 estimates from the three LCEs each provide independent estimates.

Figure 10: First-mode SSH CEOFs for the Ekman time period by frequency band. Bands are labeled at the top of each four-panel band-group. Normalized CEOF amplitude is presented in the left panels of each group, and phase (in degrees) in the right panels. PIES and model results are shown in the upper and lower panels of each group, respectively. For all panels: Bathymetry (gray contours; 1000 m intervals), PIES locations (black dots), and mean Loop Current position (thick black line) are included. Percentage of total variance explained by the first mode is indicated in the upper-right of each amplitude plot.

For the high-frequency bands (20–10 and 10–3 day; Figures 10–12 i–p & 13 a–h), the model and observations differed from one another. This discrepancy was most notable for Eddy Ekman (Figure 10), where high spatial amplitudes in observations were confined to the northwestern portion of the array along the LC mean path, while the modeled peak was displaced slightly inward of the LC path. However, both model and observations show that these high-frequency meanders were strongest along the northeast portion of the array, except for Eddy Hadal, where the LC was located noticeably more to the

Figure 11: Same as Figure 10, but for Eddy Franklin.

Figure 12: Same as Figure 10, but for Eddy Hadal.

510 west than during other eddy events and high-frequency meanders were found
 511 along the eastern LC path. Propagation in the high-frequency bands was
 512 clockwise for all eddy events, yet the phase gradient differed between model
 513 and observations. Overall, the model showed little change in spatial phase,
 514 indicating fast propagation. This was most apparent for eddies Ekman and
 515 Franklin.

516 To quantify propagation patterns seen in COEFs, phase speed and wavenum-
 517 ber were calculated from CEOF phase fields (see Section 2.3) for each com-

Figure 13: Same as Figure 10, but for Eddy Icarus in the 20-10 and 10-3 day bands only.

Figure 14: Phase speed vs. wavenumber estimates from HYCOM31.0 (gray) and PIES (black) SSH CEOFs. Error bars are standard error. Groupings from bottom to top correspond to 100-40, 40-20, 20-10, and 10-3 day frequency bands.

518 bination of eddy and frequency band (Figure 14 and Table 2). As band fre-
519 quency increased, phase speeds increased and wavelengths decreased. Mean
520 phase speeds are within 8% and 2% of each other for the 100–40 and 40–
521 20 day band, respectively, indicating good agreement. On the other hand,
522 HYCOM31.0 CEOF phase speeds for the two high-frequency bands were
523 unrealistically large (see Table 2), and therefore not included in Figure 14.

524 In order to investigate whether data assimilation played a role in the
525 discrepancies observed between HYCOM31.0 and DynLoop results at high
526 frequencies (20–3 day band), a non data-assimilative (free running) HYCOM

Table 2: Loop Current meander phase speed (c_p), wavenumber (k), and wavelength (λ) for each combination of eddy (first column) and band (second column) derived from SSH CEOF phase fields from PIES and HYCOM31.0. Italicized values were considered unreasonable and not included in Figure 14.

	Band	PIES			HYCOM31.0		
		c_p (m s ⁻¹)	k (10 ⁻² km ⁻¹)	λ (km)	c_p (m s ⁻¹)	k (10 ⁻² km ⁻¹)	λ (km)
Ekman <i>4 May – 1 Sep.</i> <i>2009</i>	100–40	0.11	1.26	498.9	0.12	1.14	551.6
	40–20	0.19	1.81	347.3	0.17	1.82	345.5
	20–10	0.22	2.59	243.0	<i>0.84</i>	<i>0.80</i>	<i>782.6</i>
	10–3	0.58	2.82	223.2	<i>2.51</i>	<i>0.78</i>	<i>803.1</i>
Franklin <i>1 Feb. – 1 Sep.</i> <i>2010</i>	100–40	0.09	1.52	412.2	0.09	1.50	418.2
	40–20	0.24	1.62	387.3	0.23	1.44	435.4
	20–10	0.22	2.54	247.5	<i>0.93</i>	<i>1.06</i>	<i>592.1</i>
	10–3	0.59	2.75	228.8	<i>5.15</i>	<i>0.48</i>	<i>1301.1</i>
Hadal <i>1 Mar. – 1 Aug.</i> <i>2011</i>	100–40	0.11	1.39	453.2	0.12	1.24	508.5
	40–20	0.16	1.81	346.8	0.20	1.43	439.4
	20–10	0.25	2.25	279.2	<i>0.51</i>	<i>1.57</i>	<i>401.2</i>
	10–3	0.52	3.26	192.8	<i>1.45</i>	<i>1.27</i>	<i>496.2</i>
Icarus <i>1 Sep. – 23 Oct.</i> <i>2011</i>	20–10	0.29	2.08	302.8	<i>0.43</i>	<i>1.43</i>	<i>439.9</i>
	10–3	0.59	2.78	225.8	<i>1.31</i>	<i>1.61</i>	<i>391.3</i>
	—	—	—	—	—	—	—

Table 3: Same as Table 1, but for three eddy time periods from free-running expt.02.2.

