Dynamics of the Periphery Current in Rhode Island Sound

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Abstract

Observations reveal a near-surface circulation around the periphery of Rhode Island Sound (RIS) that occurs in summer stratified conditions and disappears in winter when weak solar insolation and wind stirring result in strong vertical mixing. According to a series of numerical simulations and theoretical analysis, we attribute the summer intensification of this “periphery current” to a circulation produced by seasonal bottom thermal fronts often observed from May to September. The strength of the thermal fronts is proportional to the surface solar radiation. Meanwhile, the simulations capture a continuous topographically rectified tidal residual current in RIS and a pair of opposite-sign headland eddies around Montauk Point in Block Island Sound (BIS). Our analysis suggests the importance of nonlinear vorticity advection and frictional torques in BIS. On the other hand, the vorticity balance shows the importance of velocity torque by contributing to the opposite headland gyres, while the effect by planetary vorticity stretching is negligible over BIS but important in RIS.

Keywords: seasonal variability, coastal ocean, bottom thermal front, tidal mixing, tidal rectification

1. Introduction

Rhode Island Sound (RIS), a semicircular embayment off the southern Rhode Island and Massachusetts coast, is located offshore of the mouth of Narragansett

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29 Bay and bounded to the west by Point Judith and Block Island (BI) and to the
 30 east by Martha's Vineyard and the Elizabeth Islands (Fig. 1). Waters in RIS
 31 are connected with Block Island Sound (BIS) through the gap between Point
 32 Judith and BI. Based on field measurement, Codiga and Ullman (2010) identified
 33 a cyclonic, near-surface circulation around the periphery of RIS that is usually
 34 present in summer, stratified conditions and then disappears in winter when
 35 weak solar insolation and wind stirring results in strong vertical mixing. In
 36 summer, the circulation is characterized by residual current moving along the
 37 coast of Rhode Island, flowing north into RIS from the southeast, to the west at
 38 the inner shelf (Kincaid et al., 2003) where it encounters potentially significant
 39 dynamical exchange processes associated with Narragansett Bay, and then to the
 40 southwest in the western portion of RIS and south of Block Island (Ullman and
 41 Codiga, 2004). A seasonal thermocline accompanies this circulation in spring
 42 and summer (Shonting and Cook, 1970). During the winter months this entire
 43 flow pattern collapses, with much weaker currents and different trajectories
 44 (Codiga and Ullman, 2010).

45 Tidal residual effect is a significant dynamical contributor to the circulation
 46 in RIS (Luo et al., 2013), but the dynamics that gives rise to the strong tide-
 47 generated gyre around RIS is still an open issue. Pingree (1978) and Zimmerman
 48 (1981) introduced a novel approach to understanding the mechanisms of tidal
 49 residual circulations by examining the vorticity of tidal flows, and Robinson
 50 (1981) provided further insights based on the transfer of tidal vorticity to the
 51 mean field. Moreover, Robinson (1981) illustrated the unique role of quadratic
 52 bottom friction when there was a topographic slope perpendicular to the tidal
 53 residual currents. To fully understand the sources of tidal residual circulations,
 54 Ridderinkhof (1989) used the depth-averaged vorticity equation to show that
 55 the essential balance was provided by the advection and dissipation of tidal
 56 vorticity generated by column stretching and squeezing, and by the frictional
 57 torque due to depth gradients and velocity shear. One purpose of this study is
 58 to detail the dynamics of the tidal residual circulation in and around RIS.

59 In addition to the tidal residual current, another competing hypothesis for

60 the circulation in RIS involves a thermohaline circulation. The circulation is
 61 in thermal wind balance with the sharp horizontal, temperature and salinity
 62 fronts, which are usually present in spring and summer. The horizontal fronts
 63 are produced by the spatially varying distribution of tidal mixing (Hill et al.,
 64 2008), which has been found in many other shallow tidal areas in the Northern
 65 Hemisphere (Horsburgh, 2000). Codiga and Ullman (2010) suggested the pos-
 66 sibility that the summer intensification of the RIS periphery flow was tied to
 67 sharpened horizontal density gradients associated with tidal mixing fronts.

68 In addition to the periphery current, a pair of eddies around Montauk Point
 69 was present in the numerical simulations of both Edwards (2004) and Luo et al.
 70 (2013). Similar closely-connected headland eddy pairs have been found in other
 71 regions and their dynamics have been attributed to many different mechanisms,
 72 with bottom frictional torque (Pingree, 1978; Geyer and Signell, 1990) and no-
 73 slip lateral ‘wall’ boundary friction proposed by Zimmerman (1981) being the
 74 most popular. In the mechanism by Zimmerman (1981), the velocity gradient
 75 normal to the wall is associated with the wall’s lateral frictional boundary layer
 76 that is then transported to the mean field from the oscillating tidal field by the
 77 nonlinear advection of vorticity. Based on the analysis of the transport vorticity
 78 equation around an idealized, symmetric promontory in a numerical simulation,
 79 Park and Wang (2000) attributed the eddies to the transfer of topographic
 80 vorticity, which is consistent with the interpretation by Robinson (1981). In
 81 essence, the side-wall boundary works the same way in generating the velocity
 82 gradients as a slope that gradually shoals around a promontory (Zimmerman,
 83 1981). In BIS, the topographic features make the situation rather more com-
 84 plicated. Edwards (2004) has emphasized the importance of vortex stretching
 85 and tilting for the anticyclonic eddy south of Montauk Point without, however,
 86 connecting it with the cyclonic eddy to the north of Montauk Point.

87 The purpose of this study is to examine the sources of the topographically
 88 rectified residual currents in RIS and BIS, and to investigate the role of both
 89 tidal residual currents and deep thermohaline gradients in the summer circu-
 90 lation in RIS through both a series of three-dimensional numerical simulations

and theoretical analysis. The numerical model configuration, design and model validation are described in Sections 2 and 3. The tides over RIS and BIS are depicted in Section 4, and the contributions from bottom thermal gradients are presented in Section 5. In Section 6, we discuss the tidal rectification process, and a criterion is used to quantitatively describe the contributions from steric height and non-steric height. A summary is given in Section 7.

2. Methods

The numerical simulations are performed using the Regional Ocean Modeling System (ROMS; Shchepetkin and McWilliams, 1998, 2003, 2005; Haidvogel et al., 2000) that employs a terrain-following coordinate for the vertical. The high resolution ‘local’ domain covers RIS, BIS, Long Island Sound (LIS) and the adjacent inner shelf area, with a uniform horizontal grid of 800 m and 30 terrain-following vertical layers, and this domain is one-way nested within a ‘regional’ domain with a constant horizontal resolution of 5 km (Fig. 2). For the regional-scale model, the open boundary conditions are derived from a global eddy resolving model simulation produced by the Hybrid Coordinate Ocean Model (HYCOM)/Navy Coupled Ocean Data Assimilation (NCODA; Chassignet et al., 2003; Cummings, 2005).

