

Modulation of the M_2 internal tides near the Luzon Strait by westward-propagating mesoscale eddies in the Philippine Sea

Zheng Guo^a, Anzhou Cao^{b,*}, Xu Chen^c, Jinbao Song^b, Xinyu Guo^d

^a Marine Science and Technology College, Zhejiang Ocean University, Zhoushan 316022, China

^b State Key Laboratory of Ocean Sensing & Ocean College, Zhejiang University, Zhoushan 316021, China

^c Key Laboratory of Physical Oceanography, Ocean University of China, Qingdao 266100, China

^d Center for Marine Environmental Studies, Ehime University, Matsuyama 790-8577, Japan

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ABSTRACT

Mesoscale eddies are prevalent in the Philippine Sea and frequently impinge on the Kuroshio Current, affecting internal tides (ITs) originating from the Luzon Strait (LS). This process is investigated in this study using idealized experiments, which provide a controlled environment to quantify variations of the M_2 ITs as a single cyclonic eddy (CE) or anticyclonic eddy (AE) approaches the LS from the east. Our results reveal an asymmetric influence: AEs induce more pronounced changes in the M_2 tidal conversion than CEs. Variations in conversion rates are primarily controlled by changes in bottom pressure perturbations, which are linked to the product of subtidal buoyancy gradients and IT vertical velocity. Another noteworthy finding is that these mesoscale eddies influence the M_2 ITs differently in two stages: during propagation across the Philippine Sea and upon impingement with the Kuroshio Current. In the first stage, mesoscale eddies modulate the propagation of the M_2 ITs through direct interaction, inducing incoherent M_2 ITs in the Philippine Sea. The percentage of incoherent ITs exceeds 50% along the eddy paths. Advection from eddy-induced currents contribute significantly more than background stratification and relative vorticity. During the second stage, as the eddies impinge on the Kuroshio Current and move northward, variations in the M_2 IT energetics within the LS are evident. The IT incoherence increases not only in the Philippine Sea but also in the LS and South China Sea. These findings are important for understanding the IT variability and incoherence in the Philippine Sea and South China Sea.

1. Introduction

The Luzon Strait (LS) is one of the most important generation sites for internal tides (ITs) globally, owing to its distinctive double meridional ridges, pronounced stratification, and strong astronomical tides (Ramp et al., 2004; Alford et al., 2011). These energetic ITs propagate westward into the South China Sea and eastward into the Philippine Sea, extending their reaching over thousands of kilometers (Zhao, 2014; Xu et al., 2016).

Significant variability of ITs from the LS has been revealed through altimetry and mooring data (e.g. Zaron, 2017; Cao et al., 2017). This variability arises due to changes in the IT generation, which can be influenced by variations in the background circulation and stratification at the generation site and the phasing of remotely generated ITs (Jan et al., 2012; Kerry et al., 2014, 2016; Xu et al., 2021; Cao et al., 2022; 2026). Additionally, the propagation of ITs is modulated by the

background subtidal currents. In this process, mesoscale eddies play an important role, as evidenced by both observational data and theoretical and numerical models. Studies by Park and Watts (2006) and Rainville and Pinkel (2006) utilized two-dimensional ray tracing to examine the propagation of ITs through mesoscale currents in the southwestern Japan/East Sea and the Hawaiian archipelago, respectively. Chavanne et al. (2010a, 2010b) interpreted temporal variability in both amplitude and phase of semidiurnal baroclinic currents near the Hawaiian Ridge with a three-dimensional ray-tracing approach, taking into consideration the beam-like structure of the ITs close to their generation site. Near the LS, hydrographic observations reveal the changes in propagation speed and energy refraction of the mode-1 semidiurnal ITs, as well as strengthened mode-2 ITs when a pair of eddies passed (Huang et al., 2018). The dynamics behind the interaction between ITs and mesoscale eddies has been extensively investigated using idealized experiments. Dunphy and Lamb (2014) demonstrated the formation of hot

* Corresponding author.

E-mail address: caozhou@zju.edu.cn (A. Cao).

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and cold spots of energy fluxes in beam-like patterns after mode-1 ITs traversed a barotropic eddy. The interaction with a mode-1 baroclinic eddy led to the energy transfer from mode-1 ITs to modes two and higher. Wang and Legg (2023) further examined the dissipation of ITs within mesoscale eddies, finding that energy transfer from low to high modes in the baroclinic eddy structure and the trapping of those high modes by the eddy velocity field are essential for enhanced dissipation. These findings highlight a more universal dynamical phenomenon within the broader context of wave-current interactions: mesoscale eddies can substantially alter effective wave properties, propagation pathways, and deep-ocean energy transmission through vorticity-induced modifications of the background environment (Kunze, 1985; Whalen et al., 2018; Ma et al., 2022). For instance, based on moored current observations, Ma et al. (2022) demonstrated that typhoon-induced near-inertial waves (NIWs) showed a significant blue/red shift as the observation site was located outside/within an intense anticyclonic eddy. Such background vorticity and its spatial gradients modulate the wavenumber vectors, thereby controlling the vertical group velocity and deep-sea energy penetration of NIWs. Although these prior observational studies primarily focused on NIWs, the underlying dynamical processes are closely related to the mechanisms governing IT-eddy interactions.