	Band	Mode Variance
		Mode-1 / Mode-2
1 Jan. – 15 Mar. <i>1957</i>	100–40 day	77.5% / 20.3%
	40–20 day	63.2% / 26.7%
	20–10 day	52.2% / 28.7%
	10–3 day	37.6% / 21.4%
1 May. – 1 Aug. <i>1957</i>	100–40 day	71.8% / 25.2%
	40–20 day	70.2% / 22.1%
	20–10 day	41.1% / 27.9%
	10–3 day	38.4% / 20.9%
1 Apr. – 15 Jul. <i>1958</i>	100–40 day	84.6% / 12.6%
	40–20 day	64.0% / 24.6%
	20–10 day	56.8% / 18.8%
	10–3 day	34.3% / 18.7%

527 configuration was examined. The free running model, HYCOM GOMl0.04
 528 experiment 02.2, utilized the same horizontal resolution and number of hy-
 529 brid vertical layers as HYCOM31.0 (see Dukhovskoy et al. (2015) for a de-
 530 tailed description). Three LCE eddy events were identified that resembled
 531 the DynLoop observational period. SSH CEOFs were calculated for each of
 532 the three eddies in the four frequency bands. These were used to compute
 533 phase speed and wavenumbers. The first mode CEOFs are shown in Fig-
 534 ures 15 through 17, and Table 3 provides the variance explained by the first
 535 two CEOF modes in each band. Because of the large amplitude (high vari-
 536 ance) signals occurring on the West Florida Shelf, the highest frequency (10–3
 537 day) band CEOFs excluded model data east of 84°W. Figures 15 through
 538 17 share similar characteristics to what was observed in DynLoop during the
 539 three eddy events. Consistent with observations, there was a tendency for
 540 low-frequency (100–20 day) and high-frequency (20–10 day) meanders to be

Figure 15: CEOFs of band-passed SSH from free-running HYCOM expt.02.2 during model dates 01 Jan. to 15 Mar. 1957. Frequency bands (rows) increase in frequency from top to bottom. First-mode CEOF amplitude (left column) and phase in degrees (right column) are overlaid with mean Loop Current position (thick black line) from model SSH and bathymetry (gray contours; 1000 m interval). Percentage of total variance explained by the first mode is printed in the upper-right of each phase plot. Propagation is in the direction of increasing phase (light to dark; right panels).

541 strongest along the eastern and western edges of the LC, respectively. Unlike
 542 the CEOFs for HYCOM31.0, spatial phase fields from the free-running model
 543 show both the high- and low-frequency signal propagating along the LC at
 544 speeds comparable to observations; recall that HYCOM31.0 high-frequency

Figure 16: Same as Figure 15, but for free-running model dates 01 May – 01 Aug. 1957.

545 phase speeds were unrealistically large. This suggests an improvement over
 546 HYCOM31.0 at these high-frequencies. Phase speeds and wavenumbers de-
 547 rived from expt_02.2 matched closely with those from PIES observations for
 548 all four frequency bands (Figure 18): differences from observations in both

Figure 17: Same as Figure 15, but for free-running model dates 01 Apr. – 15 Jul. 1958.

Figure 18: Phase speed vs. wavenumber comparison derived from CEOFs of assimilated (gray circles) and free-running HYCOM (gray diamonds), and from PIES (black) sea surface height for each frequency band. Error bars are standard error. Groupings from bottom to top correspond to 100–40, 40–20, 20–10, and 10–3 day frequency bands.

549 phase speed and wavenumber were less than 9% and 4% in the 100–20 day
 550 band and less than 4% and 1% in the 20–3 day band. These results imply that
 551 the high-frequency altimeter sampling and assimilation could have negative
 552 impacts on the accuracy of phase speeds in the data-assimilative HYCOM
 553 — this needs to be further investigated by the HYCOM development team.

554 5.2. Stream Function Case Study: Upper-Deep Layer Coupling

555 Our stream function case studies focus on the 100–40 day band because
 556 observations showed coherent upper-deep structure in stream function with
 557 a 90° along-stream phase offset consistent with baroclinic instability (Dono-
 558 hue et al., 2016). Figures 19–21 show three case studies of upper (200 m
 559 relative to 2500 m) and deep (2500 m) 100–40 day band-passed stream func-
 560 tion for eddies Ekman, Franklin, and Hadal, respectively. All three cases

Table 4: Same as Table 2, but for three eddy periods from free-running expt_02.2.