The model uses the Mellor Yamada level 2.5 scheme for turbulence closure (Mellor and Yamada, 1972) with the background vertical mixing coefficient of $5.0 \times 10^{-6} \text{ m}^2/\text{s}$ and a quadratic bottom drag formulation. The free surface and vertically integrated boundary conditions are specified by Chapman (1985) and Flather (1976), respectively. Compared with the former configuration used in Luo et al. (2013), we have improved the resolutions over Narragansett Bay to better capture its dynamical exchanges with RIS, and increased the number of vertical layers from 15 to 30 to better resolve the vertical structure, especially in the bottom boundary layer. In all the experiments, we set WTYPE as 5, which corresponds to Jerlov’s water type III (Paulson and Simpson, 1977; Jerlov, 1976).

Our ‘realistic’ experiment (REAL) is driven by atmospheric forcing, observed

120 river discharges and tides in an effort to simulate the seasonal variability of
 121 RIS circulation (Table 1). Daily averaged climatological surface atmospheric
 122 fluxes (including winds, air temperature, air pressure, relative humidity, rainfall
 123 rate, short wave, and long wave radiations) from 2004 to 2009 were obtained
 124 from the North American Regional Reanalysis (NARR; Mesinger et al., 2006).
 125 Bulk formulas are applied to calculate latent and sensible heat fluxes. It is
 126 recognized that, as emphasized by He et al. (2004) and MacCready et al. (2009),
 127 accurate high-frequency surface forcing is important for numerical models to
 128 simulate detailed features of coastal ocean processes. Our simulations may not
 129 optimally capture characteristics of flow at timescales less than one day due to
 130 our use of daily-average surface forcing, which reduces the wind variance, and
 131 thus the variance of wind-driven currents on such short timescales. However,
 132 our use of daily-average surface forcing is defensible based on the reasonable
 133 expectation that nonlinear rectification due to higher-frequency fluctuations in
 134 surface forcing is not strong, so remains a weaker influence on the dynamics
 135 than the fully-resolved tidal and subtidal timescales that are the main focus of
 136 our investigations.

137 The averaged wind from 2004 to 2009 (Fig. 2) is generally southeastward in
 138 winter when it is strongest and northeastward in summer, consistent with the
 139 data at the deployments carried out by the Ocean Special Area Management
 140 Plan (OSAMP) in 2009 and 2010 (Ullman and Codiga, 2010). Therefore, the
 141 forcing to the model is similar to the observations. For verification purpose, we
 142 have carried out another similar experiment driven by original, daily, NARR
 143 forcing from 2004 to 2009. There is minor difference in monthly mean cur-
 144 rents between these two experiments, which indicates that the modeled cyclonic
 145 circulation is not an anomalous case. Since the dynamics behind these two ex-
 146 periments is the same, the experiment driven by climatological mean of forcing
 147 is used.

148 Daily river discharges for the major rivers (from north to south - Taunton
 149 River, Blackstone River, Pawtuxet River, Pawcatuck River, Connecticut River,
 150 and Housatonic River) were obtained from the United States Geological Survey

151 (USGS; <http://waterdata.usgs.gov/nwis>), and we re-scaled these to account for
 152 the ungauged portions of the watershed to improve the salinity in LIS and BIS
 153 (Zhang et al., 2010). This re-scaling has improved the discrepancy between
 154 model salinity results and observations from a maximum difference of ~ 3 PSU
 155 (using un-scaled data) to within 0.5 PSU. REAL utilizes 5 major tides (M_2 ,
 156 N_2 , S_2 , O_1 and K_1), which represent 80% of the variance of the total depth-
 157 averaged tidal kinetic energy as obtained from the Advanced Circulation Model
 158 for Oceanic, Coastal and Estuarine Waters (ADCIRC) tidal simulation (Luettich
 159 et al., 1992).

160 To study the topographically-rectified tidal current, we configured a hierar-
 161 chy of process-oriented experiments, specifically targeted at understanding the
 162 fundamental dynamics of tidal rectification mechanisms associated with topog-
 163 raphy, Coriolis force, and momentum advection. To simplify the problem, only
 164 the M_2 tide, which accounts for 85% to 90% of the energy from the five major
 165 constituents, was retained in all of these experiments. Because of the unique
 166 role played by quadratic bottom friction, the quadratic bottom drag formula-
 167 tion was used in all tidal simulations with a drag coefficient of $C_D = 7.5 \times 10^{-3}$.
 168 These process-oriented experiments are shown in Table 1 and defined as fol-
 169 lows: ‘RealM2’ - realistic bathymetry, with both Coriolis force and momentum
 170 advection retained; ‘Flat’ - ‘RealM2’ but with a flattened bottom; ‘NoRotate’ -
 171 ‘RealM2’ but non-rotating; and ‘NoAdv’ - RealM2 without momentum advec-
 172 tion.

173 To examine the effects of winds and remote forcing on the seasonal variability
 174 of the circulation, in addition to the experiment REAL, two experiments are
 175 undertaken (Table 1). Compared with REAL, the case NoWind removes the
 176 effects of winds; the experiment NoRemote does not have the remote effect. In
 177 contrast to the open boundary conditions in Real, in NoRemote, we analytically
 178 set the averaged velocities across the open boundaries as 0.

179 Two additional process-oriented experiments, as shown in Table 1, are set
 180 to explore the contribution of bottom intensified thermohaline gradients for
 181 balancing the summer-intensified circulation. The first (‘HeatStr’) is driven by

182 climatological mean surface heating in August and starts from a homogenous
 183 rest state. The second ('HeatWk') is driven with half the solar radiation of
 184 'HeatStr'. We run both experiments for three months and examine the time-
 185 averaged results in the third month.

186 3. Model-Observations Comparisons

187 Before performing an analysis of the model results, we validate our simu-
 188 lations through comparisons with all available observations. Luo et al. (2013)
 189 have shown that our realistically forced model of the region does a good job
 190 of simulating tides over the whole domain by [comparisons of](#) tidal ellipses and
 191 sea level elevations, as well as temperature and salinity on the outer shelf. In
 192 Fig. 3, we also validate model tidal sea level variations with observations from
 193 bottom-mounted acoustic Doppler current profilers (ADCP) at stations PO-S
 194 and PO-F, whose locations are marked in Fig. 1 (Ullman and Codiga, 2010).
 195 Root mean square errors (RMSEs) for sites PO-S and PO-F are 0.16 m and 0.04
 196 m, respectively.

197 We next compare the seasonal mean, depth-averaged model currents with
 198 moored observations from 2009 and 2010 (Fig. 4) that are a combination of
 199 data from the Rhode Island Ocean Special Area Management Plan project and
 200 an additional project funded by Rhode Island Sea Grant and Rhode Island
 201 Endeavor Program (Wertman et al., 2016, in preparation). In summer, the
 202 model successfully captures the westward jet offshore of Narragansett Bay as
 203 well as the southwestward current leaving RIS that appears in the observations.
 204 Both the model and observations show relatively weak onshore velocities in RIS,
 205 with stronger currents (~ 15 cm/s) located to the south of Block Island in a
 206 region that connects RIS and BIS. In winter, the seasonal mean depth-averaged
 207 currents are diminished for both model and observations, and the directions are
 208 more variable. Because the currents over the studied region are influenced by
 209 winds, and winds during winter are strong and highly variable, the comparisons
 210 are acceptable. The RMSEs for the u- and v-component velocities in summer

211 are 2.2 cm/s and 0.8 cm/s , respectively; in winter, 1.9 cm/s and 1.7 cm/s ,
212 respectively.

