Energy Exchange between the Mesoscale Oceanic Eddies and Wind-Forced Near-Inertial Oscillations

ZHAO JING
Department of Oceanography, Texas A&M University, College Station, Texas, and Physical Oceanography Laboratory/Qingdao Collaborative Innovation Center of Marine Science and Technology, Ocean University of China, and Qingdao National Laboratory for Marine Science and Technology, Qingdao, China

LIXIN WU
Physical Oceanography Laboratory/Qingdao Collaborative Innovation Center of Marine Science and Technology, Ocean University of China, and Qingdao National Laboratory for Marine Science and Technology, Qingdao, China

XIAOHUI MA
Department of Oceanography, Texas A&M University, College Station, Texas, and Physical Oceanography Laboratory/Qingdao Collaborative Innovation Center of Marine Science and Technology, Ocean University of China, and Qingdao National Laboratory for Marine Science and Technology, Qingdao, China

(Manuscript received 15 September 2016, in final form 3 November 2016)

ABSTRACT

In this study, the energy exchange between mesoscale eddies and wind-forced near-inertial oscillations (NIOs) is theoretically analyzed using a slab mixed layer model modified by including the geostrophic flow. In the presence of strain, there is a permanent energy transfer from mesoscale eddies to NIOs forced by isotropic wind stress. The energy transfer efficiency, that is, the ratio of the energy transfer rate to the near-inertial wind work, is proportional to $S^2/f_{\text{eff}}^2$, where $S^2$ is the total strain variance, $f_{\text{eff}} = \sqrt{(f + \zeta/2)^2 - S^2/4}$ is the effective Coriolis frequency, and $\zeta$ is the relative vorticity. The theories derived from the modified slab mixed layer model are verified by the realistic numerical simulation obtained from a coupled regional climate model (CRCM) configured over the North Pacific. Pronounced energy transfer from mesoscale eddies to wind-forced NIOs is localized in the Kuroshio Extension region associated with both strong near-inertial wind work and strain variance. The energy transfer efficiency in anticyclonic eddies is about twice the value in cyclonic eddies in the Kuroshio Extension region because of the influence of $\zeta$ on $f_{\text{eff}}$, which may contribute to shaping the dominance of cyclonic eddies than anticyclonic eddies in that region.

1. Introduction

Near-inertial oscillations (NIOs) are a ubiquitous feature throughout the global upper ocean (Garrett 2001). They form a pronounced peak in the ocean current frequency spectrum, containing half of the kinetic energy and a substantial portion of the shear in the internal wave field (Ferrari and Wunsch 2009). They are of central importance to a variety of ocean processes, including the mixed layer deepening (Greatbatch 1984; Price et al. 1986; Jochum et al. 2013) and phytoplankton dispersion (Franks 1995). In addition, they are thought to provide an energy source for abyssal diapycnal mixing, which affects the uptake of heat and carbon by the oceans as well as climate changes (Munk and Wunsch 1998; Wunsch and Ferrari 2004; Jing and Wu 2014).

As a natural resonant frequency of fluids on a rotating planet, NIOs are efficiently forced by time-varying wind stresses. Most of the wind work on NIOs occurs when

---

Corresponding author e-mail: Zhao Jing, jingzhao198763@tamu.edu

DOI: 10.1175/JPO-D-16-0214.1

© 2017 American Meteorological Society.
moving wind storm systems pass over the ocean (Alford 2003; Jing et al. 2016). Given that the storm tracks are remarkably coincident with regions of strong mesoscale eddy activities (Zhai et al. 2005), there are ample opportunities for interactions between NIOs and mesoscale eddies. In particular, the energy exchange between NIOs and mesoscale eddies is proposed to be a potentially important component in the oceanic energy budget (e.g., Ferrari and Wunsch 2009; Gertz and Straub 2009; Polzin 2010; Danioux et al. 2012; Thomas 2012; Vanneste 2013; Xie and Vanneste 2015). On one hand, it can be a significant sink of mesoscale eddy kinetic energy. On the other hand, it may also provide a substantial portion of energy for NIOs to induce diapycnal mixing.

Both NIOs and mesoscale eddies are most energetic in the mixed layer, implying that the mixed layer might be an important region for energy exchange between NIOs and mesoscale eddies. However, most of the existing mechanisms focus on the energy exchange between NIOs and mesoscale eddies in a regime without external forcing (e.g., Müller 1976; Bühler and McIntyre 2005; Thomas 2012; Vanneste 2013; Xie and Vanneste 2015). They are not always applicable to the mixed layer where NIOs are intermittently forced by passing wind storms. So far it remains unclear what mechanism is responsible for the permanent energy exchange between mesoscale eddies and NIOs under the wind forcing and how important it is to the oceanic energy budget. In a recent study by Whitt and Thomas (2015), it is suggested that the energy transfer from mesoscale eddies to wind-forced NIOs is proportional to \((\zeta/ f)^2 W\), where \(W\) is the wind work on NIOs, \(f\) is the Coriolis frequency, and \(\zeta\) is the relative vorticity of mesoscale eddies. However, Whitt and Thomas (2015) considered a unidirectional laterally sheared geostrophic flow, that is, \(U = U(y)\), \(V = 0\). In this case, \(\zeta = -U_y\), can be alternatively interpreted as the shear strain, making the mechanism responsible for the energy exchange ambiguous. In this study, we attempt to understand whether the strain, relative vorticity, or both is responsible for the permanent energy exchange by considering a geostrophic flow of the generalized form. To facilitate theoretical analyses, we consider a geostrophic flow characterized by a small Rossby number \(R_o\), which is largely satisfied by mesoscale eddies in the ocean. In fact, numerical simulations suggest that the theoretical results derived in this study show good skills in the Kuroshio Extension region where energetic mesoscale eddies are associated with a root-mean-square (RMS) \(R_o\) of 0.1 ~ 0.3. The paper is organized as follows: A slab mixed layer model modified by including the effect of geostrophic flow is presented in section 2. In section 3, we perform theoretical and numerical analyses based on the modified slab model. The theoretical results in section 3 are further tested against realistic numerical simulations obtained from a high-resolution coupled regional climate model (CRCM) configured over the entire North Pacific (Ma et al. 2016). The numerical simulations also provide an opportunity to evaluate the role of energy exchange between mesoscale eddies and wind-forced NIOs in the oceanic energy budget. Finally, conclusions are presented in section 5 followed by discussion.