	Band	c_p	k	λ
	(days)	(m s ⁻¹)	(10 ⁻² km ⁻¹)	(km)
1 Jan. – 15 Mar. <i>1957</i>	100–40	0.09	1.48	424.4
	40–20	0.17	1.81	346.7
	20–10	0.28	2.11	298.1
	10–3	0.49	3.31	190.0
1 May. – 1 Aug. <i>1957</i>	100–40	0.11	1.36	461.9
	40–20	0.19	1.53	409.7
	20–10	0.28	2.07	304.1
	10–3	0.52	3.14	200.2
1 Apr. – 15 Jul. <i>1958</i>	100–40	0.09	1.59	395.0
	40–20	0.20	1.51	417.5
	20–10	0.25	2.28	275.4
	10–3	0.54	3.01	208.4

561 demonstrated that strong deep eddies that occur during LCE formation.
562 Additionally, each deep cyclone (anticyclone) tended to be paired, but offset
563 downstream from an upper cyclone (anticyclone) in a pattern indicative of
564 baroclinic instability (Cushman-Roisin, 1994). These patterns, seen in ob-
565 servations, were reproduced by HYCOM31.0. In each case study, examples
566 of these upper-deep pairs are identified in the following descriptions, with
567 the deep cyclone or anticyclone denoted by letters A–D in each figure.

568 During Eddy Ekman’s separation, an upper-deep cyclone pair (A) entered
569 the mapping array from the north on 22 June 2009 and propagated clockwise
570 along the eastern edge of the array to arrive in the southeast portion of the
571 array on 22 July 2009 (Figure 19a–f). At that time, a second upper-deep pair
572 (B), an anticyclonic pair, entered the array from the north. The features were
573 seen in stream function fields from both observations and HYCOM31.0, and
574 matched closely in amplitude, shape, size, and position. Anticyclone pair

Figure 19: Upper (200 m relative to 2500 m; shading) and deep (2500 m; contours) 100–40 day band-passed stream function comparison between observations and HYCOM31.0 at six-day intervals during Eddy Ekman. Green contours indicate altimeter-measured and modeled Loop Current mean position for PIES and HYCOM31.0, respectively. The mapping array is outlined in black with PIES sites indicated by small circles. Gray contours show 1000, 2000, and 3000 m bathymetry.

Figure 20: Same as Figure 19, but for Eddy Franklin.

575 B followed a similar trajectory to that of A and was found in the central
576 eastern array on 3 August 2009 (Figure 19 f–h), at which time eddy pair
577 A appeared to have dissipated in HYCOM31.0. Maps of observed stream
578 function on August 3rd showed A exiting the array to the south, but its fate
579 was unclear due to the spatial limits of the array. From these maps, it seems
580 likely that A and/or B played a role in the first detachment of Ekman: as
581 the deep cyclone associated with pair A exited the array the LC experienced
582 a necking down and eventual detachment on 9 August 2009. On 3 August
583 2009, upper-deep cyclone pair C entered at the base of the Mississippi Fan
584 near the northwest corner of the array, propagated southward, and appeared
585 to dissipate after Eddy Ekman underwent a detachment around 9 August.

586 Two offset upper-deep eddy pairs, A and B, were present on 19 May 2010,
587 the first day of the Eddy Franklin case study (Figure 20a), in addition to
588 a more southern cyclone pair seen clearly in HYCOM31.0. Eddy pairs A
589 and B propagated southward along the continental slope and appeared to
590 facilitate Franklin’s first detachment around 12 June 2010 (Figures 20 b–
591 e). Both features were well represented by HYCOM31.0. Anticyclone pair
592 A dissipated around June 6th, while cyclone pair B continued to propagate
593 southward followed by anticyclone pair C, which appeared on 18 June 2010.
594 The latter two pairs assisted in a second detachment of Franklin between the
595 6th and 12th of July (Figure 20f–j). Cyclone pair D entered the array on the
596 18 July 2010 and played a role in the final separation of LCE Franklin.

597 During Eddy Hadal, similar to the Ekman and Franklin cases, a series of
598 southward-propagating cyclone and anticyclone pairs appeared (Figure 21).
599 In the Eddy Hadal case study, the correspondence between observations and

Figure 21: Same as Figure 19, but for Eddy Hadal using one-week intervals.

600 HYCOM31.0 was not as strong. Upper and deep eddies occurred in roughly
601 the same location, but deep eddies in HYCOM31.0 appeared more elongated
602 than those of observations.

603 **6. Discussion and Conclusion**

604 A full-water-column mesoscale-resolving observational dataset that recorded
605 four LC eddy shedding events permitted an in-depth model-data comparison.
606 The $1/25^\circ$ data-assimilative GOM HYCOM 31.0 was compared to observa-
607 tions in three categories of metrics: statistical point comparisons, broad-scale
608 spatial comparisons, and process-based phenomenological comparisons. The
609 first category sought to quantify correlations, RMSD, and variance ratios.
610 Because the overall aim of this study was to evaluate the model’s ability to
611 accurately represent processes involved in the LCE formation/detachment
612 cycle, the second and third metric categories focused on assessment of the
613 model’s representation of LC meander variability, wavenumber-frequency
614 characteristics, and upper-deep coupling during LCE formation.