213 On the inner shelf of RIS, close to the mouth of Narragansett Bay, the model
214 captures the seasonal variability of circulation and water exchange between Nar-
215 ragansett Bay and RIS that was observed by acoustic Doppler current profiler
216 surveys between April 1998 and July 1999 by Kincaid et al. (2003) (Table 2).
217 Consistent with the surveys, during summer, a relatively strong transport of
218 $\sim 4000 \text{ m}^3/\text{s}$ enters the survey area through the Brenton Reef line; weaker trans-
219 ports flow into the survey region through the East Passage and West Passage
220 lines (please see Fig. 1 for the location of these lines). Most of the transport into
221 the region comes through the Narragansett Beach line, with a relatively smaller
222 transport through the South Line. During winter, water exchange over the
223 survey region changes significantly with a much-diminished transport through
224 the Brenton Reef, Narragansett Beach and South Line, even reversing direction
225 through the Brenton Reef and South Line in January and February.

226 Compared with the observations, in winter, the transport through the West
227 Passage is underestimated; In summer, the model produces a much smaller
228 transport through Narragansett Beach and a larger transport through the South
229 Line. This discrepancy may be due to the model’s treatment of rivers in Narra-
230 gansett Bay, in which we include all river input to Narragansett Bay from the
231 three major rivers (Pilson, 1985). Nevertheless, the correspondence in seasonal
232 variation of transport and monthly mean currents justifies using the model to
233 study the dynamics of the circulation in RIS.

234 4. Tidal Experiments

235 4.1. Tidal Residual Circulation

236 Previous studies by Luo et al. (2013) suggested that our model well captures
237 the tides by tidal ellipses and sea level elevations comparisons. Fig. 5 shows the
238 modeled depth-averaged tidal ellipses over the studied region. Consistent with
239 the CODAR observations (Ullman and Codiga, 2010), over BIS, most of the

240 ellipses rotate clockwise. Because of the resonance of tides at the mouth of LIS,
 241 the ellipses are highly elliptical with the semi-major axis reaching to 1 m/s ,
 242 while the semi-minor axis much smaller. Most of the ellipses throughout RIS
 243 rotate clockwise, except those in its northwestern portion, to the southwest of
 244 Narragansett Bay. Compared with BIS, the RIS ellipses are much smaller, with
 245 semi-major axis smaller than 10 cm/s , and they are larger at the offshore than
 246 at the onshore region.

247 The tidal residual circulation driven by M_2 is characterized by a weak (~ 3
 248 cm/s) peripheral current around RIS (Fig. 6). It originates from the south of
 249 Martha's Vineyard, moves southwestward toward Block Island and eventually
 250 joins the circulation around Block Island. In BIS and LIS, the residual current
 251 forms an eddy pair around Montauk Point, between which there is an eastward
 252 current that is larger than 5 cm/s . This eddy pair pattern has also been found
 253 in different regions (e.g., Pingree, 1978; Geyer and Signell, 1990). In addition,
 254 a weak cyclonic eddy is present to the east of the strong anti-cyclonic eddy
 255 (southwest of Block Island). Because of the complex regional geography and
 256 topography and stronger tidal currents, the generation mechanism(s) that give
 257 rise to these eddies is rather complex and will be discussed more in Section 6.

258 Fig. 6 also shows the relative vorticity in the region of interest. Residual
 259 eddies in RIS includes positive relative vorticity eddies inside most of RIS, neg-
 260 ative relative vorticity eddies along the coast where currents encounter irregular
 261 promontories, and negative relative vorticity eddies in the west of RIS. The pos-
 262 itive relative vorticity eddies inside RIS are set by a residual current that moves
 263 cyclonically inside RIS and along its periphery. Flow moving out of RIS to the
 264 east of BI is identified with the negative relative vorticity eddies in the west
 265 of RIS. In the southeast of RIS, around Gay Head, a pair of oppositely signed
 266 eddies occurs, with the positive cyclonic eddy to the west and anticyclonic eddy
 267 to the east. As an important feature generated by tides, this pattern has been
 268 captured by both observations (Geyer and Signell, 1990) and numerical simu-
 269 lations (Park and Wang, 2000), and has been ascribed to a balance between
 270 vorticity advection and topographic vorticity.

271 4.2. Process-oriented Experiments

272 In this study we configure a hierarchy of process-oriented experiments to
 273 understand the role of topography, Coriolis force and nonlinear advection in
 274 setting this circulation. Fig. 7 shows the depth-averaged tidal-rectified residual
 275 currents for experiments a) 5 major Tides b) RealM2, c) Flat, d) NoRotate,
 276 and e) NoAdv. In each case the currents are averaged for the last month of a
 277 3-month simulation. There is little difference between the experiments driven
 278 by five major tides and by M_2 (Figs. 7a,b).

279 The experiment, Flat, is a flat bottom simplification of experiment RealM2.
 280 The comparisons of the circulations between RealM2 and Flat (Figs. 7b, c) show
 281 rather conclusively that the major tidal residual current patterns in RIS and
 282 BIS are mainly topographically-rectified. For example, due to the absence of
 283 bottom topography the eddy pair around Montauk Point is significantly weaker,
 284 as is the cyclonic eddy south of the eddy pair. Moreover, the anticyclonic eddy
 285 around BI and peripheral circulation around RIS almost disappears, leaving a
 286 relatively sluggish westward flow in the northeast region of RIS. We therefore
 287 conclude that topography is the dominant contributor to the residual current
 288 in the study region.

289 The experiment NoRotate emphasizes that we cannot ignore the role of the
 290 Coriolis force for dynamical understanding of the periphery current in RIS. In
 291 contrast to RealM2, NoRotate does not produce an anticyclonic circulation in
 292 RIS; instead it generates weak, negative relative vorticity over most of RIS.
 293 The current flowing out of Vineyard Sound, the source water for the periphery
 294 current, flows directly offshore out of the sound instead of diverting to the
 295 right of the current direction as would be the case with rotation. We therefore
 296 conclude that, in addition to topography the Coriolis force needs to be included
 297 in our dynamical understanding of the periphery current, consistent with the
 298 studies of Huthnance (1973) and Loder (1980).

299 Even though an eddy pair is generated in the west of BIS in experiment
 300 NoRotate, as compared with the eddy structure generated by RealM2 the en-
 301 tire paired eddy structure is tilted cyclonically and is weaker. In addition, less

302 water is trapped in the eddy north of Montauk Point, with stronger flow pass-
 303 ing through the gap between Point Judith and Block Island, contributing to a
 304 broader anti-cyclonic circulation east of Block Island. Besides, the weak cyclonic
 305 eddy southwest of Block Island is smaller and weaker. Considering the differ-
 306 ences in these eddy structures between experiments RealM2 and NoRotate, we
 307 can conclude that mechanisms associated with the Coriolis force is also impor-
 308 tant for dynamical understanding of the eddies in BIS. On the other hand, the
 309 fact that completely omitting the Coriolis force does not eliminate the eddies
 310 suggests that other mechanisms independent of earth rotation but associated
 311 with bottom slope apparently play a more important role, which motivates a
 312 detailed analysis of the vorticity balance in the next section.

