Modulation of Near-Inertial Oscillations by Low-Frequency Current Variations on the Inner Scotian Shelf

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ABSTRACT
Near-inertial oscillations (NIOs) on the inner Scotian shelf are studied using observations, a simple slab model, and two operational shelf circulation models. High-frequency radar and ADCP observations from December 2015 to February 2016 show that individual NIO events forced by time-varying wind stress typically lasted for three to four inertial periods. NIOs with speeds exceeding 0.25 m s\(^{-1}\) were observed in the offshore part of the study region, but their amplitudes decreased shoreward within \(40\) km of the coast. The NIOs had spatial scales of \(80\) and \(40\) km in the alongshore and cross-shore directions, respectively. The NIO phases varied moving from west to east, consistent with the typical movement of winter storms across the study region. Evolving rotary spectral analysis reveals that the peak frequency \(f_p\) of the NIOs varied with time by \(\pm7\%\) of the local inertial frequency. The variation in \(f_p\) can be explained in part by local wind forcing as demonstrated by the slab model. The remaining variation in \(f_p\) can be explained in part by variations in the background vorticity associated with changes in the strength and position of the Nova Scotia Current, an unstable baroclinic boundary current that runs along the coast to the southwest. Two operational shelf circulation models are used to examine the abovementioned features in the high-frequency-radar and ADCP observations. The models reproduce the spatial structure of the NIOs and, in a qualitative sense, the temporal variations of \(f_p\).

1. Introduction
Near-inertial oscillations (NIOs) are ubiquitous throughout the ocean. They are associated with strong vertical shear and contribute to upper-ocean mixing (e.g., Zhai et al. 2009; Jochum et al. 2013). The downward flux of near-inertial energy associated with NIOs is also thought to contribute to diapycnal mixing in the deep ocean (e.g., Alford et al. 2016). The classical mechanism for the generation of wind-driven NIOs includes two stages. During the first (short) stage, a storm passing overhead generates currents in the surface ocean mixed layer (ML). During the second (longer) stage, these currents undergo “Rossby adjustment” and NIOs are generated. Gill (1984) used vertical modes of a flat bottom ocean to study the near-inertial waves driven by NIOs in the wake of a storm. In this case, the near-inertial wave frequency is given by \(\omega_n^2 = f^2 + c_n^2 f\), where \(f\) is the planetary vorticity, \(c_n\) is the eigenvalue for the \(n\)th mode, and \(l\) is the horizontal wavenumber. Because the horizontal scale \(l^{-1}\) is limited by the wind and the proximity of the coast, their frequency is slightly higher than \(f\). At the base of the ML, near-inertial internal waves are generated through the horizontal convergence/divergence of the ML (“inertial pumping”). Zervakis and Levine (1995) noted that near-inertial energy can propagate downward into deeper layers as low modes leave the generation area. The resulting frequency changes at a fixed point are a complicated combination of contributions from each mode.

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showed that the effective inertial frequency is \( \sim 1.005f \) and that the frequency increases with depth. On a \( \beta \) plane, the near-inertial waves with \( \omega > f \) can propagate poleward until they reach the turning latitude (Anderson and Gill 1979; Gill 1984). Near-inertial waves generated near their turning latitudes propagate equatorward and vertically downward until they reach the seafloor with their frequencies exceeding the local \( f \) (on the order of 10% of \( f \) at midlatitudes; Garrett 2001). At the seafloor, these waves can be further reflected equatorward until their frequencies become \( 2f \) and they decay by parametric subharmonic instability (Nagasawa et al. 2000).

In addition to the \( \beta \) effect, low-frequency flows can influence the generation and propagation of NIOs. In a seminal paper, Kunze (1985; see also Mooers 1975) showed that the effective inertial frequency \( f_e \) of NIOs can be modified by a low-frequency flow as follows:

\[
 f_e \approx f + \zeta/2,
\]

(1)

where \( \zeta \) is the relative vorticity. The frequency shift of \( \zeta/2 \) has been observed to influence the generation of NIOs in regions where the diurnal wind forcing period matches \( f_e \) (Mihanović et al. 2016). Once generated, trapping and amplification of propagating NIOs can occur in regions of negative vorticity (Kunze 1985). This has been confirmed by observations of elevated near-inertial energy in anticyclonic eddies (Elipot et al. 2010). These anticyclonic eddies can transfer the near-inertial energy to the deep ocean through the “inertial chimney” effect (e.g., Zhai et al. 2005, 2007). Horizontal advection of NIOs by large-scale geostrophic flows, such as the Gulf Stream, can also be important in redistributing the near-inertial energy (Zhai et al. 2004), and nonlinear interactions between NIOs and low-frequency flows can lead to an exchange of energy (e.g., Müller 1976). For example, in the Kuroshio Extension region, the efficiency of energy exchange in anticyclonic eddies is about twice that of cyclonic eddies as a result of the frequency shift by the relative vorticity (Jing et al. 2017).

Equation (1) has been validated in the tropical Pacific Ocean (Poulain et al. 1992) and the global ocean (Elipot et al. 2010) based on analyses of observed surface drifter tracks. Both of these open ocean studies focused on frequency modulation of the inertial frequency in regions with energetic mesoscale eddies. Equation (1) has also been validated closer to coastal boundaries. For example, Mihanović et al. (2016) showed that changes in the Leeuwin Current off southwestern Australia can modulate \( f_e \) by more than 50%. Shearman (2005) found that the strong relative vorticity associated with the shelf-break front off New England can cause a significant reduction in \( f_e \). Bondur et al. (2013) found that the background currents on the Hawaii shelf could induce strong variations of \( f_e \), leading to the anomalous counterclockwise rotation of NIOs in this region.