615 Statistical point-comparisons showed that in the upper ocean HYCOM31.0
616 and DynLoop agree well. This was especially true of the temperature com-
617 parisons: above-thermocline array-averaged correlation was 0.93, normalized
618 RMSD ranged between 0.21 and 0.76, and variance was comparable between
619 model and observations. This indicates that the NCODA vertical projec-
620 tion of synthetic temperature profiles derived from altimeter SSH works well
621 in the Gulf. SSH variance was dominated by the large array-scale nearly-
622 annual cycle of LC advance and retreat; the PIES/HYCOM31.0 SSH time
623 series comparison (summarized in Figure 4b), therefore, showed no statis-

624 tical distinction between sites on or off OSTM/Jason-2 altimeter ground
625 tracks. Distinct differences between upper and deep velocity comparisons
626 were apparent: mean velocity correlations above and below 900 m were ap-
627 proximately 0.7 and 0.2, respectively, and modeled upper- and deep-ocean
628 velocity variances were, on average, 21% and 35% less than observed vari-
629 ances above and below 900 m depth, respectively.

630 To focus on the mesoscale circulation, the spatial pattern of SSH variance
631 in four frequency bands was evaluated. In the 100–40 and 40–20 day bands,
632 modeled and observed SSH revealed meanders that grew and propagated
633 downstream along the eastern portion of the LC, with phase speeds between
634 0.09 and 0.24 m s⁻¹. Mean phase speeds from HYCOM31.0 and observations
635 agreed within 8% and 2% in the 100–40 and 40–20 day band, respectively. Al-
636 though the spatial variance pattern for the two high-frequency bands (20–10
637 and 10–3 day) looked similar, propagation speeds did not agree well: model
638 phase speeds were unrealistically large. This was consistent with the result
639 that SSH coherence between HYCOM31.0 and PIES SSH fell off rapidly for
640 frequencies higher than 1/20 d⁻¹. We speculate that, for the high-frequencies,
641 altimeter sampling influences the agreement between observations and model,
642 noting that phase speeds determined from a comparable free-running ver-
643 sion of GOM HYCOM differed from observed values by less than 9% for *all*
644 frequency bands. Liu et al. (2014) assessed the relative skill of a suite of
645 altimeter-derived surface current products and model output. They found
646 that the altimeter-derived products performed slightly better than the 1/25°
647 GOM data-assimilative HYCOM, and suggested that increased data cover-
648 age might improve HYCOM’s performance. Outstanding questions, such as

649 the one raised by the DynLoop program as to whether high-frequency mean-
650 ders propagate along the full length of the LC, are therefore currently best
651 addressed with a free-running model.

652 Observations and numerical models indicate that deep eddies play a role
653 in the separation of LCEs (Hurlburt and Thompson, 1980, 1982; Sturges
654 et al., 1993; Welsh and Inoue, 2000; Oey, 2008; Donohue et al., 2015, 2016).
655 Both HYCOM31.0 and observations showed that deep EKE increased during
656 LCE separation, although the amplitude of modeled deep EKE was about
657 half that observed. A comparison of world-wide current meter observations
658 to a free running $1/12^\circ$ global HYCOM configuration (Scott et al., 2010)
659 showed that the deep kinetic energy was also significantly reduced (by up to
660 a factor of three) when compared to observations, but that data assimilation
661 brought modeled kinetic energy close to observed levels. Scott et al. (2010)
662 did suggest that the quadratic bottom drag value, C_d , used in HYCOM may
663 play a role in reduced model TKE. Higher resolution may also be necessary
664 when modeling the GOM: recent modeling studies indicate that resolutions
665 higher than $1/32^\circ$ may be necessary to properly resolve deep EKE (Hurlburt
666 and Hogan 2000; Chassignet and Xu, personal communication).

667 Within the 100–40 day band, HYCOM31.0 reproduced patterns indica-
668 tive of baroclinic instability, that is, a vertical offset between upper and
669 deep stream function. While modeled deep eddies differed slightly spa-
670 tially and temporally from observations, the joint development of an upper
671 ocean meander along the eastern side of the LC and train of upper-deep
672 cyclone/anticyclone pairs that precede separation were contained within the
673 model solution.

674 Further analysis of the $1/25^\circ$ GOM data-assimilative HYCOM would pro-
675 vide insight into LCE formation and separation, offering a larger spatiotem-
676 poral window than observational arrays. For example, the trigger for the
677 development of the long wavelength meander is not well understood. Do LC
678 frontal eddies generate deep vorticity as they stretch and move off the Missis-
679 sippi Fan as suggested by Le Hénaff et al. (2012) or do pre-existing external
680 deep eddies generated near the West Florida Shelf interact with the LC?
681 Interestingly, the HYCOM31.0 case studies in Figures 19–21 suggest that
682 both mechanism might be operating. Model analysis would provide insight
683 into the radiation of the deep energy generated during LCE separation. At
684 the present time, the pathways of deep energy radiation, feedbacks between
685 upper and deep circulation, especially in regions of steep topography, are not
686 well understood due to limited observations.