313 To conclude this set of experiments, we turn off momentum advection (NoAd-
 314 vect) and obtain a very simple and weak circulation without any of the major
 315 circulation patterns discussed above. According to Pingree (1978) and Rid-
 316 derinkhof (1989), the essential tidal rectification process is the advection and
 317 dissipation of vortices generated by periodic tidal currents, i.e. it is advection
 318 that transfers vorticity from the periodic tidal currents to the residual circula-
 319 tion. Hence no obvious residual current is generated without advection terms.

320 In RIS, the tidal residual circulation has the same direction, that is, west-
 321 ward, as the observations shown in Fig. 4. However, its magnitude is much
 322 smaller, and away from the coast, there is a large difference between the ob-
 323 servations and the tide-only simulation. This indicates that tides are not the
 324 dominate contributor to the summer circulation.

325 **5. Thermohaline Gradients**

326 *5.1. Seasonal Variability*

327 Seasonally averaged surface circulations in summer and winter from our
 328 REAL experiment are shown in Figs. 9a, b. Consistent with the limited obser-
 329 vations (Kincaid et al., 2003; Ullman and Codiga, 2010), the summer circulation
 330 features a peripheral circulation within RIS originating from the southwest of

Martha's Vineyard, moving along the coast of Rhode Island, and flowing out of RIS to the southeast of Block Island before joining the southwestward downshelf flow in BIS. The circulation pattern changes dramatically in winter when the cyclonic pattern in RIS essentially collapses, especially in the middle and eastern regions of RIS. The current out of BIS steers offshore, however its magnitude does not change as significantly as RIS over the seasonal cycle.

The seasonal variations in RIS are also reflected in vertical structure as shown in Figs. 9c, d. The section is from Block Island to the east of RIS (black line in Fig. 8). During summer, when strong solar insolation occurs, the transitions between the tidally mixed regions (in eastern and western RIS) and the stratified region in the interior of RIS are shown by the robust deep thermal fronts at the ends of the section with weak thermal fronts near the surface. Accompanying the deep thermal fronts is a strong northward jet along the eastern portion of the section and a strong southward jet along the west, defining the peripheral circulation around RIS. During this time period, salinity remains much less stratified at both inner shelf and outer shelf with a horizontal change of <0.2 PSU over this region.

During winter, with the significant decrease of insolation and an increase of surface torque due to an increase in the seasonally varying winds, both temperature and salinity are vertically homogenized over most of the region. At the same time, currents in the top layers turn offshore and diminish to <1 cm/s except for the region within 10 km off Block Island that was influenced by the flow from BIS through the gap between Block Island and Point Judith.

5.2. *Winds and Remote-forcing Effects*

Experiments NoWind and NoRemote are used to examine the importance of winds and remote effects on the seasonal variability of the circulation. Figs. 10a, b show the summer and winter circulation for the experiment NoWind, and Figs. 10c, d for the case NoRemote. Similar to the case REAL, both experiments show strong seasonal variability of the circulation with robust summer currents around the periphery of Rhode Island. The strong circulation disappears in

361 winter.

362 Without remote effects, in summer and winter, the currents entering RIS
363 through the south boundary of RIS disappears. Because of mass conservation,
364 the circulation in RIS becomes complicated. However, the robust peripheral
365 current around RIS and the southwestward current to the south of Martha’s
366 Vineyard still exist. This indicates that winds and remote effects are not re-
367 sponsible for the summer intensification of the circulation.

368 5.3. Bottom Thermal Fronts in Idealized Experiments

369 In coastal seas, a transition between the stratified, offshore region and the
370 tidally mixed, onshore region often generates bottom and surface thermal fronts,
371 which are associated with tidal mixing fronts (Hill et al., 2008; Takeoka et al.,
372 1997). Simpson and Hunter (1974) used an empirical criterion, H/U^3 (where
373 H is the water depth and U is the root-mean-square value of depth-averaged
374 tidal velocity), to estimate the locations of tidal mixing fronts. The criterion
375 describes the energy balance between the demand for potential energy for a fully
376 mixed condition and the supply of vertical mixing energy from the tidal current,
377 with small values representing efficient vertical mixing.

378 Fig. 8 plots the $\log(H/U^3)$ for M_2 tidal currents in RIS and BIS. It shows
379 intense mixing over BIS and eastern and western RIS. In BIS, because buoyancy
380 is also due to the freshwater entrained out of LIS and not just the tidal flow,
381 this diagnostic does not capture the full story of frontal structure. However in
382 RIS, especially away from the mouth of Narragansett Bay, freshwater effect is
383 weak (Codiga and Ullman, 2010), and it is therefore good to use this criterion
384 to predict the locations of thermal fronts.

385 We examine the impacts of the bottom and surface thermal fronts on the
386 summer circulation by analyzing the results of experiments HeatStr and HeatWk
387 shown in Table 1. Monthly mean surface circulations from these two cases
388 are shown in Figs. 11a, b. Consistent with the REAL simulation, currents in
389 both experiments feature a circulation around RIS. Differences in the surface
390 currents (HeatStr-HeatWk; Fig. 12a) give a circulation difference pattern in the

391 cyclonic direction in RIS, demonstrating that a surface heating reduction slows
 392 the peripheral circulation around the periphery of RIS by ~ 5 cm/s. In the
 393 interior of RIS, the circulation difference is negligible. Between the peripheral
 394 region with westward circulation and the interior of RIS, there is a weak (~ 1
 395 cm/s) eastward **current** (Fig. 12a), corresponding to a wider cyclonic jet in
 396 HeatWk than in HeatStr.

397 Fig. 12b shows the cross-sectional structures of temperature (black contours)
 398 and velocity (colors) across the section defined in Fig. 8. At the ends of the
 399 section where $\log(H/U^3) \leq 4.2$, water columns are vertically mixed by tidal
 400 mixing. In the interior where $\log(H/U^3) \geq 5$, stratification is strong due to
 401 weakening of mixing. Temperature contrasts between the vertically mixed wa-
 402 ter columns and the deeper, stratified water column form a prominent bottom
 403 thermal front and a less prominent surface front. By thermal wind balance, the
 404 bottom thermal front drives a westward circulation at the surface, while the
 405 surface thermal front tends to generate an opposite circulation (Fig. 13). Thus
 406 the surface westward circulations in Figs. 11a, b emphasize the importance of
 407 the bottom thermal front.

408 Given that different pools of water along the cross-shelf direction respond
 409 to surface heating reductions differently (Hill et al., 2008), changes in thermal
 410 fronts are expected as follows. Because surface heating penetrates to a deep
 411 depth at the inner shelf shallower waters, the temperature reduction for the inner
 412 shelf waters is obvious over the entire water column. At the offshore deep waters,
 413 surface heating only affects the surface warm pool, and the **isolated** bottom
 414 cold pool keeps intact. Therefore, we expect reductions in both surface and
 415 bottom thermal fronts, with clear dynamical implications. From experiments
 416 HeatStr to HeatWk, theoretically, the bottom thermal front tends to generate a
 417 weaker westward circulation around RIS, while the surface thermal front tends
 418 to generate a weaker eastward circulation, as shown in Fig. 13.