2. The modified slab mixed layer model

The slab mixed layer model (Pollard and Millard 1970) has been widely used to simulate the generation of NIOs in the mixed layer by wind stress (e.g., D’Asaro 1985; Alford 2001; Alford 2003; Jiang et al. 2005). The slab mixed layer model modified by including the geostrophic flow is

\[
\begin{align}
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} &= f v - r u + T_x, \quad \text{and} \\
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} &= -f u - r v + T_y
\end{align}
\]

(Weller 1982), where \((u, v)\) is the near-inertial current in the mixed layer; \((U, V)\) is the geostrophic flow, which is horizontally nondivergent; \((T_x, T_y)\) is the wind stress \((r_x, r_y)\) normalized by the product of seawater density \(\rho_0\) and mixed layer depth \(H_M\) and \(r\) is the radiative damping parameter modeling the slow downward radiation of NIOs from the mixed layer to the thermocline. Previous studies suggest that \(r \ll f\), and under this condition the modeled near-inertial current is not sensitive to the value of \(r\) (D’Asaro 1985; Alford 2001; Park et al. 2009).

The validity of (1) is built on the following two assumptions: (i) the \(R_o\) of geostrophic flow is much smaller than unity but is much larger than the Burger number \(B_u\) of wind-forced NIOs, and (ii) the horizontal scale of winds is much larger than that of mesoscale eddies. As shown in Appendix A, these two assumptions are reasonable for mesoscale eddies and wind-forced NIOs at the midlatitudes.

---

1. Müller (1976) and Bühler and McIntyre (2005) deal with not only NIOs but also higher-frequency internal waves.
The wind work on NIOs $W$ and energy transfer rate $\varepsilon$ from mesoscale eddies to NIOs can be derived from the kinetic energy budget of NIOs:

$$\frac{\partial KE}{\partial t} = \varepsilon - 2rKE + W,$$

(2)

where $KE = \rho_0 H_M (u^2 + v^2)/2$ is the near-inertial kinetic energy in the mixed layer and

$$W = \tau_x u + \tau_y v, \quad \text{and} \quad \varepsilon = -\rho_0 H_M (uu_{x} + uv_{y} + vv_{y} + v\omega_x + \omega_y v).$$

(3a)

3. Theoretical and numerical analyses based on the modified slab mixed layer model

a. Theoretical analyses

In this section, we will derive the analytical solution to the energy exchange between the mesoscale eddies and wind-forced NIOs using the modified slab mixed layer model (1). To elucidate the physics underlying (1), we rewrite (1) in the matrix (tensor) form:

$$\frac{\partial}{\partial t} \begin{pmatrix} u \\ v \end{pmatrix} = \begin{pmatrix} -r & f \\ -f & -r \end{pmatrix} \begin{pmatrix} u \\ v \end{pmatrix} - \begin{pmatrix} U_x \\ U_y \end{pmatrix} \begin{pmatrix} u \\ v \end{pmatrix} + \begin{pmatrix} T_x \\ T_y \end{pmatrix}.$$

(4)

The deformation tensor of mesoscale eddies can be further decomposed into the strain tensor and relative vorticity tensor:

$$\begin{pmatrix} U_x & U_y \\ V_x & V_y \end{pmatrix} = \frac{1}{2} \begin{pmatrix} S_{nx} & S_{nx} \\ S_{ny} & S_{ny} \end{pmatrix} + \frac{1}{2} \begin{pmatrix} 0 & -\xi \\ -\xi & 0 \end{pmatrix},$$

(5)

where $S_{nx} = 2U_x$ is the normal strain component in the $x$ direction, $S_{ny} = 2V_y$ is the normal strain component in the $y$ direction, $S_x = V_x + U_y$ is the shear strain, and $\xi = V_x - U_y$ is the relative vorticity. As mesoscale eddies are assumed to be geostrophic and thus horizontally nondivergent, we have $S_{nx} = -S_{ny} = S_{n}.$

As the strain tensor is symmetric, it is always possible to choose the directions of the orthogonal axes (i.e., the principal axes) of reference so that the nondiagonal elements are zero (Batchelor 1967). In this case, (4) can be simplified as

$$\begin{pmatrix} u' \\ v' \end{pmatrix} = \begin{pmatrix} -r - f/2 & f + \xi/2 \\ -f - \xi/2 & -r + S_{n}/2 \end{pmatrix} \begin{pmatrix} u' \\ v' \end{pmatrix} + \begin{pmatrix} T_x' \\ T_y' \end{pmatrix},$$

(6)

where variables denoted by primes correspond to values in the principal axes. It should be noted that both the total strain variance (i.e., the sum of normal strain variance and shear strain variance) and relative vorticity are invariants for the rotational transform so that $\xi' = \xi$ and $S_{n}' = S_{n}^2 + S_{\xi}^2.$