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701 Baker-Yeboah, S., Watts, D. R., Byrne, D. A., 2009. Measurements of sea sur-
702 face height variability in the eastern South Atlantic from pressure-sensor
703 equipped inverted echo sounders: baroclinic and barotropic components.
704 J. Atmos. Oceanic Technol. 26 (12), 2593–2609.

705 Barnett, T., 1983. Interaction of the monsoon and Pacific Trade Wind system
706 at interannual time scales. Part II: The tropical band. Mon. Weather Rev.
707 112, 2380–2387.

708 Bendat, J. S., Piersol, A. G., 2000. Random Data: Analysis and Measurement
709 Procedures, third edition Edition. Wiley, New York.

710 Bleck, R., 2002. An oceanic general circulation model framed in hybrid
711 isopycnic-Cartesian coordinates. Ocean Modelling 4 (1), 55–88.

712 Bretherton, F. P., Davis, R. E., Fandry, C., 1976. A technique for objective
713 analysis and design of oceanographic experiments applied to MODE-73.
714 In: Deep Sea Research and Oceanographic Abstracts. Vol. 23. Elsevier,
715 pp. 559–582.

716 Bunge, L., Ochoa, J., Badan, A., Candela, J., Sheinbaum, J., 2002. Deep
717 flows in the Yucatán Channel and their relation to changes in the Loop
718 Current extension. J. Geophys. Res. 107(C12):3233.

719 Chang, Y.-L., Oey, L.-Y., 2011. Loop Current cycle: Coupled response of the
720 Loop Current with deep flows. J. Phys. Oceanogr. 41 (3), 458–471.

721 Chang, Y.-L., Oey, L.-Y., 2012. Why does the Loop Current tend to shed
722 more eddies in summer and winter? *Geophys. Res. Lett.* 39, L05605.

723 Chang, Y.-L., Oey, L.-Y., 2013. Loop Current growth and eddy shedding us-
724 ing models and observations: Numerical process experiments and satellite
725 altimetry data. *J. Phys. Oceanogr.* 43, 669–689.

726 Chassignet, E., Hurlburt, H., Metzger, E. J., Smedstad, O., Cummings, J.,
727 Halliwell, G., Bleck, R., Baraille, R., Wallcraft, A., Lozano, C., Tolman, H.,
728 Srinivasan, A., Hankin, S., Cornillon, P., Weisberg, R., Barth, A., He, R.,
729 Werner, F., Wilkin, J., 2009. US GODAE: Global Ocean Prediction with
730 the HYbrid Coordinate Ocean Model (HYCOM). *Oceanography* 22 (2),
731 64–75.

732 Chassignet, E. P., Hurlburt, H. E., Smedstad, O. M., Barron, C. N., Ko,
733 D. S., Rhodes, R. C., Shriver, J. F., Wallcraft, A. J., Arnone, R. A., 2005.
734 Assessment of data assimilative ocean models in the Gulf of Mexico using
735 ocean color. In: Sturges, W., Lugo-Fernández, A. (Eds.), *Circulation in
736 the Gulf of Mexico: Observations and Models*. Vol. 161 of *Geophysical
737 Monograph Series*. American Geophysical Union, Washington, D.C., pp.
738 87–100.

739 Chassignet, E. P., Hurlburt, H. E., Smedstad, O. M., Halliwell, G. R., Hogan,
740 P. J., Wallcraft, A. J., Baraille, R., Bleck, R., 2007. The HYCOM (HYbrid
741 Coordinate Ocean Model) data assimilative system. *J. Marine Syst.* 65 (1–
742 4), 60–83.

743 Chassignet, E. P., Hurlburt, H. E., Smedstad, O. M., Halliwell, G. R., Wal-

744 fraft, A. J., Metzger, E. J., Blanton, B. O., Lozano, C., Rao, D. B.,
745 Hogan, P. J., Srinivasan, A., 2006. Generalized vertical coordinates for
746 eddy-resolving global and coastal ocean forecasts. *Oceanography* 19, 20–
747 31.

748 Chassignet, E. P., Smith, L. T., Halliwell, G. R., Bleck, R., 2003. North At-
749 lantic simulation with the HYbrid Coordinate Ocean Model (HYCOM):
750 Impact of the vertical coordinate choice, reference density, and thermo-
751 baricity. *J. of Phys. Oceanogr.* 33, 2504–2526.

752 Chérubin, L. M., Sturges, W., Chassignet, E. P., 2005. Deep flow variability
753 in the vicinity of the Yucatan Straits from a high-resolution numerical
754 simulation. *J. Geophys. Res.* 110 (C4), C04009.

755 Cione, J. J., Uhlhorn, E. W., Aug. 2003. Sea surface temperature variability
756 in hurricanes: Implications with respect to intensity change. *Mon. Weather*
757 *Rev.* 131 (8), 1783–1796.

758 Cochrane, J., 1972. Separation of an anticyclone and subsequent development
759 in the Loop Current (1969). In: Capuro, L., Reids, J. (Eds.), *Contributions*
760 *on the Physical Oceanography of the Gulf of Mexico*. No. 2 in Texas A&M
761 *University Oceanographic Studies*. Gulf Publishing Co., Houston, TX, pp.
762 pp. 91–106.