419 Consistent with the above analysis, a decrease in shortwave radiation weak-
 420 ens the bottom and top thermal fronts. Differences between these two exper-
 421 iments (HeatStr - HeatWk; Fig. 12b) show a significantly (~ 5 cm/s) reduced

422 northward velocity at the eastern end of the section, and a significantly (~ 5
 423 cm/s) reduced southward velocity at the western end, together comprising a
 424 weaker cyclonic circulation in experiment HeatWk than in HeatStr. The robust
 425 changes at the ends of the section are accompanied by weak changes in deeper
 426 regions and moving towards the interior. Specially, from 10 km to 30 km off
 427 Block Island, stronger southward velocities occur in experiment HeatWk; from
 428 30 km to 45 km, stronger northward velocities occur in HeatWk, corresponding
 429 to the weak (~ 1 cm/s) eastward circulation difference in Figs. 11a that is found
 430 offshore of the stronger westward difference. According to the above discussion
 431 of the schematic shown in Fig. 13, this feature is attributed to the weakening of
 432 the shallow thermal front in experiment HeatWk, and it can explain why that
 433 experiment exhibits a wider westward jet than experiment HeatStr. Moving to-
 434 ward the interior of RIS from the inner shelf region, a strong top thermal front
 435 in experiment HeatStr limits the westward circulation generated by the bottom
 436 thermal front (where it gets relatively weaker), while in experiment HeatWk
 437 the decrease of the surface thermal front weakens the resistance to the cyclonic
 438 vorticity generated by the bottom thermal front.

439 Overall, the circulation change around the periphery of RIS simulated in
 440 the case HeatWk is a direct consequence of the reduction of surface heating
 441 and its corresponding bottom thermal fronts, while the change in the width of
 442 the jet is due to the changes in surface thermal fronts. Therefore, we conclude
 443 that bottom thermal fronts are essential to the intensified peripheral circulation
 444 around RIS in summer, with the surface thermal front adjusting the width of
 445 the jet. Conducting a similar analysis, we attribute the stronger downshelf
 446 current along Long Island to the formation of a stronger bottom thermal front,
 447 agreeing with the study of Ullman and Codiga (2004) who found an increase in
 448 the horizontal density gradient by analysis of historical hydrographic data.

449 5.4. *Geostrophic Balance*

450 To further study this, we calculate the geostrophic component using the
 451 thermal wind balance, by casting the equations into a right-handed coordinate

452 system with x directed upshelf along the shore of Rhode Island and y directed
 453 onshore. The thermal wind balance is:

$$\frac{\partial v}{\partial z} = -\frac{g}{\rho f} \frac{\partial \rho}{\partial x} \quad (1)$$

$$\frac{\partial u}{\partial z} = +\frac{g}{\rho f} \frac{\partial \rho}{\partial y} \quad (2)$$

455 where z is taken to be vertically upwards, g is the gravitational acceleration,
 456 u is the eastward current and v is the northward current. To compute the
 457 absolute velocities from the thermal wind equation we assume a static bottom
 458 pool, $u = v = 0$, following Horsburgh (2000). Geostrophic calculations are
 459 performed to test whether the observed residual flow pattern is in agreement
 460 with the pattern obtained from the baroclinic forcing.

461 The seasonal mean velocities calculated from the thermal wind equation
 462 and the full-physics model results for the case REAL are shown in Fig 14.
 463 Geostrophic currents in winter are weak, as is expected during this well-mixed
 464 season. In summer, the strong westward jet along the shore of Rhode Island
 465 and the weaker offshore current appear in both the thermal wind calculation
 466 and the full-physics model results. The flow is mainly geostrophic in summer,
 467 except at depth, which is possibly an artifact from the simplified thermal wind
 468 calculation without considering the bottom Ekman layer and from arbitrarily
 469 taking the bottom pool as static - according to Fig. 9c, the bottom currents are
 470 weak, within 2 cm/s. Nevertheless, in summer geostrophic currents dominate
 471 the cyclonic circulation in RIS.

472 6. Discussion

473 6.1. Tidal Rectification over BIS

474 According to the process-oriented, idealized experiments, Coriolis effect is
 475 important for the generation of the tidal residual current in RIS. The generation
 476 of cyclonic circulation over RIS by the Coriolis effect and nonlinear transfer of
 477 vorticity is similar to the generation of the residual current over the Georges
 478 Bank (Loder, 1980) and over sand banks (Zimmerman, 1981).

479 The generation mechanism around Montauk Point is more complicated than
 480 RIS due to its complex topography and tidal currents. In the following, we focus
 481 on the tidal rectification over BIS. The essential tidal rectification process is the
 482 advection and dissipation of vortices generated by periodic tidal currents (Pin-
 483 gree, 1978; Ridderinkhof, 1989). We follow Robinson (1981) and Ridderinkhof
 484 (1989) and use the vorticity equation to analyze the tidal rectification over BIS.
 485 The vorticity budget can be written as:

$$\begin{aligned}
 \underbrace{\frac{\partial \omega}{\partial t}}_a + \underbrace{u \frac{\partial \omega}{\partial x} + v \frac{\partial \omega}{\partial y}}_b = & \underbrace{-\omega \nabla \cdot \vec{u}}_c - \underbrace{f \nabla \cdot \vec{u}}_d - \underbrace{\frac{k \sqrt{u^2 + v^2}}{H^2} \vec{u} \times \nabla h}_e \\
 & + \underbrace{\frac{k}{H} \vec{u} \times \nabla (\sqrt{u^2 + v^2})}_f - \underbrace{\frac{k \sqrt{u^2 + v^2}}{H} \omega}_g \\
 & + \underbrace{\nu \nabla^2 \omega}_i
 \end{aligned} \tag{3}$$

486 where $\vec{u}(u, v)$ is the vertically integrated velocity components; H is the
 487 sum of the local water depth (h) and the local tidal elevation above mean
 488 sea level; ω is the relative vorticity of the depth-averaged velocity; f is the
 489 Coriolis parameter; k is the (constant) bottom friction coefficient; and ν is the
 490 horizontal eddy viscosity. The vorticity budget can be addressed with respect
 491 to the following individual terms: the local (a) and advective (b) changes of
 492 relative vorticity; stretching of the water column by relative (c) or planetary (d)
 493 vorticity; frictional torques due to the sloping bottom obliquely transverse to
 494 the tidal currents (e) and the horizontal velocity shear (f); and the damping of
 495 vorticity by bottom friction (g) and horizontal diffusion (i). Term a is identically
 496 zero after averaging over a tidal cycle. Terms e , f and g all originate from the
 497 bottom friction. The term f only occurs when a quadratic bottom friction
 498 formulation is used.