Equation (6) is a $2 \times 2$ system of first-order, linear, ordinary differential equations. The eigenvalue of the coefficient matrix on the right-hand side of (6) is

$$\lambda = -r \pm i \sqrt{\left(f + \frac{\xi}{2}\right)^2 - \frac{S_{n}^2}{4}},$$

(7)

The real part of $\lambda$ corresponds to the exponential decay of NIOs due to the linear damping in (1). The imaginary part of $\lambda$ is the effective Coriolis frequency $f_{\text{eff}},$ which is the natural frequency of NIOs in the presence of mesoscale eddies (Fomin 1973; Chavanne et al. 2012). Both the strain and relative vorticity can affect $f_{\text{eff}}$ (Fig. 1). The strain only affects $f_{\text{eff}}$ to $O(\text{Ro}^2),$ whereas the relative vorticity affects $f_{\text{eff}}$ to $O(\text{Ro}).$ Therefore, for mesoscale eddies characterized by a small Ro, the relative vorticity has a more dominant influence on $f_{\text{eff}}$ than the strain.

Using the relations $\xi' = \xi$ and $S_{n}' = S_{n}^2 + S_{\xi}^2,$ the expression of $f_{\text{eff}}$ in the original axes is

$$f_{\text{eff}} = \sqrt{\left(f + \frac{\xi}{2}\right)^2 - \frac{S_{n}^2 + S_{\xi}^2}{4}},$$

(8)
Note that (8) reduces to the expression $f_{\text{eff}} = \sqrt{f_U}$ derived by Whitt and Thomas (2015) for a unidirectional laterally sheared geostrophic flow.

The response of the initially quiescent mixed layer to wind forcing can be divided into two stages: a transient spinup followed by forced-dissipative equilibrium. As we are only interested in the second stage, (6) can be solved in the frequency domain using a Fourier transform:

$$
\begin{bmatrix}
\omega + r + S_n/2 & -f - \zeta/2 \\
f + \zeta/2 & \omega + r - S_n/2
\end{bmatrix}
\begin{bmatrix}
u' \\
v' 
\end{bmatrix} = 
\begin{bmatrix}
\bar{T}_x \\
\bar{T}_y
\end{bmatrix}, 
(9)
$$

where $\omega$ is the frequency and variables with a tilde represent the Fourier transform. To derive (9), mesoscale eddies are assumed to be almost steady within an inertial period, which is not unreasonable as mesoscale eddies typically evolve on a time scale much longer than the inertial period.

The solution to (9) is

\begin{align}
\bar{u}' &= \frac{(r - S_n/2 + i\omega)\bar{T}_x + (f + \zeta/2)\bar{T}_y}{(i\omega + r)^2 + f_{\text{eff}}^2}, \quad \text{and} \\
\bar{v}' &= \frac{(r + S_n/2 + i\omega)\bar{T}_x - (f + \zeta/2)\bar{T}_y}{(i\omega + r)^2 + f_{\text{eff}}^2}.
\end{align}

(10a)

(10b)

According to Parseval’s theorem,

\begin{align}
E\{W\} &= \rho_0 H_M E \left\{ \int \frac{r(r^2 + f_{\text{eff}}^2 + \omega^2)}{[(\omega + f_{\text{eff}})^2 + r^2][(\omega - f_{\text{eff}})^2 + r^2]} \left( |\bar{T}_x|^2 + |\bar{T}_y|^2 \right) \, d\omega \right\}, \quad \text{and} \\
E\{e\} &= \frac{\rho_0 H_M}{2} E \left\{ \int \frac{rS_n^2}{[(\omega + f_{\text{eff}})^2 + r^2][(\omega - f_{\text{eff}})^2 + r^2]} \left( |\bar{T}_x|^2 + |\bar{T}_y|^2 \right) \, d\omega \right\},
\end{align}

where $E\{\cdot\}$ represents the ensemble mean.

As the integrand in (12b) is definitely positive, the energy is irreversibly transferred from mesoscale eddies to wind-forced NIOs. This permanent energy transfer corresponds to a negative correlation between $\langle u'u' \rangle - \langle v'v' \rangle$ and $S_n$, since $\epsilon = -\rho_0 H_M (u'u' - v'v') S_n$, in the principal axes. The mechanism responsible for this negative correlation can be understood based on the equations governing the evolution of $u'u'$ and $v'v'$:

\begin{align}
\frac{\partial u'u'}{\partial t} &= \left( f + \frac{\zeta}{2} \right) u'v' - ru'u' - u'u' S_n + u'T_y, \\
\frac{\partial v'v'}{\partial t} &= - \left( f + \frac{\zeta}{2} \right) u'u' - rv'u' + v'v' S_n + v'T_x.
\end{align}

(13a)

(13b)

For wind-forced NIOs, $u'T_x$ and $v'T_y$ act as production terms for $u'u'$ and $v'v'$, respectively. The influence of strain on $u'u'$ and $v'v'$ depends on the sign of $S_n$. Here, we assume $S_n > 0$ without the loss of generality. In this case, the strain acts to reduce $u'u'$ produced by winds through the energy exchange with mesoscale eddies, as the momentum flux $u'u'$ is upgradient. The opposite is true for $v'v'$. Therefore, the strain induces a negative correlation between $\langle u'u' \rangle - \langle v'v' \rangle$ and $S_n$. and a
net energy transfer from mesoscale eddies to wind-forced NIOs.