763 Cummings, J. A., 2005. Operational multivariate ocean data assimilation. *Q.*
764 *J. Roy. Meteor. Soc.* 131 (613), 3583–3604.

765 Cummings, J. A., Smedstad, O. M., 2013. Variational data assimilation for
766 the global ocean. In: Park, S. K., Xu, L. (Eds.), *Data Assimilation for At-*

767 mospheric, Oceanic and Hydrologic Applications (Vol. II). Springer Berlin
768 Heidelberg, pp. 303–343.

769 Cushman-Roisin, B., 1994. Introduction to Geophysical Fluid Dynamics.
770 Prentice-Hall, New Jersey.

771 Donohue, K. A., Watts, D. R., Hamilton, P., Leben, R., Kennelly, M., 2016.
772 Loop Current Eddy separation and baroclinic instability. *Dyn. Atmos.*
773 *Oceans* (accepted).

774 Donohue, K. A., Watts, D. R., Hamilton, P., Leben, R., Kennelly, M., Lugo-
775 Fernández, A., 2015. Gulf of Mexico Loop Current path variability. *Dyn.*
776 *Atmos. Oceans* (accepted).

777 Donohue, K. A., Watts, D. R., Tracey, K. L., Greene, A. D., Kennelly, M.,
778 2010. Mapping circulation in the Kuroshio Extension with an array of
779 current and pressure recording inverted echo sounders. *J. Atmos. Oceanic*
780 *Technol.* 27, 507–527.

781 Dukhovskoy, D. S., Leben, R. R., Chassignet, E. P., Hall, C. A., Morey,
782 S. L., Nedbor-Gross, R., 2015. Characterization of the uncertainty of Loop
783 Current metrics using a multidecadal numerical simulation and altimeter
784 observations. *Deep Sea Res. Pt. I* 100, 140–158.

785 Fox, D., Teague, W., Barron, C., Carnes, M., Lee, C., 2002. The Modular
786 Ocean Data Analysis System (MODAS). *J. Atmos. Ocean Tech.* 19, 240–
787 252.

788 Gopalakrishnan, G., Rudnick, D. L., Owens, W. B., 2013b. State estimates

789 and forecasts of the Loop Current in the Gulf of Mexico using the MITgem
790 and its adjoint. *J. Geophys. Res. Oceans* 118, 3292–3314.

791 Hamilton, P., Donohue, K., Hall, C., Leben, R. R., Quian, H., Sheinbaum,
792 J., Watts, D. R., 2014. Observations and dynamics of the Loop Current.
793 Tech. Rep. OSC Study BOEM 2015-006, US Dept. of the Interior, Bureau
794 of Ocean Energy Management, Gulf of Mexico OSC Region, New Orleans,
795 LA.
796 URL http://www.po.gso.uri.edu/dynamics/pub_index.html

797 Hamilton, P., Lugo-Fernández, A., Sheinbaum, J., 2016. A Loop Current ex-
798 periment: Field and remote measurements. *Dyn. Atmos. Ocean.* (in press).

799 Hannachi, A., Jolliffe, I., Stephenson, D., 2007. Empirical orthogonal func-
800 tions and related techniques in atmospheric science: A review. *Intl. J. of*
801 *Clim.* 27 (9), 1119–1152.

802 Harris, F., 1978. On the use of windows for harmonic analysis with the dis-
803 crete fourier transform. *Proceedings of the IEEE* 66 (1), 51–83.

804 Hurlburt, H. E., Hogan, P. J., 2000. Impact of $1/8^\circ$ to $1/64^\circ$ resolution on
805 Gulf Stream model-data comparisons in basin-scale subtropical Atlantic
806 Ocean models. *Dyn. Atmos. Oceans* 32, 283–329.

807 Hurlburt, H. E., Thompson, J. D., 1980. A numerical study of Loop Current
808 intrusions and eddy sheeding. *J. Phys. Oceanogr.* 10, 1611–1651.

809 Hurlburt, H. E., Thompson, J. D., 1982. The dynamics of Loop Current and
810 shed eddies in a numerical model of the Gulf of Mexico. In: Nihoul, J.

811 (Ed.), *Hydrodynamics of Semi-enclosed Seas*. Elsevier Sci., New York, pp.
812 243–298.

813 Kantha, L., Choi, J.-K., Schaudt, K. J., Cooper, C. K., 2005. A regional data-
814 assimilative model for operational use in the Gulf of Mexico. In: Sturges,
815 W., Lugo-Fernández, A. (Eds.), *Circulation in the Gulf of Mexico: Obser-
816 vations and Models*. Vol. 161 of Geophysical Monograph Series. American
817 Geophysical Union, Washington, D.C., pp. pp. 165–180.

818 Le Hénaff, M., Kourafalou, V. H., Dussurget, R., Lumpkin, R., 2014. Cy-
819 clonic activity in the eastern Gulf of Mexico: Characterization from along-
820 track altimetry and in situ drifter trajectories. *Prog. in Oceanogr.* 120,
821 120–138.