This study focuses on the inner half of the Scotian shelf (ScS) off the east coast of Canada. This region features a persistent southwestward coastal jet [known as the Nova Scotia Current (NSC)] with peak surface speeds reaching 0.3 m s\(^{-1}\) centered at approximately 45 km from the coast. The NSC varies on time scales from days to seasons (Petrie 1987; Dever et al. 2016) and occasionally oscillates in the alongshore direction with a mean wavelength of \( \sim 50 \) km (Petrie 1987). Anderson and Smith (1989) observed strong NIOs with maximum speeds of order 0.2 m s\(^{-1}\) on the ScS using moored current meters located on the 150-m isobath. The same authors also found that the peak frequency of NIOs was slightly subinertial and suggested this may be caused by the shear of the mean NSC.

The Marine Environmental Observation Prediction and Response (MEOPAR) research network deployed a high-frequency (HF) radar system off Halifax Harbour in 2015 to monitor surface ocean currents over the inner ScS (Figs. 1, 2). The radial currents observed by the two antennas were processed to give hourly maps of surface currents defined on a horizontal grid with a spacing of 6 km. In this study, the HF-radar data are complemented by observed time-varying vertical profiles of horizontal currents made by an ADCP located close to the center of the mean NSC and within the region monitored by the HF radar. These two new observational datasets provide an excellent opportunity to examine NIOs on the inner ScS and, in particular, how NIOs are modulated by an unstable coastal boundary current.

To help to interpret these new observational datasets, we take advantage of three models. The simplest is a slab model used to quantify the effect of local wind forcing. We also use two prototype operational shelf circulation models, the Dalhousie Coastal Ocean Forecast System (DalCoast; Ohashi et al. 2006) and the Gulf of Maine and Scotian Shelf system (GoMSS; Katavouta et al. 2016). DalCoast is based on the Princeton Ocean Model (POM; Mellor 2004), and GoMSS is based on the Nucleus for European Modelling of the Ocean (NEMO) framework (Molines et al. 2014). Both DalCoast and GoMSS have undergone extensive validation in terms of their ability to simulate tidal and subtidal variations. These models, however, use different numerics and different vertical coordinates. A question that remains to be addressed is how well they can simulate NIOs.

In this study, we first use the new observations made by the HF radar and ADCP to describe the intensity, the intermittency, and, for the first time, the spatial structure of NIOs over the inner ScS. We also attempt to explain the observed changes in the frequency of peak near-inertial energy \( f_p \) using Eq. (1) with the changes in \( \zeta \) estimated from the low-pass-filtered HF-radar observations. We then...
use the three models to help understand how the slowly varying background circulation influences NIOs on the ScS. We also use the observations to assess the performance of the two operational models in the near-inertial band and to identify possible ways of improving the models for practical applications, including the forecasting of currents on the inner shelf.

Section 2 provides a statistical description of the HF-radar and ADCP observations and compares them with results produced by the slab model forced by the local wind. Section 3 describes the two operational models. Section 4 describes the results produced by the two operational models and their comparison with the observations. Section 5 is a summary and discussion.

2. Processing and analysis of observations

a. Processing of the HR-radar and ADCP data

The HF-radar system consists of two long-range Coastal Ocean Dynamics Applications Radar (CODAR) Seasonde sets located at Sandy Cove and Clam Harbour (Fig. 2) off the Atlantic coast of Nova Scotia, Canada. Each radar operates at a central frequency of 4.8 MHz and provides hourly radial surface currents with 6-km resolution and coverage up to 200 km from the coastal radar site. The HF-radar measurements correspond to the ocean currents averaged from the surface to a depth on the order of $l/(4p)$, where $l$ is the Bragg wavelength (Stewart and Joy 1974). This depth corresponds to approximately 2.5 m given the system’s transmit frequency. We will henceforth refer to the currents observed by the HF radar as “surface” currents.

The HF-radar data used in this study were processed by the HF Radar Network (HFRNet; http://cordc.ucsd.edu/projects/mapping/) using the following procedure. The data were first mapped onto regional grids defined using an equidistant cylindrical projection with a grid spacing of 6 km. Surface current vectors were then estimated using a least squares fit to at least three radial velocities within a search radius of 10 km from each grid point. As part of the data quality control, a velocity...
A threshold of 3.0 m s\(^{-1}\) was applied to both the radial and total velocities. One issue associated with this method is that the solution becomes unstable when the radial current components are close to parallel. This occurs near the line connecting the two radar sites and in the far field. Errors associated with the radar geometry are usually quantified by the geometric dilution of precision (GDOP; Chapman et al. 1997), a unitless measure of uncertainty. Low GDOP values correspond to a preferred geometric configuration. The GDOP map for the present system is given in Fig. 2. Grid points with a GDOP greater than 3 are not used in the present study.

Radio interference at either unit can lead to gaps in the observed HF-radar record. The radio interference usually occurs between sunset and sunrise and is often related to changes in the ionosphere. Hourly data at grid points with more than 50% temporal coverage (Fig. 2) for three winter months (December 2015–February 2016) are used in this study (this winter period was chosen because the HF-radar data coverage is relatively high).

During these three winter months, vertical profiles of currents were also measured by a bottom-mounted, upward-looking ADCP at location T2 on the “Halifax Line” (Fig. 1). The ADCP was deployed by the Ocean Tracking Network to monitor the NSC. (Two additional ADCP were also deployed in the same general area, but they were not functional during the study period.) The ADCP currents were averaged over 30-min time windows and 4-m vertical bins. The shallowest bin was centered at 20 m from the sea surface and the deepest bin was 10 m off the bottom.

### b. The slab model

Slab models have been widely used to simulate wind-driven inertial oscillations of surface current (e.g., Pollard and Millard 1970; D’Asaro 1985; Paduan et al. 1989). The equation for the ML velocity \( \mathbf{u} = u + i v \) is assumed to be

\[
\frac{\partial \mathbf{u}}{\partial t} + i \mathbf{u} + \lambda \mathbf{u} = \frac{\tau}{\rho h},
\]

where \( f \) is the inertial frequency, \( i = (-1)^{1/2} \), \( \lambda \) is a linear damping coefficient, \( \tau = \tau_s + ir_c \) is the wind stress, \( \rho \) is water density, and \( h \) is the mixed layer depth (MLD).