The efficiency of energy transfer from mesoscale eddies to wind-forced NIOs can be measured by the ratio of the ensemble-mean energy transfer rate to ensemble-mean wind work, that is, \( R = E\{e\}/E\{W\} \). Given that both the expressions for \( E\{e\} \) and \( E\{W\} \) contain integrals, the expression for \( R \) is too complicated to make a clear, dynamical interpretation. As mentioned in section 2, \( \gamma \) is much smaller than \( f \) so that \( r \ll f \text{eff} \) in case of a small \( \text{Ro} \). This motivates us to approximate \( R \) by its limit for \( n/f \text{eff} \to 0 \):

\[
R_L = \lim_{n/f_{\text{eff}} \to 0} R = E\left\{ \frac{S_n^2}{4f_{\text{eff}}^2} \right\}. \quad (14)
\]

A detailed derivation for (14) is provided in appendix B. Using the relations \( \zeta' = \xi \) and \( S_{n'}^2 = S_n^2 + S_{\gamma}^2 \), the counterpart of \( R_L \) in the original axes is

\[
R_L = E\left\{ \frac{S_n^2 + S_{\gamma}^2}{4f_{\text{eff}}^2} \right\}. \quad (15)
\]

It should be noted that (15) reduces to the expression \( e/W = U^2/[4(f - U_{\gamma})] \), derived by Whitt and Thomas (2015) for a unidirectional laterally sheared geostrophic flow \( [U = U(y), V = 0] \). However, the more general case considered in this study reveals that it is the strain that is responsible for the permanent energy transfer from mesoscale eddies to wind-forced NIOs, as \( R_L \) is always zero in absence of strain. The relative vorticity alone does not induce any permanent energy exchange. But it can have an influence on the energy transfer efficiency in the presence of strain by modifying \( f_{\text{eff}} \). A negative \( \zeta \) acts to decrease \( f_{\text{eff}}^2 \), leading to an increased value of \( R_L \) (Fig. 2). The opposite is true when \( \zeta \) is positive. However, this does not mean that the relative vorticity in a turbulent geostrophic flow with both positive and negative values has no net effects on the energy transfer efficiency. In fact, \( R_L \) averaged in an anticyclonic eddy and a cyclonic eddy of equal strength is always larger than the value in absence of relative vorticity according to the inequality \( \frac{1}{2}[(a + b)^{-2} + (a - b)^{-2}] > a^{-2} \) for \( |a| > |b| \) (Fig. 2).

b. Numerical simulations using the modified slab mixed layer model

In this section, the validity of the analytical solutions is tested against the numerical simulations based on the modified slab mixed layer model [(1)]. The wind stress is obtained from the Kuroshio Extension Observatory (KEO) mooring during September 2007–September 2008. As the focus is on the wind-forced NIOs, the wind stress is high-pass filtered with a cutoff frequency of 0.2f. For the high-pass filtered wind stress, the frequency spectra of \( \tau_x \) and \( \tau_y \) have similar energy levels (Fig. 3a). The correlation coefficient between \( \tau_x \) and \( \tau_y \) is about –0.02, not significant at the 5% significance level. Therefore, the high-pass filtered wind stress can be treated as isotropic. The rotary spectrum of high-pass filtered wind stress reveals slightly stronger CW rotation (Fig. 3b). But the difference between the CW- and CCW-rotated components is generally within a factor of 2 at the inertial and higher frequencies. All these features are qualitatively consistent with our assumption in section 3a.

To distinguish the influences of strain and relative vorticity on \( e \), two kinds of idealized geostrophic flow are constructed. In the experiment Ex-S, the relative vorticity is set to be zero and the normal and shear strain are modeled as \( S_n = |\text{Ro}|\cos(\omega_{\gamma} + \theta_n) \) and \( S_{\gamma} = |\text{Ro}|\cos(\omega_{\gamma} + \theta_\gamma) \), where \( \omega_{\gamma} = 1/60 \text{cpd} \), \( \theta_n \) and \( \theta_\gamma \) are random phases independent from each other, and \( |\text{Ro}|^2 \) corresponds to total strain variance. In the experiment Ex-SV, the normal and shear strain are the same as those in Ex-S, but the relative vorticity is modeled as \( \zeta = \sqrt{2}|\text{Ro}|\cos(\omega_{\gamma} + \theta_\zeta) \), where \( \theta_\zeta \) is a random phase independent from \( \theta_n \) and \( \theta_\gamma \).

The Ro is varied from 0.1 to 0.6, which covers the typical values in the Kuroshio Extension region (see section 4 for details). For each value of Ro, we perform 100 Monte Carlo simulations in which \( \theta_n \), \( \theta_\gamma \), and \( \theta_\zeta \) are generated
randomly from a uniform distribution between 0 and 2π. It should be noted that as the modified slab mixed layer model \((1)\) is strictly valid only for small \(R_o\), the results derived in the moderate \(R_o\) case should be treated as suggestive rather than definitive.

In both Ex-S and Ex-SV, \(E\{\phi\}\) is not sensitive to the changes of \(R_o\) (Fig. 4a). It varies by less than 30% for \(R_o\) ranging from 0.1 to 0.6, consistent with the findings reported by Klein et al. (2004). In contrast, \(E\{e\}\) exhibits pronounced enhancement with the increased \(R_o\) in both Ex-S and Ex-SV (Fig. 4b) due to the rapidly elevated energy transfer efficiency (Fig. 4c). In particular, the energy transfer efficiency is systematically higher in Ex-SV than Ex-S, and the difference becomes larger as \(R_o\) increases, suggesting the important role of relative vorticity in enhancing the energy transfer efficiency. All these features are consistent with our theoretical analyses in section 3a. In particular, the energy transfer efficiency derived from the numerical simulations agrees well with \(R_L\) with a difference less than 20%, lending supports to our theoretical solutions.