499 To diagnostically analyze the contributions of different terms in the depth-
 500 averaged vorticity equation in RIS and BIS, we plotted depth-averaged vorticity
 501 balance in RIS and BIS. Fig. 15 shows the terms of advecting (in the term

502 b), producing (c , d , e , f) and dissipating ($g + i$) vorticity averaged over 20
503 tidal cycles of hourly numerical output. Consistent with the process-oriented
504 experiments and available studies, advection of vorticity, which is responsible
505 for the nonlinear transfer of vorticity from tidal currents to residual circulation,
506 is essential. Similar to the vorticity analysis around the Gay Head (Park and
507 Wang, 2000), the residual vorticity advection is positive at the left-hand side and
508 negative at the right-hand side.

509 The dominate contributors to the vorticity production are slope torque (the
510 frictional torque associated with a sloping bottom obliquely transverse to the
511 tidal currents) and speed torque (the frictional torque associated with a velocity
512 gradient), and the effect of speed torque is larger, agreeing with Zimmerman
513 (1981). The speed torque is discussed in detail by Zimmerman (1981) using both
514 the processes of residual basin circulation and headland eddies. The planetary
515 and relative vorticity stretching effects are relatively weaker. Consistent with
516 the process-oriented experiments, the effect due to Coriolis term is small over
517 BIS.

518 According to Zimmerman (1981) and Robinson (1981), a strong topographic
519 feature cannot guarantee significant residual vorticity; Only topographic fea-
520 tures having the length scale of the tidal excursion, $L_t = 2U/\omega$, in the direction
521 of the tidal stream can produce a significant residual vorticity. If the vorticity
522 generating region is considerably larger than the tidal excursion, the oscillating
523 tidal currents are not able to produce residual vorticity, except at the ends of
524 the source region. For a topographic feature within the tidal excursion, the
525 term due to the planetary vorticity is independent of the spatial extent of the
526 topographic feature but depends on the proportional depth change, and there-
527 fore is the same for an estuarine or a deeper shelf-sea situation. In contrast, the
528 frictional torque terms (terms e and f in the depth-averaged vorticity equation)
529 are proportional to the velocity of the fluid element as it passes the generating
530 region, and to the length scale of the generating region. Hence for an estuarine
531 and a deeper shelf-sea situation, the terms associated with bottom friction are
532 greatly different. The above discussion can explain why Coriolis term is impor-

533 tant for both of Georges Bank and RIS, while slope torques are only important
 534 in BIS, where robust tidal currents occur.

535 In a numerical model, to capture the topographic features having the greatest
 536 influence in the generation of mean flows, a grid size smaller than the tidal
 537 excursion is necessary (Robinson, 1981). The tidal excursion in our study region
 538 is about 5-8 km. With the tidal excursion larger than 5 km in most of the
 539 studied region, 800 m is an appropriate grid size to solve the tidal residual
 540 current with a reasonable computation time. However, at the inner shelf of
 541 RIS, there are tidal excursions smaller than 800 m. This means our simulation
 542 may underestimate the contributions of the tidal residual current. To better
 543 solve the tidal rectification close to Narragansett Bay, a configuration covering
 544 RIS with a finer horizontal resolution of 200 m is carried out and Fig. 16 shows
 545 its circulation pattern. Its result is the same with the 800 m configuration with
 546 a 3-4 cm/s, cyclonic circulation around RIS, proving 800 m is an appropriate
 547 grid size to capture the tidal circulation in RIS with a reasonable computation
 548 time. *Note that the configuration with 200 m resolution is only used for the*
 549 *idealized, barotropic tidal experiment without remote effects through the open*
 550 *boundaries from the local-scale ROMS.*

551 6.2. Steric Height and Non-Steric Height

552 Following Csanady (1979, 1997) and Ullman and Codiga (2004), a typical
 553 pressure gradient in the across-shelf direction ($\frac{1}{\rho_0} \frac{\partial P}{\partial y}$) can be partitioned into
 554 the part arising from non-steric height and the part from steric height

$$\frac{1}{\rho_0} \frac{\partial P}{\partial y} = g \frac{\partial \eta_{ns}}{\partial y} - g \frac{\partial}{\partial y} \int_{-H_m}^z \epsilon dz' \quad (4)$$

555 where $\eta_{ns} = \eta - \eta_s$, is the surface departure of sea level elevation (η) from
 556 the surface steric height (η_s) relative to a reference level H_m ; The surface steric
 557 height $\eta_s = -\int_{-H_m}^0 \epsilon dz'$; $\epsilon = \frac{\rho}{\rho_0} - 1$, is the dimensionless anomaly of density
 558 (ρ) relative to a reference density ρ_0 .

559 The first term on the right-hand side is due to non-steric height, correspond-
 560 ing to the contribution from across-shelf pressure gradient caused by the along-

shelf current from the upstream, and the second term is due to steric height,
corresponding to the contribution from thermal fronts. Depth averaging the
equation and simplifying it following Csanady (1997) and Ullman and Codiga
(2004), we get

$$\frac{1}{\rho_0 H} \int_{-H}^0 \frac{\partial P}{\partial y} dz' = g \frac{\partial \eta_{ns}}{\partial y} + \frac{g}{H} \frac{\partial}{\partial y} \int_{-H}^0 \epsilon z' dz' - g \frac{\partial}{\partial y} \int_{-H_m}^{-H} \epsilon dz' \quad (5)$$

in which the depth-averaged steric height is written as the sum of the second
and third terms on the right-hand side. As usual, ρ_0 is taken to be the density
of the deepest water, and we take $H_m = 80$ m, the approximate depth along
the southern boundary, and calculate the pressure gradient between the onshore
and offshore ends of section B (marked in Fig. 6) due to the non-steric and steric
heights. When calculating the third term, the integral is along the bottom from
 $-H_m$ to $-H$.

We compare the non-steric part and steric part to quantitatively evalu-
ate the contributions to the cyclonic circulation around RIS from barotropic
and baroclinic components. For the experiment driven by weak surface heat-
ing, the contribution from non-steric height is $-0.75 \times 10^{-5} \text{ m/s}^2$, and from
steric height is $1.16 \times 10^{-5} \text{ m/s}^2$. Doubling surface heating from the idealized
experiment HeatWk to HeatStr, the component of non-steric height becomes
 $-1.11 \times 10^{-5} \text{ m/s}^2$, meanwhile, there is a larger increase in the steric height
contribution, which becomes $1.99 \times 10^{-5} \text{ m/s}^2$, to counterbalance the change in
non-steric height. This is consistent with our theoretical analysis that pointed
out the increase of surface heating generates a stronger cyclonic circulation
around RIS due to the increase of bottom thermal front.

Similar circulation driven by a combination of barotropic and baroclinic com-
ponents has been found in other areas, such as Yellow Sea (Naimie et al., 2000)
and Georges Bank (Loder, 1980), with contributions from steric height caused
by thermal fronts and non-steric height (barotropic component) associated with
wind piling and ambient streams. This study provides a good method to ad-
judge the contributions from different sources, which can be widely used over

589 other regions of the world.

590 7. Summary

591 Dynamical mechanisms associated with the cyclonic periphery current around
592 RIS in summer are discussed based on a series of numerical modeling results and
593 theoretical calculations. The numerical simulations indicate that tidal residual
594 currents and bottom thermohaline gradients are responsible for this summer
595 circulation. The depth-averaged vorticity balance is used to explain the sources
596 of the tidal residual currents in RIS as well as the eddy pair around Montauk
597 Point.