822 Le Hénaff, M., Kourafalou, V. H., Morel, Y., Srinivasan, A., 2012. Simulating
823 the dynamics and intensification of cyclonic Loop Current Frontal Eddies
824 in the Gulf of Mexico. *J. Geophys. Res.* 117, C02034.

825 Leben, R., 2005. Altimeter-derived Loop Current metrics. In: Sturges, W.,
826 Lugo-Fernández, A. (Eds.), *Circulation in the Gulf of Mexico: Observa-
827 tions and Models*. Vol. 161 of Geophysical Monograph Series. American
828 Geophysical Union, Washington, D.C., pp. pp. 181–201.

829 Leben, R. R., Born, G. H., Engebret, B. R., 2002. Operational altimeter
830 data processing for mesoscale monitoring. *Marine Geodesy* 25, 3–18.

831 Leben, R. R., Hall, C. A., Hoffman, N. G., 2012. The relationship between
832 Loop Current separation period and retreat latitude revisited. In: McKay,

- 833 M., Nides, J. (Eds.), Proceedings: Twenty-Sixth Gulf of Mexico Informa-
 834 tion Transfer Meeting, March 2011. U.S. Dept. of the Interior, Bureau of
 835 Ocean Energy Management, Gulf of Mexico OCS Region: New Orleans,
 836 LA, pp. 37–41 (OCS Study BOEM 2012–107).
- 837 Lin, I.-I., Wu, C.-C., Pun, I.-F., Ko, D.-S., Sep. 2008. Upper-ocean thermal
 838 structure and the Western North Pacific category 5 typhoons. Part I: Ocean
 839 features and the category 5 typhoons’ intensification. *Mon. Weather Rev.*
 840 136 (9), 3288–3306.
- 841 Liu, Y., Weisberg, R. H., 2011. Evaluation of trajectory modeling in different
 842 dynamic regions using normalized cumulative Lagrangian separation. *J.*
 843 *Geophys. Res.* 116, C09013.
- 844 Liu, Y., Weisberg, R. H., 2012. Seasonal variability on the West Florida Shelf.
 845 *Prog. in Oceanogr.* 104, 80–98.
- 846 Liu, Y., Weisberg, R. H., Vignudelli, S., Mitchum, G. T., 2014. Evalua-
 847 tion of altimetry-derived surface current products using lagrangian drifter
 848 trajectories in the eastern Gulf of Mexico. *J. Geophys. Res. Oceans* 119,
 849 2827–2842.
- 850 Lugo-Fernández, A., 2007. Is the Loop Current a chaotic oscillator? *J. Phys.*
 851 *Oceanogr.* 37, 471–489.
- 852 Lugo-Fernández, A., Leben, R. R., 2010. On the linear relationship be-
 853 tween Loop Current retreat latitude and eddy separation period. *J. Phys.*
 854 *Oceanogr.* 40 (12), 2778–2784.

- 855 Maul, G. A., 1977. The annual cycle of the Gulf Loop Current, part I, Ob-
856 servations during a one-year time series. *J. Mar. Res.* 35, 29–47.
- 857 Meinen, C. S., Watts, D. R., 2000. Vertical structure and transport on a
858 transect across the North Atlantic Current near 42°N: Time series and
859 mean. *J. Geophys. Res.: Oceans* 105 (C9), 21869–21891.
- 860 Metzger, E. J., Smedstad, O. M., Thoppil, P. G., Hurlburt, H. E., Cummings,
861 J. A., Wallcraft, A. J., Zamudio, L., Franklin, D. S., Posey, P. G., Phelps,
862 M. W., Hogan, P. J., L., B. F., DeHaan, C. J., 2014. US Navy operational
863 global ocean and Arctic ice prediction systems. *Oceanography* 27 (3), 32–
864 43.
- 865 Nedbor-Gross, R., Dukhovskoy, D. S., Bourassa, M., Morey, S. L., Chas-
866 signet, E. P., 2014. Investigation of the relationship between the Yucatán
867 Channel transport and the Loop Current area in a multi-decadal numerical
868 simulation. *Mar. Technol. Soc. J.* 48 (4), 15–26.
- 869 Nguyen, T.-T., Morey, S. L., Dukhovskoy, D. S., Chassignet, E. P., 2015.
870 Nonlocal impacts of the Loop Current on cross-slope near-bottom flow in
871 the northeastern Gulf of Mexico. *Geophys. Res. Lett.* 42 (8), 2926–2933.
- 872 Oey, L. Y., 2008. Loop Current and deep eddies. *J. Phys. Oceanogr.* 38,
873 1426–1449.
- 874 Paris, C. B., Le Hénaff, M., Aman, Z. M., Subramaniam, A., Helgers, J.,
875 Wang, D.-P., Kourafalou, V. H., Srinivasan, A., 2012. Evolution of the Ma-
876 condo well blowout: Simulating the effects of the circulation and synthetic

877 dispersants on the subsea oil transport. *Environ. Sci. Technol.* 46 (24),
878 13293–13302.