4. Realistic numerical simulations in the North Pacific

a. Model configurations

The modified slab mixed layer model neglects the refraction and advection of NIOs by mesoscale eddies (Kunze 1985; Young and Ben Jelloul 1997; Balmforth et al. 1998; Bühler and McIntyre 2005; Zhai et al. 2005, 2007). Furthermore, it does not include the shear instability induced by NIOs, systematically overestimating the NIOs especially during strong wind forcing events (Plueddemann and Farrar 2006). Given these shortages of the modified slab mixed layer model, it is thus necessary to test the validity of our theoretical solutions in realistic numerical simulations based on a 3D primitive equation model. It should be noted that the theoretical
solutions in section 3a are unlikely to hold exactly due to the great uncertainties in the modified slab mixed layer model. In the following analysis, we will focus on two important qualitative features inferred from the analytical solutions, that is, (i) the linear dependence of energy transfer efficiency on total strain variance $S_n^2 = S_n^2 + S_s^2$ and (ii) its asymmetry between cyclonic and anticyclonic eddies.

A coupled regional climate model developed at Texas A&M University (Ma et al. 2016) is used to simulate NIOs and mesoscale eddies in the North Pacific. It includes the Weather Research and Forecasting (WRF) Model (Leung et al. 2006) as the atmospheric component and Regional Ocean Modeling Systems (ROMS; Moore et al. 2004; Shchepetkin and McWilliams 2005) as the oceanic component. The CRCM is configured with ROMS and WRF both at 9-km horizontal resolution. The model domain covers the entire North Pacific. In the simulation, WRF and ROMS are coupled hourly, which is sufficiently fine to resolve wind stress variance in the near-inertial band. Readers are referred to Ma et al. (2016) for more details on the model configurations.

The CRCM simulation was initialized on 1 October 2002 from a 6-yr ROMS spinup simulation (1997–2002) and integrated for 1 yr. The period 2002–03 is chosen, as it is a relatively neutral year for both the El Niño–Southern Oscillation (ENSO) and Pacific decadal oscillation (PDO), which modulate the activities of midlatitude storms and tropical cyclones (Chang and Fu 2002; Camargo and Sobel 2005). This facilitates the comparisons of the simulated NIO strength to the climatological mean derived from the surface drifters (Chaigneau et al. 2008; Elipot and Lumpkin 2008).

b. Simulation results

The simulated mesoscale eddy activity is qualitatively consistent with that obtained from the satellites (Fig. 5). In both the CRCM and observations, the standard deviations of sea surface height (SSH) exhibit pronounced enhancement in the Kuroshio Extension region. The simulated standard deviation of SSH in the Kuroshio Extension region is larger than the observed one. But their difference is less than 50%. The stronger SSH variability in the CRCM could be due to the model bias,
the interannual variation of mesoscale eddy activities, or the insufficient resolution of satellite measurements.

The simulated NIOs also agree reasonably well with the observations derived from surface drifters (Chaigneau et al. 2008; Elipot and Lumpkin 2008). The near-inertial, current amplitude averaged over the North Pacific is 12.3 cm s⁻¹, comparable to the 11.5 cm s⁻¹ obtained from the observations (Chaigneau et al. 2008). Readers are referred to Jing et al. (2016) for detailed comparisons between the simulated NIOs in CRCM and observations.

The good agreements between the CRCM simulations and observations give us confidence that the interactions between the mesoscale eddies and wind-forced NIOs simulated in the CRCM are qualitatively reliable. In the following analysis, we will focus on the Kuroshio Extension region (30°–45°N, 142°E–180°) as both the mesoscale eddies and wind-forced NIOs are the most energetic here. The RMS Ro in this region ranges from 0.1 to 0.3 so that the small Ro assumption is loosely satisfied.

In the CRCM, the flow of mesoscale eddies is computed using the geostrophic relation based on the low-pass filtered SSH with a cutoff frequency of 0.2f. This removes the contribution of wind-induced Ekman currents to strain and relative vorticity, which is beyond the scope of this study. The near-inertial signals are attained by a high-pass filter with a cutoff frequency of 0.8f (Furuichi et al. 2008). The near-inertial wind work is computed as $W_{\text{CRCM}} = u_i \tau_x + v_i \tau_y$, where $(u_i, v_i)$ is the near-inertial current on the sea surface. Sensitivity tests suggest that changing the cutoff frequency from 0.7f to 0.9f does not have any substantial impact on the following conclusions. The energy transfer rate $\varepsilon_{\text{CRCM}}$ from the geostrophic flow to wind-forced NIOs is computed following (3b) by assuming that the near-inertial current and geostrophic flow are vertically uniform in the surface boundary layer. The boundary layer height $H_{\text{bl}}$ is derived from the K-profile parameterization (Large et al. 1994). It should be noted that the assumption of uniform flow in the surface boundary layer might be violated in wintertime with the deep boundary layer as the downward transport of momentum from the sea surface to the bottom of the boundary layer may take a time longer than the inertial period (D’Asaro 2014). But this assumption is adequate for our qualitative analysis here.

As only one simulation run is available, the ensemble mean is alternatively evaluated as the zonal mean between 142°E and 180°. Figure 6a displays the zonal mean $\varepsilon_{\text{CRCM}}$ in the Kuroshio Extension region; $\varepsilon_{\text{CRCM}}$ is positive throughout 30°–45°N, suggesting a permanent energy transfer from mesoscale eddies to wind-forced NIOs. The term $\varepsilon_{\text{CRCM}}$ exhibits strong latitudinal variability with a pronounced peak around 37°N. The latitudinal variation of $\varepsilon_{\text{CRCM}}$ follows both $W_{\text{CRCM}}$ and $S^2/f_{\text{eff}}^2$ (Figs. 6b,c). In particular, the energy transfer efficiency $R_{\text{CRCM}}$, defined as $\varepsilon_{\text{CRCM}}/W_{\text{CRCM}}$, is tightly correlated to $S^2/f_{\text{eff}}^2$ with a correlation coefficient of 0.85 (significant at 5% significance level). A logarithm regression analysis reveals that $R_{\text{CRCM}}$ is proportional to $(S^2/f_{\text{eff}}^2)^{\beta}$ with a 95% confidence interval of $\beta$ ranging from 0.97 to 1.14 (Fig. 7). This is consistent with our analytical solution [(15)], which predicts $\beta = 1$.