As discussed in previous studies (e.g., D’Asaro 1985; Park et al. 2009), the damping parameter \( \lambda \) represents the decay of NIOs in the ML. Possible mechanisms responsible for this decay include energy radiation into the interior of the ocean, local turbulent dissipation, and nonlinear transfer to other frequencies. Previous studies have pointed out that, in deep water, NIO decay time scales resulting from downward energy propagation are comparable to those based on observations (e.g., Balmforth and Young 1999). In shallow water, however,
the ML can extend to the seafloor, and thus bottom friction can play an active role in the decay of NIOs. With the assumption that \( u = 0 \) at \( t = 0 \), the velocity at subsequent times can be written explicitly in terms of the history of the wind stress:

\[
\mathbf{u}(t) = \frac{1}{\rho h} \int_0^t e^{-(\theta + \lambda)k(t-t')} \mathbf{\tau}(t') \, dt'.
\]  

The \( e \)-folding time of an inertial oscillation forced by a wind impulse is \( \lambda^{-1} \). If the wind stress can be modeled as a stationary process with a flat spectrum over the near-inertial band, then the rotary spectrum of the current will peak at \( \omega = f \). The half-width of the spectral peak, determined in terms of half-power points, is \( \lambda \), thereby providing an alternative interpretation of \( \lambda \). D’Asaro (1985) suggested that typical values of \( \lambda^{-1} \) lie in the range of 2–10 days in deep water. In shallow water with the ML extending to the bottom, Lagerloef and Muench (1987) showed that \( \lambda \) is equivalent to \( \lambda'/h \), where \( \lambda' \) is the linear bottom drag coefficient. Applying typical values of \( \lambda' = 0.1 \text{ cm s}^{-1} \) and \( h = 50 \text{ m} \), the estimated \( \lambda^{-1} \) is \( \sim 14 \text{ h} \) in shallow water. In our study, we set \( \lambda^{-1} \) to three inertial periods at location T2. This estimate was based on analysis of the observed current time series in both the time and frequency domains. Specifically we visually examined the observed time series at location T2 to estimate the decay time scale of significant NIOs following their generation (and hence \( \lambda^{-1} \)). We also estimated \( \lambda \) from the half-width of the near-inertial spectral peak of the observed currents.

The wind forcing for the slab model was calculated from hourly surface winds extracted from the Climate Forecast System, version 2 (CFSv2; Saha et al. 2014). Wind stress was calculated using a quadratic formula with the drag coefficient given by the bulk formula of Large and Pond (1981) and Powell et al. (2003).

c. **Observed monthly means**

The monthly mean surface currents calculated from the HF-radar observations indicate a persistent southwest-alongshore jet known as the NSC for all three months (Figs. 3a–c). The observed NSC generally occurred between the 100- and 200-m isobaths and had a width of about 30 km. The NSC weakened by about 0.07 m s\(^{-1}\) over the study period and its center migrated offshore by about 11 km between January and February (Table 1). The current maps for December and January show small-scale circulation features offshore of the NSC but they are weak and could result from the relatively low HF-radar data availability in this region (Fig. 2).

Dever et al. (2016) analyzed current observations for all three ADCPs deployed by the Ocean Tracking Network and estimated geostrophic currents from glider observations of temperature and salinity for the period 2011–14. They found that location T2 was close to the center of the mean NSC in winter, consistent with the observed means for December 2015 and January 2016 shown in Figs. 3a and 3b, respectively.

Figure 4 shows observations of monthly mean currents normal to the Halifax Line (alongshore currents) made by the HF radar and the ADCP at T2. The HF-radar observations along this section confirm the weakening and offshore migration of the NSC but also suggest that the NSC widened in February. The vertical current profiles from the ADCP show the NSC was strongest for the shallowest bin centered on 20 m and decreased toward the bottom, consistent with the finding made by Dever et al. (2016). The mean currents observed by the ADCP at 20 m were stronger than the shallower surface currents observed by the HF radar (Table 1). This difference can be explained by the mean effect of the wind on the near-surface flow.

d. **Observed near-inertial oscillations at location T2**

Time series of wind stress and observed surface currents at T2 are presented in Fig. 5. The intense NIOs in the current time series coincided with strong winds associated with the passage of winter storms (e.g., 5 and 16 December). The NIOs typically lasted for three to four inertial periods. Not all winter storms generated strong NIOs (e.g., storms on 12–16 January and 9 February). To explain this difference, the simple slab model was used to simulate the NIOs at T2 based on the local wind. The simulated currents by the slab model are dominated by NIOs. They are in reasonable agreement with the observed NIOs (Figs. 5b,c and Figs. 5e,f), confirming the important influence of the local wind in generating NIOs in this region. The results produced by the slab model also explain why some wind events were more effective than others in generating NIOs. For example, from 12 to 16 January the NIOs were partially suppressed by the counterclockwise (CCW)-rotating wind stress around 13 and 16 January. Similarly, the CCW-rotating wind stress around 9 February was rotationally unfavorable for the generation of the clockwise (CW)-rotating NIOs.

Rotary spectra (Gonella 1972) of the current time series observed by the HF radar and the ADCP (Fig. 6a) confirm the important contribution of NIOs. [In the present study, positive (negative) frequencies denote CW-rotating (CCW rotating) motions.] Rotary spectra of the HF-radar and ADCP observations are very similar in the near-inertial band for both CW and CCW components, even though they were made at depths separated by 17.5 m. It is speculated that the observed
CCW component is caused by bottom reflection (e.g., Eriksen 1982; Jordi and Wang 2008). Here, the CCW spectral peak is about one order of magnitude smaller than the CW peak. However, over steep topography, the CCW component can become relatively large. For example, in a study of NIOs around a submarine canyon, Jordi and Wang (2008) found that the CCW component resulting from bottom reflection can be as large as the...
showed that the background flow can modify the near-inertial band calculated using a sliding window of length 10 days with a 1-h overlap. (The HF-radar time series (Fig. 5), which show good agreement in the rotary coherence between the HF-radar and ADCP observations. (The cutoff periods of the band-pass filter are 15 and 20 h.) About 71% of the total variance can be explained by the first mode. The spatial structure of this mode (Fig. 7a) was large scale with lower amplitudes within ~40 km of the coast. The elements of the first mode rotated cyclonically, moving from west to east. The mode amplitude (Fig. 7e) is dominated by clockwise oscillations centered at the inertial frequency with increased amplitudes during storms such as on 5 and 16 December and 18 February. The higher-order modes are not presented because each of them individually accounts for less than 7% of the total variance of the HF-radar data.