879 Park, J.-H., Watts, D. R., Donohue, K. A., Tracey, K. L., 2012. Compar-
880 isons of sea surface height variability observed by pressure-recording in-
881 verted echo sounders and satellite altimetry in the Kuroshio Extension. *J.*
882 *of Oceanogr.* 68 (3), 401–416.

883 Schmitz, W. J. J., 2005. Cyclones and westward propagation in the shed-
884 ding of anticyclonic rings from the Loop Current. In: Sturges, W., Lugo-
885 Fernández, A. (Eds.), *Circulation in the Gulf of Mexico: Observations and*
886 *Models*. Vol. 161 of *Geophysical Monograph Series*. American Geophysical
887 Union, Washington, D.C., pp. 241–261.

888 Scott, R., Arbic, B. K., Chassignet, E. P., Coward, A. C., Maltrud, M.,
889 Srinivassan, A., Vargese, A., 2010. Total kinetic energy in four global ed-
890 dyding ocean circulation models and over 5000 current meter records. *Ocean*
891 *Modeling* 32, 157–169.

892 Shay, L. K., Jaimes, B., Brewster, J. K., Meyers, P., McCaskill, E. C.,
893 Uhlhorn, E., Marks, F., Halliwell Jr., G. R., Smedstad, O. M., Hogan, P.,
894 2011. Airborne ocean surveys of the Loop Current complex from NOAA
895 WP-3D in support of the Deepwater Horizon oil spill. In: Liu, Y., Mac-
896 Fadyen, A., Ji, Z.-G., Weisberg, R. H. (Eds.), *Monitoring and Modeling*
897 *the Deepwater Horizon Oil Spill: A Record-Breaking Enterprise*. Vol. 195
898 of *Geophysical Monograph Series*. American Geophysical Union, Washing-
899 ton, D.C., pp. 131–151.

900 Sturges, W., Evans, J. C., Welsh, S., Holland, W., Feb. 1993. Separation of
901 warm-core rings in the Gulf of Mexico. *J. Phys. Oceanogr.* 23 (2), 250–268.

902 Sturges, W., Leben, R. R., 2000. Frequency of ring separations from the Loop
903 Current in the Gulf of Mexico: A revised estimate. *J. Phys. Oceanogr.* 30,
904 1814–1818.

905 Taylor, K. E., Apr. 2001. Summarizing multiple aspects of model performance
906 in a single diagram. *J. Geophys. Res.* 106 (D7), 7183–7192.

907 Thompson, R. O. R. Y., 1979. Coherence significance levels. *J. of Atmos. Sci.*
908 36 (10), 2020–2021.

909 Vukovich, F. M., 1988. Loop Current boundary variations. *J. Geophys. Res.:*
910 *Oceans* 93 (C12), 15585–15591.

911 Walker, N., Myint, S., Babin, A., Haag, A., 2003. Advances in satellite ra-
912 diometry for the surveillance of surface temperatures, ocean eddies and
913 upwelling processes in the Gulf of Mexico using GOES-8 measurements
914 during summer. *Geophys. Res. Let.* 30 (16), 1854.

915 Watts, D. R., Tracey, K. L., Friedlander, A. I., 1989. Producing accurate
916 maps of the Gulf Stream thermal front using objective analysis. *J. Geophys.*
917 *Res.: Oceans* 94 (C6), 8040–8052.

918 Watts, R. D., Qian, X., Tracey, K. L., 2001. Mapping abyssal currents and
919 pressure fields under the meandering Gulf Stream. *J. Atmos. Ocean Tech.*
920 18, 1052–1067.

- 921 Welsh, S. E., Inoue, M., 2000. Loop Current rings and the deep circulation
922 in the Gulf of Mexico. *J. Geophys. Res.: Oceans* 105 (C7), 16951–16959.
- 923 Xu, F.-H., Chang, Y.-L., Oey, L.-Y., Hamilton, P., 2013. Loop Current
924 growth and eddy shedding using models and observations: analyses of
925 the july 2011 eddy-shedding event. *J. Phys. Oceanogr.* 43, 1015–1027.
- 926 Yablonsky, R. M., Ginis, I., Sep. 2012. Impact of a warm ocean eddy’s cir-
927 culation on hurricane-induced sea surface cooling with implications for
928 hurricane intensity. *Mon. Weather Rev.* 141 (3), 997–1021.
- 929 Yin, X. Q., Oey, L. Y., 2007. Bred-ensemble ocean forecast of Loop Current
930 and rings. *Ocean Modell.* 17, 300–326.
- 931 Zeng, X., Li, Y., He, R., 2015. Predictability of the Loop Current variation
932 and eddy shedding process in the Gulf of Mexico using an artificial neural
933 network approach. *J. Atmos. Oceanic Technol.* 32 (5), 1098–1111.

Figure 21 - Stream Fn Hadal