The energy transfer $\varepsilon_{\text{CRCM}}$ exhibits evident asymmetry between cyclonic eddies and anticyclonic eddies (Figs. 8a,b). The mean $\varepsilon_{\text{CRCM}}$ in the cyclonic eddies is 0.05 mW m⁻², while it increases to 0.12 mW m⁻² in the anticyclonic eddies. The difference is statistically significant at the 5% significance level based on a ranksum test. This asymmetry of $\varepsilon_{\text{CRCM}}$ is not due to the difference of near-inertial wind work as the mean $W_{\text{CRCM}}$ in the anticyclonic eddies is only 4% larger than that in the cyclonic eddies (Figs. 8c,d). Nor can it be simply ascribed to the concentration of NIOs in the anticyclonic eddies through the trapping mechanism (Kunze 1985), as the mean, near-inertial kinetic energy.
\[ \rho_0 \frac{H_{sl}(u_i^2 + v_i^2)}{2} \] in the surface boundary layer differs by less than 10% between the cyclonic and anticyclonic eddies (Figs. 8e,f). Therefore, the enhanced \( e_{CRCM} \) in the anticyclonic eddies suggests a higher energy transfer efficiency in the anticyclonic eddies than in the cyclonic eddies. In fact, the mean \( R_{CRCM} \) in the cyclonic eddies is only about 1.01% but increases to 2.19% in the anticyclonic eddies (Figs. 8g,h). The evident asymmetry of \( R_{CRCM} \) between cyclonic and anticyclonic eddies is consistent with our analytical solutions (Fig. 2), lending further supports for the validity of our analytical solutions in the reality.

5. Conclusions

In this study, we analyzed the energy exchange between mesoscale eddies and wind-forced NIOs based on both a modified slab mixed layer model and realistic numerical simulations obtained from a 3D primitive equation model. The major conclusions are summarized as follows:

1) In the presence of strain, there is a permanent energy transfer from mesoscale eddies to NIOs forced by isotropic wind stress. The energy transfer efficiency is proportional to \( S^2/f_{eff}^2 \).

2) It is the strain that is responsible for the permanent energy transfer from mesoscale eddies to wind-forced NIOs. The relative vorticity alone does not induce any permanent exchange. But it can significantly affect the energy transfer efficiency in the presence of strain due to its modification of \( f_{eff} \). The energy transfer efficiency is enhanced in the anticyclonic eddies but reduced in the cyclonic eddies.

3) The energy exchange between mesoscale eddies and wind-forced NIOs in the realistic numerical simulation is qualitatively consistent with the analytical solutions derived from the modified slab mixed layer model. Pronounced energy transfer from mesoscale eddies to wind-forced NIOs is localized in the Kuroshio Extension region associated with strong near-inertial wind work and strain variance. The energy transfer...
efficiency in the anticyclonic eddies is about twice the value in the cyclonic eddies.

6. Discussion

The Kuroshio Extension region is recognized as one of the regions associated with the strongest mesoscale eddy activity. However, even in the Kuroshio Extension region, the RMS Ro is generally within 0.1–0.3. In this case, the energy transfer from mesoscale eddies to wind-forced NIOs is expected to play a negligible role in the oceanic near-inertial energy budget as our theoretical solutions suggest that the energy transfer efficiency is proportional to Ro² for small Ro. In fact, the mean energy transfer rate in the Kuroshio Extension region derived from the CRCM is about 0.09 mW m⁻², only accounting for ~2% of the near-inertial wind work (4.8 mW m⁻²) in this region. Given that the mesoscale eddy activities tend to be overestimated by the CRCM (Fig. 5), the energy transfer rate is probably even smaller in reality. It might be possible that the energy exchange between submesoscale (<10 km) flow and wind-forced NIOs plays a more important role in the energy budget of NIOs as the submesoscale flow is characterized by an O(1) Ro.

Although the overall magnitude of the energy transfer rate from mesoscale eddies to wind-forced NIOs is negligible, it could make an important contribution to shaping the dominance of cyclonic eddies than anticyclonic eddies in the upper ocean. Recent observations based on satellites reveal that the cyclonic eddies are characterized by stronger eddy kinetic energy than anticyclonic eddies in the upper ocean. Recent observations based on satellites reveal that the cyclonic eddies are characterized by stronger eddy kinetic energy than anticyclonic eddies in the upper ocean. Recent observations based on satellites reveal that the cyclonic eddies are characterized by stronger eddy kinetic energy than anticyclonic eddies in the Kuroshio Extension region (Cheng et al. 2014). As demonstrated in this paper, the energy transfer efficiency from mesoscale eddies to wind-forced NIOs is significantly enhanced in the anticyclonic eddies compared to the cyclonic eddies. This could lead to larger turbulent horizontal viscosity in the anticyclonic eddies both through the stronger stress induced by NIOs or through the elevated shear flow dispersion driven by the combined action of vertical mixing and vertical shear of NIOs (Young et al. 1982; Klein et al. 2003), contributing to stronger eddy kinetic energy in the cyclonic eddies than anticyclonic eddies.