To explain the phase differences of NIOs moving from west to east (Fig. 7a), we forced the slab model with reanalysis wind stress to simulate the local current response at each HF-radar grid point. The results of an EOF analysis of the slab-model simulations are presented in Fig. 7b. It is noted that the complex elements of the first EOF rotate cyclonically, moving from west to east, in general agreement with observations (Fig. 7a). This implies that the phase differences moving from west to east are due primarily to the atmospheric forcing. This is consistent with the fact that storms generally move from west to east in the study region during winter.

To examine the spatial structure of the NIOs in more detail, we spectrally analyzed each gridpoint time series of observed HF-radar current and mapped the results across the grid. The amplitudes of the NIOs, defined by the square root of the integral of the rotary spectrum between 0.85 and 1.06 cycles per day (cpd) (Fig. 6b). This high coherence is consistent with visual inspection of the two observed current time series (Fig. 5), which show good agreement in amplitude and timing during the strong storm-induced NIOs of 5 and 16 December and 18 February. During moderate wind events, when the Ekman depth was relatively shallow and vertical shear near the surface was large, discrepancies between the two time series are evident.

As mentioned in the introduction, Kunze (1985) showed that the background flow can modify \( f_e \) through changes in \( \zeta \) according to Eq. (1). To estimate the magnitude of this effect, we present in Fig. 6c the evolving rotary spectrum (ERS) of the ADCP time series in the near-inertial band calculated using a sliding window of length 10 days with a 1-h overlap. (The HF-radar time series at T2 has too many missing observations to undertake this type of analysis.) The choice of a 10-day window is a compromise between the need for a long window to increase spectral resolution and degrees of freedom, and a short window to capture the intermittency of the NIOs. It is clear from Fig. 6c that \( f_e \) changed with time. For example, the NIO associated with the 16 December storm had \( f_p \approx 1.06 f \) in contrast to \( f_p \approx 0.93 f \) around 18 February. We also show \( f_e \) calculated from Eq. (1) using \( \zeta \) at T2 estimated from the HF-radar data averaged in time using the same 10-day sliding window. The similarity of the variability of \( f_p \) and \( f_e \) during the NIOs in December and February is consistent with the relative vorticity correction given by Eq. (1) and thus the idea that changes in the position and intensity of the NSC influence NIOs on the inner ScS. To illustrate, the ~6% shift of \( f_p \) in mid-December coincided with an inward excursion of the NSC and the ~7% shift in late February coincided with an offshore excursion that changed the sign of \( \zeta \) at T2 (see Figs. 3 and 4). It is noted that the agreement between variations in \( f_p \) and \( f_e \) is not perfect. The discrepancy between \( f_p \) and \( f_e \) during the NIOs in January can be explained in part by local wind forcing as demonstrated by the slab model (Fig. 6d). Overall, however, adding \( \zeta \) to the ERS of the slab model (Fig. 6c) leads to closer agreement between the ERS of the slab model and the ADCP observations.

Table 1. Monthly means of observed and simulated currents normal to the Halifax Line. The 2.5-m values are the peak observed or modeled mean current (m s\(^{-1}\)) and the numbers in parentheses are the distance from shore (km) at which it occurs. The ADCP values are the mean current (m s\(^{-1}\)) observed by the ADCP at location T2, which is 42 km from shore. The corresponding 20-m values for DalCoast and GoMSS are the modeled values at T2.

<table>
<thead>
<tr>
<th></th>
<th>Dec 2015</th>
<th>Jan 2016</th>
<th>Feb 2016</th>
</tr>
</thead>
<tbody>
<tr>
<td>HF radar</td>
<td>0.29 (42)</td>
<td>0.26 (42)</td>
<td>0.22 (53)</td>
</tr>
<tr>
<td>ADCP</td>
<td>0.32</td>
<td>0.34</td>
<td>0.30</td>
</tr>
<tr>
<td>DalCoast</td>
<td>0.27 (45)</td>
<td>0.25 (45)</td>
<td>0.20 (49)</td>
</tr>
<tr>
<td>GoMSS</td>
<td>0.26 (38)</td>
<td>0.25 (38)</td>
<td>0.21 (34)</td>
</tr>
<tr>
<td>GoMSS</td>
<td>0.26</td>
<td>0.25</td>
<td>0.20</td>
</tr>
</tbody>
</table>
The inhibition is explained in terms of the leakage of near-inertial energy both downward and offshore. Kundu et al. (1983, their Fig. 6) showed that NIOs undergo an exponential decay over 5 days within one Rossby radius of the coast as a result of leakage of energy both horizontally and vertically. The horizontal leakage from the coastal region leads to an offshore increase in NIO energy over the same time period.

The internal Rossby radius for the study area is \( \sim 15 \) km (Dever 2017). In agreement with Kundu et al. (1983), we found a slight initial increase in NIO amplitude during well-defined NIO events at a distance of approximately 1 Rossby radius from shore (not shown). However, we found that the coastal suppression of NIO amplitude occurred over a scale of \( \sim 40 \) km, well beyond 1 Rossby radius.