598 The essential role of tidal rectification is to nonlinearly transfer vorticity
599 from the periodic tidal currents to the mean residual circulation. We examined
600 the sources of tidal residual vorticity based on process-oriented experiments and
601 the calculation of the depth-averaged vorticity balance and found that over BIS,
602 slope torque and speed torque are important for the generation of the headland
603 eddies, while over RIS, planetary vorticity stretching is the most important
604 process for generating the westward residual circulation.

605 The formation of tidal eddies around Montauk Point [has](#) long been ascribed
606 to ‘wall’ friction or frictional torque by the topography shoaling toward land
607 (Pingree, 1978; Zimmerman, 1981; Robinson, 1981) or to nonlinear advection
608 (Pingree, 1978; Park and Wang, 2000). Our analysis suggests the importance of
609 nonlinear vorticity advection and frictional torques in BIS. On the other hand,
610 the vorticity balance shows the importance of velocity torque by contributing
611 to the opposite gyres in the same sense, while the effect by planetary vorticity
612 stretching is negligible over BIS. The major role of the Coriolis force is to trap or
613 turn more waters in BIS, instead of transporting those waters into RIS through
614 the gap between BI and Point Judith.

615 The impact of bottom thermal fronts in RIS is tested with two experiments:
616 one driven by climatological mean surface heating in August and the other by
617 artificially cutting the short wave radiation in half. Comparisons between the

618 surface circulations and cross-sectional structures of currents and temperature
619 prove the importance of deep thermal fronts for the summer intensification of
620 cyclonic circulation in RIS and the southwestward current along the southern
621 shore of Long Island. Meanwhile, the experiments show that weaker surface
622 heating generates a wider cyclonic jet because of the smaller contribution from
623 a weaker surface thermal front.

624 To adjudge the contributions to the cyclonic circulation from the barotropic
625 part associated with across-shelf sea level variation and the baroclinic part as-
626 sociated with thermal fronts, we compared the pressure gradients due to steric
627 height and non-steric height respectively and found the dominant roles played
628 by the thermal fronts when surface heating is strong. This method can be
629 widely used over other similar regions with circulation contributions from both
630 barotropic and baroclinic components.

631 Because the pool of cold bottom waters offshore of the bottom thermohaline
632 fronts is often static and organic rich, whereas the zones inshore of the fronts are
633 more turbid and rich with phytoplankton, bottom thermal fronts have biolog-
634 ical and environmental implications by providing natural boundaries between
635 different biological communities. Therefore, the impact of the construction and
636 destruction of deep thermohaline gradients on the benthic habitat heterogeneity
637 and primary production is important. According to the observations by Malek
638 et al. (2010) and Nixon (2010), the demersal fish community in RIS was abun-
639 dant and the phytoplankton density decreased in summer. However, little is
640 known about the exact relationship between the physical structures and biolog-
641 ical activities in RIS or elsewhere. This study may provide useful background
642 information for biological oceanographers to study the relationship between the
643 distribution of biological communities and the construction of the deep thermo-
644 haline fronts.

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755 **Table Captions**

756 **Table 1:** List of Process-oriented Seasonal-variability and Tidal Experiments
757 with ROMS

758 **Table 2:** Tidally Averaged Transport (Unit: m^3/s) along the Cruise Lines
759 Shown in Figure 1. Positive: Transport Entering the Survey Region;
760 Negative: Transport Exiting the Region.

Table 1: List of Process-oriented Seasonal-variability and Tidal Experiments with ROMS

Seas-var Run	Tides ^a	River (m^3/s)	Atm. Forcing ^b	Remote Effects
REAL	Yes	Realistic	Realistic	Yes
NoWind	Yes	Realistic	No Wind	Yes
NoRemote	Yes	Realistic	Realistic	No
HeatStr	Yes	Constant	August Mean	No
HeatWk	Yes	Constant	Halve Swrad ^c	No
Tidal Run ^d	Tides Used	Bathymetry	Earth Rotation	Nolinear Advection
RealM2	M_2	Realistic	Yes	Yes
Flat	M_2	Flat	Yes	Yes
NoRotate	M_2	Realistic	No	Yes
NoAdv	M_2	Realistic	Yes	No

a. Five major tides are used, including M_2 , N_2 , S_2 , O_1 , K_1 .

b. Atmospheric forcing includes air temperature, air pressure, relative humidity, short wave, and long wave radiations.

c. Swrad is abbreviated form of short wave radiation. Compared with HeatStr, In HeatWk, we halved the short wave radiation.

d. This set of experiments does not include atmospheric forcing and river discharge.

Table 2: Tidally Averaged Transport (Unit: m^3/s) along the Cruise Lines Shown in Figure 1. Positive: Transport Entering the Survey Region; Negative: Transport Exiting the Region.

Transect Lines	Seasons	Observations	Model Results
Brenton Reef	Summer	5300 +/-300	4334
	Winter	-200 +/-250	-220
East Passage	Summer	1100 +/-400	776
	Winter	1000 +/-500	550
West Passage	Summer	100 +/-150	214
	Winter	700 +/-100	167
Narragansett Beach	Summer	-5400 +/-600	-3123
	Winter	-2400 +/-300	-879
South Line	Summer	-1000 +/-200	-1936
	Winter	300 +/-250	332

761 **Figure Captions**

762 **Figure 1:** Bathymetry around RIS and BIS. Place names: Long Island Sound
763 (LIS), Montauk Point (MP), Block island Sound (BIS), Block Island (BI),
764 Rhode Island Sound (RIS), Narragansett Bay (NB), Buzzards Bay (BB),
765 Vineyard Sound (VS), Martha’s Vineyard (MV). The cruise lines at the
766 mouth of Narragansett Bay are represented by black lines: (1) Brenton
767 Reef, (2) East Passage, (3) South Line, (4) West Passage, (5) Narragansett
768 Beach. The red stars are the outfalls of, from left to right, the Connecticut
769 River, Pawcatuck River, Pawtuxet River, Blackstone River and Taunton
770 River.

771 **Figure 2:** Model configurations. The local-scale ROMS grid (plot every 6 grid
772 points) has an uniform horizontal resolution of 800 m, and the regional-
773 scale ROMS grid (plot every 6 grid points) is uniform with a resolution of
774 5 Km.

775 **Figure 3:** Comparisons of tidal sea level variations between model results and
776 observations at the sites of a) PO-S and b) PO-F (Ullman and Codiga,
777 2010).

778 **Figure 4:** Comparison of the seasonal mean, depth-averaged velocity fields in
779 summer and winter between the Model (blue arrows) and observations
780 (red arrows). The field data are collected in 2009 and 2010 (Ullman and
781 Codiga, 2010).

782 **Figure 5:** M_2 tidal ellipses over BIS and RIS. Blue and green ellipses are for
783 model with blue representing clockwise, and green counterclockwise. Red
784 ellipses are observations from the Coastal Ocean Dynamics Applications
785 Radar (CODAR; (Ullman and Codiga, 2004)).