This study suggests that the strain is a crucial ingredient for the energy exchange between mesoscale eddies and NIOs during the wind forcing stage. It is worth noting that the important role of strain in the energy exchange has also been demonstrated by Bühler and McIntyre (2005) and Thomas (2012) at the post-forcing stage when winds have vanished and NIOs are refracted and advected by mesoscale eddies. In fact, the physical term responsible for the energy exchange is the same among Bühler and McIntyre (2005), Thomas (2012), and this study, that is, \(-((u'u') - (v'v'))S_{\tau}/2\). However, the mechanisms leading to nonzero Reynolds stress induced by NIOs differ among different studies due to the differences in model setup. In Thomas (2012), nonzero Reynolds stress is due to the vertical shear of background flow, which affects the polarization relation of NIOs. In Bühler and McIntyre (2005), nonzero Reynolds stress results from a positive Okubo–Weiss parameter (Provenzale 1999) imposing a permanent influence on the azimuth of the horizontal wave vector of NIOs. In contrast, the key mechanism in this study is the combined effect of isotropic wind forcing and strain, which leads to stronger inertial response in the direction where background flow converges. The mechanisms proposed by Bühler and McIntyre (2005) and Thomas (2012) probably dominate the post–wind forcing stage, while the mechanism in this study might be more important during the wind forcing stage.

Acknowledgments. This work is supported by China National Natural Science Foundation (NSFC) Key Project (41130859), NSFC Major Project (41490643), National Major Research Plan of Global Change (2013CB956201), and NSFC-Shandong Joint Fund for Marine Science Research Centers. Z. J. is partly supported by the China Scholarship Council. The CRCM simulation is supported by U.S. National Science Foundation Grants AGS-1462127 and AGS-1067937. We are grateful to Dr. Ping Chang and Dr. Raffaele Montuoro for their development effort for CRCM. We thank Kuroshio Extension Observatory and AVISO for providing the data through online access.

APPENDIX A

Scaling the Slab Mixed Layer Model Equations

In the mixed layer, the linearized horizontal momentum equation for NIOs in the presence of geostrophic flow is

\[ \frac{\partial \mathbf{u}_h}{\partial t} + \mathbf{u}_h \nabla_h \mathbf{u}_h + \nabla_h \mathbf{u}_h \mathbf{U}_h + w \frac{\partial \mathbf{U}_h}{\partial z} = f \mathbf{u}_h \times \mathbf{k} - \nabla_h \rho + \frac{1}{\rho_0} \frac{\partial \tau}{\partial z}, \]

where \( \mathbf{U}_h = (U, V) \) is the horizontal velocity vector of geostrophic flow, \( \mathbf{u}_h = (u, v) \) is the horizontal velocity of NIOs, \( w \) is the vertical velocity of NIOs, \( \rho \) is the pressure of NIOs normalized by seawater density, \( \tau = (\tau_x, \tau_y) \) is
the wind stress, $\mathbf{k}$ is a unit vector pointing upward, and $\nabla_h = (\partial/\partial x, \partial/\partial y)$.

Let $L_N$ ($L_G$), $H_N$ ($H_G$), and $u^*$ ($U^*$) represent the horizontal scale, vertical scale, and horizontal velocity magnitude of NIOs (geostrophic flow). And let $\tau^*$ and $N_0$ denote the magnitude of wind stress and buoyancy frequency in the surface mixed layer. The magnitude of individual terms in (A1) is

$$\mathbf{u}_h \nabla_h \mathbf{u}_h = f u^* \mathbf{Ro},$$

where $\mathbf{Ro} = U^*(fL_G)$ is the Rossby number of geostrophic flow, $Bu = N_0^2H_G^2/(f^2L_N^2)$ is the Burger number of NIOs, and $H_M$ is the mixed layer depth.

The relative importance of each term in (A1) depends on the nondimensional parameters $L_G/L_N$, $H_N/H_G$, $\mathbf{Ro}$, and $Bu$. For wind-forced NIOs, $L_N$ is basically determined by winds, and $H_N$ can be chosen as $H_M$. As the vertical extent of geostrophic flow is typically much deeper than the mixed layer, we have $H_N/H_G \ll 1$. Assuming $L_G/L_N \ll 1$ and $Bu \ll \mathbf{Ro}$, (A1) can be approximated as

$$\frac{\partial \mathbf{u}_h}{\partial t} + \mathbf{u}_h \nabla_h \mathbf{u}_h = f \mathbf{u}_h \times \mathbf{k} + \frac{1}{\rho_0} \frac{\partial \mathbf{\tau}}{\partial z}. \quad (A2)$$

Vertically integrating (A2) in the mixed layer and adding a linear damping term yields (1).

The assumptions $L_G/L_N \ll 1$ and $Bu \ll \mathbf{Ro}$ are reasonable at the midlatitudes. At the midlatitudes, NIOs are primarily generated by synoptic storms whose horizontal scale is $\sim 1000$ km. In contrast, the horizontal scale of mesoscale eddies is $\sim 100$ km so that $L_G/L_N \ll 1$. In addition, given $f \sim 10^{-4}$ rad s$^{-1}$, $H_M \sim 50$ m, and $N_0 \sim 10^{-3}$ rad s$^{-1}$, $Bu$ is $\sim 2.5 \times 10^{-7}$. This is much smaller than the RMS Ro ($0.1 \sim 0.3$) of mesoscale eddies in the Kuroshio Extension region derived from our numerical simulations. Therefore, it is reasonable to assume $L_G/L_N \ll 1$ and $Bu \ll \mathbf{Ro}$.