To explain this discrepancy with the theory of Kundu et al. (1983), we plotted the distribution of near-inertial kinetic energy over the whole model domain (not shown) as predicted by the numerical models described
in section 3. We found the simulated NIOs depend strongly on water depth even in regions well away from the coast. For example, in the Gulf of Maine, where the offshore gradient of water depth near the coast is relatively small, the coastal suppression of NIOs occurs over larger offshore distances than the ScS. [Note that the coastal suppression of NIOs in the Gulf of Maine is consistent with the absence of the near-inertial signals in the observed wave height time series at nearshore buoys (Wang and Sheng 2018).] Away from the coast, we found that NIOs are effectively suppressed over shallow regions such as Western Bank and Georges Bank (Fig. 1). Kundu and Thomson [1985, their Eq. (22)] showed that the inertial current amplitude under a fast-moving storm is proportional to \( \tau_0(1 - h/D)/(fh) \), where \( \tau_0 \) is the wind stress magnitude and \( D \) is the water depth. This implies that the amplitude increases with \( D \) and decreases with \( h \). MLD in shallow water is usually increased because of tidal and wind mixing. Furthermore, the ML in shallow water can extend to the seafloor, where bottom friction becomes important. We therefore conclude that shallow water depth, increased MLD, and bottom friction contribute to the observed suppression of NIOs within ~40 km of the coast in the study area.

The spatial scale of the NIOs was estimated from the rotary coherence between the HF-radar observations at location T2 and all other grid points. The coherence was evaluated at \( f_p \). The coherence magnitude is anisotropic.
it exceeds 0.5 within ~40 km of T2 in the alongshore direction, and ~20 km in the cross-shore direction. The phase of the coherence (Fig. 8g) increases from west to east, consistent with the EOF analysis.

Figure 9a shows the horizontal distribution of the relative frequency shift ($\Delta f = f_p - f$) for the HF-radar observations. The observed $f_p$ is below $f$ within ~50 km of the coast and slightly above $f$ seaward of this distance. To explain the spatial variation in $f_p$, $\zeta/2$ was estimated from current fields observed by the HF radar. Figure 9d shows the ERS of slab-model simulations. As in (d) except that the ERS of the slab-model simulations has been shifted at each frequency by $\zeta/2$. All ERS were calculated using a 10-day sliding window and a spectral window of width 0.379 cpd. The frequency axis is limited to the near-inertial band. In (c)–(e), the black dashed lines represent the local inertial frequency ($\omega = f$) and the blue lines show $f_p$ (plotted only for frequencies with relatively high near-inertial energy).

To further examine the relevance of the present study to other locations, we also analyzed the HF-radar observations made off the Oregon coast (see appendix A for details). Overall, the coastal suppression of NIOs and the modulation of $f_p$ by the background vorticity off the Oregon coast are consistent with our findings for the ScS.

3. Description of the two ocean models

a. DalCoast

DalCoast is a 3D, sigma-coordinate, primitive equation ocean circulation model based on the POM (Mellor 2004).
The model domain covers the Gulf of St. Lawrence, the ScS, and the Gulf of Maine and adjacent deep waters (Fig. 1). The grid spacing is 1/16° (~7 km) in both the longitudinal and latitudinal directions. There are 40 sigma levels in the vertical with the highest concentration near the surface and bottom and approximately equal spacing in the interior. The model topography is based on the General Bathymetric Chart of the Oceans (GEBCO) bathymetric dataset (http://www.gebco.net/).

Hourly surface wind and atmospheric pressure at sea level were extracted from the CFSv2 (Saha et al. 2014) to drive the model. Wind stress was calculated as described in section 2b. In addition to forcing by wind stress and atmospheric pressure, DalCoast is also driven by the net heat and freshwater fluxes at the sea surface and freshwater runoff from major rivers in the region. At the model’s open boundaries, the model is driven by 1) wind-induced hourly sea level and depth-averaged currents produced by a barotropic model covering the northwest Atlantic Ocean (72°–42°W, 38°–60°N) with a resolution of 1/12°, 2) tidal forcing specified in terms of hourly sea levels and depth-averaged currents predicted by the Oregon State University (OSU) Tidal Inversion Software (OTIS) for eight tidal constituents (M2, S2, N2, K2, K1, O1, P1, and Q1), and 3) daily values of temperature, salinity, and large-scale density-driven currents provided by an ocean-ice numerical model of the northwest Atlantic Ocean (Urrego-Blanco and Sheng 2012). Note that these three contributors to the open boundary condition have not been adjusted to match each other. Spectral nudging (Thompson et al. 2007; Wright et al. 2006) and the semiprognostic method (Sheng et al. 2001) are used to reduce bias in temperature and salinity in DalCoast and the model of Urrego-Blanco and Sheng (2012). Both models are nudged to the mean and seasonal cycle of the monthly climatology of Geshelin et al. (1999).

b. GoMSS

GoMSS is a 3D, z-coordinate, primitive equation ocean circulation model based on NEMO (Molines et al. 2014).
Its domain covers the ScS, the Gulf of Maine, and adjacent deep waters (Fig. 1). The grid spacing is 1/36° (~2.8 km) in both longitudinal and latitudinal directions. There are 50 z levels with a spacing that varies from 1 m near the surface to 458 m at the deepest level (5500 m). Partial cells are used to better represent the bathymetry. The use of the “variable volume level” approach (Levier et al. 2007) allows the thickness of the vertical levels to vary with changes in the sea surface elevation. The model bathymetry is based primarily on the 2-arc-min gridded global relief dataset ETOPO2v2 (NOAA National Geophysical Data Center, now part of the NOAA National Geophysical Data Center).
National Centers for Environmental Information). Higher-resolution data provided by R. Karsten (Acadia University, 2014, personal communication) were used to improve the bathymetry in the inner Gulf of Maine.