786 **Figure 6:** Relative vorticity in the case M2 implemented with realistic bathymetry,
787 Coriolis force and quadratic bottom friction formulation with bottom drag
788 coefficient of $C_d = 7.5 \times 10^{-3}$.

789 **Figure 7:** Depth-averaged residual currents for the process-oriented tidal ex-
790 periments of a) 5 major Tides b) M2; c) Flat; d) NoRotate e) NoAdv.

791 **Figure 8:** Tidal mixing parameter $\log_{10}(H/U^3)$ for M_2 tidal currents. The
792 black line is for the section referenced in Figs. 9, 11, 12, and 14

793 **Figure 9:** Surface currents and cross-sectional structure of along shelf veloc-
794 ity (in cm/s; color), temperature (in $^{\circ}C$; black contours) and salinity (in
795 PSU; white contours) in summer (averaged over July, August and Septem-
796 ber) and winter (averaged over January, February and March) for the ex-
797 periment [REAL](#), which is driven by monthly climatological atmospheric
798 forcing, rivers, tides and realistic open boundary conditions from regional
799 ROMS (Luo et al., 2013). The section is marked as black line in Fig. 8.

800 **Figure 10:** Seasonally mean surface currents for the experiments without wind
801 effects (NoWind) and without remote effects (NoRemote) in summer and
802 winter. a) Summer currents for NoWind; b) Winter currents for NoWind;
803 c) Summer currents for NoRemote and d) Winter currents for NoRemote.

804 **Figure 11:** Time averaged surface circulation and cross-sectional structures of
805 temperature (in $^{\circ}C$; black contours) and velocity (in cm/s; color) from
806 experiments (top) HeatWk and (bottom) HeatStr. The section is marked
807 as black line in Fig. 8.

808 **Figure 12:** Differences of a) surface circulation and b) cross-sectional struc-
809 tures of temperature (in $^{\circ}C$; black contours) and velocity (in cm/s; color)
810 between the [experiments](#) HeatStr and HeatWk (HeatStr-HeatWk). The
811 section is marked as black line in Fig. 8.

812 **Figure 13:** A schematic of the theoretical circulation change from the case
813 HeatStr (red) to the case HeatWk (green) made by the (left) bottom
814 thermal front and (right) surface thermal front respectively.

815 **Figure 14:** Geostrophic velocities by thermal wind equation (black lines) and
816 direct model output (colors) across the section athwart RIS in July for the

817 case REAL. The section is marked as black line in Fig. 8.

818 **Figure 15:** Depth-averaged vorticity balance. The relative vorticity is gener-
819 ated by relative vorticity stretching (term c), planetary vorticity stretching
820 (term d), frictional torques associated with bottom slopes perpendicular
821 to tidal currents (term e) and velocity shear (term f), transported by ad-
822 vection (term b) and dissipated by bottom friction and horizontal diffusion
823 (terms $g + i$).

824 **Figure 16:** Depth-averaged residual currents for the configuration with uni-
825 form horizontal resolution of 200 m.

Figure 1: Bathymetry around RIS and BIS. Place names: Long Island Sound (LIS), Montauk Point (MP), Block island Sound (BIS), Block Island (BI), Rhode Island Sound (RIS), Narragansett Bay (NB), Buzzards Bay (BB), Vineyard Sound (VS), Martha's Vineyard (MV). The cruise lines at the mouth of Narragansett Bay are represented by black lines: (1) Brenton Reef, (2) East Passage, (3) South Line, (4) West Passage, (5) Narragansett Beach. The red stars are the outfalls of, from left to right, the Connecticut River, Pawcatuck River, Pawtuxet River, Blackstone River and Taunton River.

Figure 2: Model configurations. The local-scale ROMS grid (plot every 6 grid points) has an uniform horizontal resolution of 800 m, and the regional-scale ROMS grid (plot every 6 grid points) is uniform with a resolution of 5 Km.

Figure 3: Comparisons of tidal sea level variations between model results and observations at the sites of a) PO-S and b) PO-F (Ullman and Codiga, 2010).

Figure 4: Comparison of the seasonal mean, depth-averaged velocity fields in summer and winter between the Model (blue arrows) and observations (red arrows). The field data are collected in 2009 and 2010 (Ullman and Codiga, 2010)

Figure 5: M_2 tidal ellipses over BIS and RIS. Blue and green ellipses are for model with blue representing clockwise, and green counterclockwise. Red ellipses are observations from the Coastal Ocean Dynamics Applications Radar (CODAR; Ullman and Codiga (2004)).

Figure 6: Relative vorticity in the case M2 implemented with realistic bathymetry, Coriolis force and quadratic bottom friction formulation with bottom drag coefficient of $C_d = 7.5 \times 10^{-3}$.

Figure 7: Depth-averaged residual currents for the process-oriented tidal experiments of a) 5 major Tides b) M2; c) Flat; d) NoRotate e) NoAdv.

Figure 8: Tidal mixing parameter $\log_{10}(H/U^3)$ for M_2 tidal currents. The black line is for the section referenced in Figs. 9, 11, 12, and 14

Figure 9: Surface currents and cross-sectional structure of along shelf velocity (in cm/s; color), temperature (in $^{\circ}C$; black contours) and salinity (in PSU; white contours) in summer (averaged over July, August and September) and winter (averaged over January, February and March) for the experiment **REAL, which is** driven by monthly climatological atmospheric forcing, rivers, tides and realistic open boundary conditions from regional ROMS (Luo et al., 2013). The section is marked as black line in Fig. 8.

Figure 10: Seasonally mean surface currents for the experiments without wind effects (NoWind) and without remote effects (NoRemote) in summer and winter. a) Summer currents for NoWind; b) Winter currents for NoWind; c) Summer currents for NoRemote and d) Winter currents for NoRemote.

Figure 11: Time averaged surface circulation and cross-sectional structures of temperature (in $^{\circ}$ C; black contours) and velocity (in cm/s; color) from experiments (top) HeatWk and (bottom) HeatStr. The section is marked as black line in Fig. 8.

Figure 12: Differences of a) surface circulation and b) cross-sectional structures of temperature (in $^{\circ}$ C; black contours) and velocity (in cm/s; color) between the [experiments](#) HeatStr and HeatWk (HeatStr-HeatWk). The section is marked as black line in Fig. 8.

Figure 13: A schematic of the theoretical circulation change from the case HeatStr (red) to the case HeatWk (green) made by the (left) bottom thermal front and (right) surface thermal front respectively.

Figure 14: Geostrophic velocities by thermal wind equation (black lines) and direct model output (colors) across the section athwart RIS in July for the case REAL. The section is marked as black line in Fig. 8.

Figure 15: Depth-averaged vorticity balance. The relative vorticity is generated by relative vorticity stretching (term *c*), planetary vorticity stretching (term *d*), frictional torques associated with bottom slopes perpendicular to tidal currents (term *e*) and velocity shear (term *f*), transported by advection (term *b*) and dissipated by bottom friction and horizontal diffusion (terms *g* + *i*).

Figure 16: Depth-averaged residual currents for the configuration with uniform horizontal resolution of 200 m.