Mesoscale eddies are prevalent near the LS. To the west of the LS, the northern South China Sea experiences mesoscale eddies 35%–60% of the time (Chen et al., 2011). Numerical simulations by Guo et al. (2023) demonstrated the refraction of the M_2 ITs by anticyclonic eddies (AEs) and cyclonic eddies (CEs) in the South China Sea. Fan et al. (2024) focused on variations in the energetics of ITs during their interactions with eddies. East of the LS, in the Philippine Sea, a well-defined zonal band of high eddy kinetic energy exists between 18°N and 26°N, to which mesoscale eddies contribute (Qiu, 1999; Waterman et al., 2011). Most of these mesoscale eddies are generated between the Subtropical Countercurrent and North Equatorial Current. They propagate westward at a mean translation speed of $O(0.1 \text{ m/s})$, with a mean radius of $\sim 130 \text{ km}$ for CEs and $\sim 120 \text{ km}$ for AEs (Yang et al., 2013a; Cheng et al., 2014). Their trajectories spatially overlap with the primary path of ITs originating from the LS. The inevitable interaction between mesoscale eddies and ITs modulates propagation of ITs. Moreover, these eddies frequently intrude into the Kuroshio Current regime (Movie S1). Analysis of 19.5 years of satellite altimeter data reveals that 315 eddies approached the Luzon-Taiwan eastern coast, predominantly concentrating at latitudes 18°N–19°N and 22°N–23°N (Cheng et al., 2017). Chang et al. (2015) showed that AEs enhanced the strength of the Kuroshio Current, while CEs tended to weaken it. The eddy-Kuroshio Current interactions east of Taiwan have been found to have remote effects in the LS, i.e., Kuroshio Current intrusion (Jan et al., 2017). Changes in the Kuroshio Current within the LS affect generation and propagation of ITs (Jan et al., 2012; Xu et al., 2021; Cao et al., 2026). Such changes give rise to incoherent ITs, which remain within the tidal frequency band but exhibit irregular and unpredictable amplitude and phase (Cao et al., 2022). These studies suggest that mesoscale eddies in the Philippine Sea could modulate the propagation of ITs and affect their energetics through interactions with the Kuroshio Current, subsequently leading to an increase in IT incoherence. This is a topic that has yet to be thoroughly investigated.

The primary goal of this study is to investigate the influence of a mesoscale eddy on the M_2 ITs as it propagates through the Philippine Sea and impinges on the Kuroshio Current. To achieve this, we conduct a series of idealized numerical experiments contextualized to the LS, simultaneously simulating ITs and subtidal motions including the Kuroshio Current and mesoscale eddies produced using the Rankine Vortex winds (Holland, 1980). Section 2 describes the model settings and methods used for analyzing the results. The results from numerical experiments are presented in Section 3, followed by a discussion in Section 4. Section 5 summarizes the findings of this study.

2. Methodology

2.1. Model setup

The investigation into the influence of mesoscale eddies on the M_2 ITs is based on the Coastal and Regional Ocean COMMunity (CROCO) model (<https://www.croco-ocean.org>). Built on the regional ocean modelling system from Institut de Recherche pour le Développement (ROMS AGRIF), the CROCO has been successfully applied to various scenarios (Marchesiello et al., 2019; Renault et al., 2017; Cao et al., 2023; Guo et al., 2023).

The model domain spans from 115.5°E to 131.5°E in longitude and from 16°N to 26°N in latitude, with the LS situated in the center. The bathymetry is extracted from the 1-min gridded elevations/bathymetry for the world (ETOPO1) data (Amante and Eakins, 2009). Horizontally, the model employs a grid spacing of $1/20^\circ$ (approximately 5 km), while vertically, it uses 25 unevenly distributed sigma layers, providing finer resolution near the surface and bottom (Table S1). To simulate the M_2 tides, surface elevations and barotropic currents of the M_2 constitute are sourced from the Oregon State University Ocean Topography Experiment TOPEX/Poseidon Global Inverse Solution (TPX07; Egbert and Erofeeva, 2002) and introduced at the four open boundaries. No astronomical body force is applied in the model interior. Additionally, a 0.5° wide sponge layer is implemented to absorb baroclinic energy and prevent artificial reflection of ITs at the open boundaries.