Finally, it should be noted that mesoscale eddies can imprint their horizontal scales on NIOs (Young and Ben Jelloul 1997; Balmforth et al. 1998), which might violate the assumption $L_G/L_N \ll 1$. However, this imprint effect occurs on a time scale of $(Ro)^{-1}$ (Young and Ben Jelloul 1997) and thus does not play an important role at the wind forcing stage for $Ro \ll 1$. Therefore, to guarantee the validity of (A2), we also assume $Ro \ll 1$, which is largely satisfied in the Kuroshio Extension region where the RMS Ro of mesoscale eddies ranges from 0.1 to 0.3.

APPENDIX B

Analytical Solutions to Wind Work and Energy Transfer Rate

For isotropic wind forcing with equal magnitudes of CW- and CCW-rotated components, we have the following relations:

$$E(|\bar{T}_x^2|) = E(|\bar{T}_y^2|), \quad (B1a)$$

$$E(\bar{T}_x \bar{T}_y + \bar{T}_y \bar{T}_x) = 0, \quad \text{and} \quad (B1b)$$

$$E(\bar{T}_x \bar{T}_y - \bar{T}_x \bar{T}_y) = 0, \quad (B1c)$$

where (B1a) and (B1b) correspond to the isotropy, and (B1c) corresponds to equal magnitudes of CW- and CCW-rotated components. According to (3a) and the convolution theorem, the time mean $W$ is equal to

$$W = \text{Re} \left\{ \rho_0 H_M \int \bar{u} \bar{T}_x^{\omega_x} + \bar{v} \bar{T}_y^{\omega_y} \, d\omega \right\}. \quad (B2)$$

Substituting (10) into (B2) yields

$$W = \rho_0 H_M \left[ \frac{r(r^2 + f^2_{\text{eff}} + \omega^2)}{[\omega + f_{\text{eff}}]^2 + r^2] [\omega - f_{\text{eff}}]^2 + r^2]} \right] (|\bar{T}_x^2| + |\bar{T}_y^2|) \, d\omega$$

$$- \frac{\rho_0 H_M}{2} \left[ \frac{S_{\text{ir}}^2(r^2 + f^2_{\text{eff}} - \omega^2)}{[\omega + f_{\text{eff}}]^2 + r^2] [\omega - f_{\text{eff}}]^2 + r^2]} \right] (|\bar{T}_x^2| - |\bar{T}_y^2|) \, d\omega$$

$$- \frac{\rho_0 H_M}{2} \left[ \frac{2ir\omega(f + \zeta/2)}{[\omega + f_{\text{eff}}]^2 + r^2] [\omega - f_{\text{eff}}]^2 + r^2]} \right] (\bar{T}_x \bar{T}_y - \bar{T}_x \bar{T}_y) \, d\omega. \quad (B3)$$
Taking the ensemble mean on both sides of (B3) and using (B1) yields (12a).

Substituting (11) into (3b) yields
\[
e = -\frac{\rho_0 H W S_n}{2} \int \left( \frac{r^2 + \omega^2 - f_{\text{eff}}^2}{(\omega + f_{\text{eff}})^2 + r^2} \right) \frac{d\omega}{(\omega - f_{\text{eff}})^2 + r^2} \]
\[
+ \frac{\rho_0 H W S_n}{2} \int \left( \frac{r^2 + \omega^2}{(\omega + f_{\text{eff}})^2 + r^2} \right) \frac{d\omega}{(\omega - f_{\text{eff}})^2 + r^2} \]
\[
- \frac{\rho_0 H W S_n}{2} \int \frac{2r(f + \zeta/2)\hat{T}_x\hat{T}_y^* + \text{c.c.}}{(\omega + f_{\text{eff}})^2 + r^2} \frac{d\omega}{(\omega - f_{\text{eff}})^2 + r^2},
\]
(B4)

Taking the ensemble mean on both sides of (B4) and using (B1) yields (12b).

For \( r \ll f \) (this is equivalent to \( r \ll f_{\text{eff}} \) for a small \( R_0 \)), the integrand \( r(\omega^2 + f_{\text{eff}}^2) \left[ \frac{1}{(\omega + f_{\text{eff}})^2 + r^2} \right] \) is strongly peaked around \( \omega = \pm f_{\text{eff}} \), reflecting the resonance between the wind stress and NIOs. In the limiting case, that is, \( r/f_{\text{eff}} \to 0 \),
\[
\lim_{r/f_{\text{eff}} \to 0} \left( \frac{r^2 + f_{\text{eff}}^2 + \omega^2}{(\omega + f_{\text{eff}})^2 + r^2} \right) \left[ \frac{1}{(\omega - f_{\text{eff}})^2 + r^2} \right] = \frac{\pi}{2f_{\text{eff}}} \left[ \delta \left( \frac{\omega - f_{\text{eff}}}{f_{\text{eff}}} \right) \right] + \delta \left( \frac{\omega + f_{\text{eff}}}{f_{\text{eff}}} \right),
\]
(B5)

where \( \delta \) is the Dirac \( \delta \) function. Substituting (B5) into (12a) yields
\[
E\{W_L\} = \lim_{r/f_{\text{eff}} \to 0} E\{W\}
= \frac{\pi \rho_0 H W}{2} \left\{ \sum_{\omega = \pm f_{\text{eff}}} |\hat{T}_x|^2 + |\hat{T}_y|^2 \right\}.
\]
(B6)

Similarly, we have
\[
E\{e_L\} = \lim_{r/f_{\text{eff}} \to 0} E\{e\}
= \frac{\pi \rho_0 H M}{8} \left\{ \sum_{\omega = \pm f_{\text{eff}}} |\hat{T}_x|^2 + |\hat{T}_y|^2 \right\}.
\]
(B7)

Combining (B6) and (B7) yields (14).

REFERENCES