Momentum and heat fluxes at the ocean surface are calculated using the same hourly atmospheric variables used to force DalCoast. The initial and open boundary conditions (excluding tides) are interpolated from daily temperature, salinity, sea surface height, and horizontal velocity fields from the Mercator global ocean forecast system (Prototype System, version 4; nominal horizontal grid spacing of 1/12°). Eight tidal constituents (M2, S2, N2, K2, K1, O1, P1, and Q1) are also used to specify tidal forcing at the model’s lateral open boundaries. The tidal elevations and transports were obtained from the finite-element-solution (FES2004) global tidal model of Lyard et al. (2006).

**c. Validation**

Both DalCoast and GoMSS have been validated extensively in the tidal, synoptic, and seasonal frequency bands [for DalCoast see Thompson et al. (2007), Ohashi and Sheng (2013, 2016), Ohashi et al. (2009a,b), and Wang and Sheng (2016); for GoMSS see Katavouta et al. (2016) and Katavouta and Thompson (2016)]. Further validation of both models in the tidal and synoptic bands, using observations of coastal sea level and current for the present study period, is described in appendix B. The following section focuses on the near-inertial band.

**4. Comparison of ocean model simulations and observations**

The monthly mean surface currents simulated by DalCoast and GoMSS are first assessed by comparison with the HF-radar and ADCP observations. This is followed by an assessment of the model simulations in the near-inertial band.

**a. Monthly means**

Both DalCoast and GoMSS simulate the well-defined southwestward NSC that is strong (0.25–0.27 m s⁻¹) in December and January and weakens (~0.20 m s⁻¹) in

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**FIG. 9.** Horizontal distributions of (a)–(c) shift in peak frequency relative to inertial frequency (f_p − f) and (d)–(f) 3-month mean of ζ/2 for the (left) HF-radar observations, (center) DalCoast simulations, and (right) GoMSS simulations.
February, consistent with the HF-radar observations (Fig. 3; Table 1). DalCoast reproduces reasonably well the width (~30 km) and the location of the observed NSC for all three months, while the NSC produced by GoMSS is too narrow (~20 km) and also too close to shore for all three months (Table 1). There is some indication in Fig. 3 that GoMSS provides more realistic simulations seaward of the NSC, including some small-scale circulation cells that are evident in the HF-radar observations.

The simulated monthly mean currents normal to the Halifax Line (alongshore currents) produced by DalCoast and GoMSS are presented in Fig. 4 as a function of depth and offshore distance. Both models produce the vertical shear and near-zero bottom velocities evident in the ADCP observations at T2. The velocity sections also confirm that the NSC simulated by GoMSS is too close to shore for all three months. Both models also produce the mean alongshore flows at 20 m that are weaker than the corresponding ADCP observations by almost 0.1 m s\(^{-1}\) (Table 1). It is speculated that this discrepancy is caused mainly by baroclinic processes associated with the movement of oceanic fronts that are not simulated correctly because of inadequate model resolutions and imperfect model physics.

The vertical distributions of monthly mean salinity along the Halifax Line simulated by the two models are also shown by black contour lines in Fig. 4. Both models produce low-salinity waters within 20–80 km of shore. There is good agreement between the salinity simulated by DalCoast and the section published by Dever et al. (2016, see their Fig. 5b) based on glider observations.
made during January–March over the period 2011–14. By comparison, GoMSS produces weaker vertical gradients of salinity in the top 50 m than DalCoast, suggesting a salinity bias error in GoMSS. The reasons for this bias and the incorrect offshore position of the NSC in GoMSS are currently under investigation.

b. Near-inertial oscillations at location T2

Time series of the simulated surface currents by DalCoast and GoMSS are presented in Fig. 10. There is generally good agreement between the simulations produced by the two models and the observations of surface currents made by the HF radar. Well-defined NIOs, clearly related to the wind forcing, are evident in both the observations and simulations.

The rotary spectra of the simulated current time series in the near-inertial band for the CW component are generally similar to those of the observed currents made by the HF radar and ADCP (Figs. 11a,b). Both models, however, did not reproduce the observed weak CCW spectral peak. This could be due to inadequate model resolution. In addition, all spectra have peaks in the vicinity of zero frequency, but DalCoast underestimates the observed low-frequency variability of currents at 20 m, indicating its simulated mean flow is less variable in time at this water depth.

Fig. 11. Rotary spectra of observed and simulated currents at location T2 at depths of (a) 2.5 and (b) 20 m. The observations at 2.5 and 20 m were made by the HF radar and ADCP, respectively. Also shown is the ERS, based on a 10-day sliding window, of the (c) ADCP observations and model simulations by (d) DalCoast and (e) GoMSS, all at a depth of 20 m. The frequency axis is limited to the near-inertial band. The dashed lines represent the local inertial frequency ($\omega = f$), and the blue lines show $f_p$ (plotted only for frequencies with relatively high near-inertial energy). The green lines show $f_e = f + \xi/2$, where $\xi$ was estimated from near-surface currents at 2.5 m observed by the HF radar [in (c)] and simulations made by DalCoast [in (d)] and GoMSS [in (e)] at 20 m.
Closer inspection of the rotary spectra for GoMSS (Figs. 11a,b) reveals that the near-inertial peaks occur at 
\[ f_p' = 1.05f \]. To explain this frequency shift, we present the ERO simulation at \( t = 20 \) min in Figs. 11d and 11e. For DalCoast, \( f_p \) is less variable than observed (Fig. 11c). For GoMSS, \( f_p \) exhibits more temporal variability that is closer to the observed \( f_p \), but there is a difference in the mean (Figs. 11c,e). To explain these results, we estimated the time variation of \( \zeta \) for both models and added the plot of \( f_e = f + \zeta/2 \) to Figs. 11d and 11e. For DalCoast, \( \zeta \) is close to zero over the 3-month period, consistent with the proximity of \( f_p \) to the inertial frequency. For the GoMSS simulations of the NIOs in December and February, the positive background vorticity is associated with a positive shift of \( f_p \), consistent with Eq. (1). The calculated \( f_e \) does not match \( f_p \) exactly. One contributor to the discrepancy is the use of 10-day sliding window to calculate \( f_p \). Given NIOs are highly intermittent, the calculated \( f_p \) at a specific time is only a smoothed approximation. In mid-January, the slab-model simulations show that the negative shift of \( f_p \) is due to the subinertial wind forcing.

c. Spatial structure of the near-inertial oscillations

To further examine the variability of the model simulations in the near-inertial band, an EOF analysis of the bandpass filtered simulations was conducted in the same way as for the HF-radar observations. Overall, DalCoast fits the first mode of the observations more closely than GoMSS. The latter overestimates the intensity of the NIOs in the southwest area of the study region. Both models, particularly DalCoast, account for more of their total variance (DalCoast: 93%; GoMSS: 85%) in the first mode than the observations (71%). This is due in part to missing values in the observed time series made by the HF radar, and the remaining discrepancy can be due to observation errors and real small-scale features that are not simulated correctly by the models because of inadequate model resolution and imperfect model physics.

A comparison of Figs. 8b and 8a demonstrates that the amplitudes of the NIOs simulated by DalCoast are in good agreement with their observed counterparts. By comparison, the NIO amplitudes produced by GoMSS are smaller than the observations close to shore (Figs. 8c,a). It is speculated that this discrepancy is caused by the unrealistically low vertical stratification and deeper ML in GoMSS in this region, in comparison with DalCoast (see Fig. 4).

Furthermore, GoMSS also overestimates the NIO amplitudes in the offshore region. To explain this overestimation, we present in Fig. 12 the bandpass filtered alongshore currents as a function of time and depth for

![Fig. 12. Hovmöller plots of bandpass-filtered alongshore currents for four periods with relatively strong NIOs at station T2 for the (top) ADCP observations and the (middle) DalCoast and (bottom) GoMSS simulations. Each column corresponds to a specific 5-day period defined by the x axis of the bottom panel. The dashed line represents the MLD estimated from the vertical gradient of density.](image)
the ADCP and the models for four periods with relatively strong NIOs. It is noted that the MLD for model results of DalCoast is more constant in time and that the density jump is weaker when compared with the results of GoMSS (Fig. 4). (The MLD can be inferred from the depth of the 180° phase difference in current shown in Fig. 12.) It is speculated that the stronger density stratification and occasionally shallower MLD are responsible for the stronger NIOs simulated by GoMSS in the offshore region. Another interesting feature in Fig. 12 is the higher mode variability evident in the ADCP observations and GoMSS simulations at depths exceeding 100 m. This is consistent with a more variable density field shown in the results of GoMSS (i.e., density jumps at ~100 m shown in Figs. 4j–l) relative to Dal-Coast (Figs. 4d–f). Specifically, the growing strength of the second mode inertial wave in the observations and GoMSS simulations could be attributable to the thick bottom ML over which the bottom stress plays an active role (MacKinnon and Gregg 2005).

The spatial coherence of the NIOs relative to location T2 simulated by GoMSS are in reasonable agreement with the HF-radar observations (Figs. 8d,f). Specifically, both maps show strong anisotropy with a more rapid drop in coherence in the onshore–offshore direction. The coherence of the DalCoast simulations decreases more slowly in all directions, with coherences exceeding 0.8 (Figs. 8d,e). This is consistent with the relatively high proportion of total variance of the DalCoast simulations accounted for by the first EOF. The different coherence scales can be explained by the fact that DalCoast’s density field is relatively smoother, and less variable in time, as a result of its use of the spectral-nudging method, and this reduces the amount of scattering of the NIOs. The phase maps of the simulated NIOs (Figs. 8h,i) show that both models reproduce reasonably well the observed NIO phase changes moving from west to east (Fig. 8g).

The horizontal distribution of the relative frequency shift ($\Delta f = f_p - f$) for the simulations produced by DalCoast and GoMSS (Figs. 9b,c) are generally similar to the frequency shift of the HF-radar observations (Fig. 9a), with negative values of $\Delta f$ close to shore and positive values with distance from shore. The maps of simulated $\Delta f$ are also broadly consistent with the corresponding maps of $\zeta/2$ (comparing Fig. 9b with Fig. 9e and Fig. 9c with Fig. 9f), thereby providing further evidence for the effect of the NSC on $f_p$ on the inner ScS. The model results of DalCoast provide more realistic maps of $\Delta f$ and $\zeta/2$ than the results of GoMSS, presumably because of its use of the spectral-nudging and semiprognostic methods that ensure its simulated NSC and background vorticity field remain close to the observed winter mean state in DalCoast.

5. Summary and discussion

The near-inertial oscillations on the inner Scotian shelf have been examined using HF-radar and ADCP observations (December 2015–February 2016), a simple
The most interesting finding is that the observed peak frequency of the NIOs $f_p$ varied with time by about 7% of the local inertial frequency $f$, as demonstrated in the evolving rotary spectrum (ERS) of the ADCP observations. The variation in $f_p$ can be explained in part by local wind forcing as demonstrated by a simple slab model. The remaining variation of $f_p$ can be explained in part by variations in the background vorticity [Eq. (1)] associated with changes in the strength and position of the unstable Nova Scotia Current. The horizontal distribution of $f_p$ estimated from the HF-radar observations shows that it was below $f$ within ~50 km of the coast and slightly above $f$ seaward of this point, consistent with the frequency shifts caused by the changes in background vorticity. This provides additional observational evidence for Eq. (1) and modification of $f_p$ by the NSC.

Analysis of the HF-radar and ADCP observations also showed that the NIOs on the inner ScS are driven primarily by time variations in wind stress associated with the passage of storms. Individual NIO events
typically lasted from three to four inertial periods. NIOs with speeds exceeding 0.25 m s\(^{-1}\) were observed in the offshore part of the study region but their amplitudes decreased shoreward within ~40 km of the coast, consistent with the effects of decreasing water depth, increasing MLD, and bottom friction as the coast is approached. The observed NIOs during the study period had spatial scales of ~80 and ~40 km in the alongshore and cross-shore directions, respectively. The NIO phases varied moving from west to east, consistent with typical movement of winter storms in this region.

The coastal suppression of NIOs and modulation of \(f_p\) by the background vorticity was also found off the coast of Oregon (appendix A) thereby indicating the wider applicability of the present study beyond the ScS.

Two operational shelf circulation models (DalCoast and GoMSS) were used to simulate the observations. Both models reproduce the observed spatial structure of the NIOs and, in a qualitative sense, the temporal and spatial variations in \(f_p\). DalCoast benefits from the use of spectral nudging and the semiprognostic approach to ensure its simulated NSC and background vorticity field remain close to the observed winter mean state. However, these bias corrections also result in underestimation of the low-frequency circulation variability, which further degrades DalCoast’s simulations of the time variation of \(f_p\), the spatial scales of the NIOs, and the high mode variability of the NIOs in the vertical. In contrast, GoMSS allows the density field to evolve freely and uses more realistic open boundary conditions provided by the Mercator global ocean forecast system. As a result, GoMSS shows better skill in simulating the low-frequency circulation variability and associated changes in NIOs than DalCoast. However, GoMSS performs less well in simulating the mean circulation and misplaces the position of the NSC, despite the much
higher resolution in GoMSS. Reasons for this bias error are currently under investigation.

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APPENDIX A

NIO off the Coast of Oregon

To examine the relevance of this study to other locations, we analyzed the HF-radar observations made off the coast of Oregon for a 3-month period in 2017. This region has a similar latitude (45°–47°N) to the ScS (~44°N) and also features a seasonally varying, buoyancy-driven coastal boundary current (Mazzini et al. 2014). NIOs have also been observed in this region (Kim and Kosro 2013), similar to the ScS.

Figure A1a shows the mean surface current off Oregon from October to December 2017 based on hourly HF-radar observations. A northward surface buoyancy-driven flow, reaching peak speeds approximately between 0.1 and 0.2 m s⁻¹, is evident within ~25 km of the coast. Figure A1b shows the spatial distribution of the amplitude of NIOs for the same period. In accord with the results of our study on the ScS, the NIOs are suppressed within ~20 km of the Oregon coast. This is consistent with the study of Kim and Kosro (2013) based on an analysis of two years (2007–2008) of HF-radar observations. They term the suppression “coastal inhibition.”

We next selected two locations (O1 and O2; Fig. A1b) associated with relatively strong background vorticity and NIOs in order to examine the effect of the background vorticity on the temporal changes of \( f_p \). Again, we used the slab model driven by reanalysis winds to quantify the effect of local wind forcing on the temporal changes of \( f_p \). In accord with the approach used for the ScS study, we selected decay time scales \( \lambda^{-1} \) based on visual examination of the time series of observed currents and the shape of the associated rotary spectra. This resulted in decay scales of four and three inertial periods at O1 and O2, respectively.

Rotary spectral analysis (Figs. A2a,b; Figs. A3a,b) shows reasonable agreement between the slab-model simulations and the HF-radar observations in the near-inertial band at both locations. The ERS (Figs. A2c–e; Figs. A3c–e) show that the vorticity correction based on the background vorticity improves the \( f_p \) simulated by the slab model, particularly around 1) 12 October and 15 November at O1 and 2) 20 October, 15 November, and 20 December at O2. This provides additional observational evidence for the relative vorticity correction given by Eq. (1).

Overall, the coastal suppression of NIOs and the modulation of \( f_p \) by the background vorticity found off the Oregon coast are consistent with our findings for the ScS. This provides support for the relevance of the ScS study to other locations.

![Fig. B1. Time series of observed and simulated (a) tidal and (b) nontidal sea surface elevations at a tide gauge in Halifax Harbour. The observed and simulated sea surface elevations are decomposed into tidal and nontidal components using the MATLAB software package ‘‘T_TIDE.’’](image)
APPENDIX B

Model Validation

To quantify model performance, we use the $\gamma^2$ statistic of Thompson and Sheng (1997):

$$\gamma^2 = \frac{\text{Var}(O - M)}{\text{Var}(O)},$$

where $\text{Var}$ denotes variance and $O$ and $M$ denote observation and simulation, respectively. Small values of $\gamma^2$ indicate good model performance. If $\gamma^2 > 1$, then the model simulations are worse than the mean of the observations. The observed and simulated sea levels for Halifax (see Fig. 1 for location) were first decomposed into tidal and nontidal components using the analysis package of Pawlowicz et al. (2002). The observed tides (Fig. B1a) are predominantly semidiurnal and have a spring–neap variation of ~0.5 m. The observed nontidal component (Fig. B1b) includes several surges associated with winter storms. Both models reproduce well the tide and nontidal component; $\gamma^2$ is 0.03 (DalCoast) and

Fig. B2. Distributions of $\gamma^2$ for the (left) alongshore and (right) cross-shore currents produced by (a),(b) DalCoast and (c),(d) GoMSS.
0.05 (GoMSS) for tide and is 0.17 for the non-tidal component for both models. The agreement also indicates that the different open boundary conditions (see section 3) used by the two models are adequate for tides and surges.

Figure B2 shows distributions of $\gamma^2$ for the alongshore (left panels) and cross-shore (right panels) currents. The $\gamma^2$ distributions for the alongshore currents (Figs. B2a,c) are similar for both models with the highest values in the vicinity of the NSC. The lowest $\gamma^2$ (below 0.4) are found close to the coast. The $\gamma^2$ distributions for the cross-shore currents (Figs. B2b,d) are generally aligned with the cross-shore direction with the highest values near the edges of the grid. DalCoast performs better than GoMSS for this component of flow. Overall, for both models, there is similarity between the large-scale features in the distributions of $\gamma^2$ and the maps of GDOP for both flow components. This implies that errors in the HF-radar observations have a significant effect on the assessment of model performance. Nevertheless, the $\gamma^2$ values are less than 0.6 over most of the grid, indicating that both models have some skill in simulating the currents observed by the HF radar.

REFERENCES