To represent the background state of the Kuroshio Current, the model is initialized using a 20-year (1996 to 2015) mean ocean state, obtained by averaging monthly data from the Simple Ocean Data Assimilation (SODA) dataset (Carton et al., 2018). The model runs for 30 days to ensure that the background field achieves a steady state.

Eddies are generated in the ocean using wind stresses based on the Rankine Vortex theory (Holland, 1980; Jan et al., 2017):

$$\frac{V_w^2}{r} + fV_w = \frac{1}{\rho_a} \frac{\partial p}{\partial r}, \quad (1)$$

where V_w is the orbital wind speed, r is the distance to the vortex wind center, f is the Coriolis parameter, ρ_a is the air density, and p is the pressure which is calculated as:

$$p(r) = P_c + (P_n - P_c)e^{-\frac{R_{max}}{r}}, \quad (2)$$

where P_c is the center pressure of the vortex wind field, R_{max} is the radius of maximum wind, and P_n is the environmental pressure far from center of vortex. The central pressure drop ($P_n - P_c$), radius of maximum wind, and duration of the wind field determine the size and strength of the eddy. Following Jan et al. (2017), a three-day anticyclonic/cyclonic wind field is applied to the spin-up model in the AE/CE cases, with $R_{max} = 100 \text{ km}$ and $P_n = 1010 \text{ hPa}$. After this period, the wind forcing is removed to allow the eddies to propagate freely, and the M_2 tidal forcing is introduced at the open boundaries. The model is forced solely by tides and integrated for 70 days, with results from the last 60 days used for analysis.

Six experiments with different settings of the initial center location, strength, and polarity of the eddy were performed (Table 1) to investigate the influence of different mesoscale eddies on ITs. The AE1 and CE1 cases represent an AE and a CE, respectively, propagating westward in the Philippine Sea and impinging on the Kuroshio Current to the east of Taiwan Island. In AE2 and CE2, the Rankine Vortex winds for generating

Table 1
Parameters used to produce the Rankine vortex winds for each case.

Case	AE1	AE2	AE3	CE1	CE2	CE3
P_c (hPa)	980	965	980	980	965	980
Initial eddy center	128°E 22.5°N	128°E 22.5°N	128°E 22°N	127.5°E 20°N	127.5°E 20°N	127.5°E 19.5°N

the eddies are increased, and the modeled eddies are stronger than those in AE1 and CE1. AE3 and CE3 are similar to AE1 and CE1, respectively, but the initial eddy center of the two eddies is 0.5° further south. For comparison, a No Eddy (NE) case is also conducted, where the tidal forcing is exerted on the spin-up model. This case also runs 70 days.

2.2. Baroclinic tidal energy equations

The momentum, buoyancy, and continuity equations governing inviscid ocean motion are (Kelly and Lermusiaux, 2016; Cao et al., 2026):

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - f v = -\frac{1}{\rho} \frac{\partial p}{\partial x}, \quad (3)$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + f u = -\frac{1}{\rho} \frac{\partial p}{\partial y}, \quad (4)$$

$$0 = -\frac{1}{\rho} \frac{\partial p}{\partial z} + b, \quad (5)$$

$$\frac{\partial b}{\partial t} + u \frac{\partial b}{\partial x} + v \frac{\partial b}{\partial y} + w \frac{\partial b}{\partial z} = 0, \quad (6)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \quad (7)$$

where u , v , and w are the zonal, meridional, and vertical velocities, respectively; t is the time; ρ is the water density; p is the pressure; and $b = -g(\rho - \rho_0)/\rho_0$ is the buoyancy with g denoting the gravity acceleration and ρ_0 denoting the reference density.

In a system where ITs coexist with subtidal motions, such as meso-scale eddies and the Kuroshio Current, the density can be decomposed as:

$$\rho(x, y, z, t) = \rho_0 + \bar{\rho}(x, y, z, t) + \rho'(x, y, z, t), \quad (8)$$

where the overbar denotes the slowly varying component associated with subtidal motions, and the prime denotes the components associated with ITs. According to the hydrostatic approximation, the pressure can be expressed as:

$$p(x, y, z, t) = p_0(z) + \bar{p}(x, y, z, t) + p'(x, y, z, t). \quad (9)$$

The velocity and buoyancy are similarly partitioned into subtidal and IT components. Let $\chi(x, y, z, t)$ represent u , v , w , or b :

$$\chi(x, y, z, t) = \bar{\chi}(x, y, z, t) + \chi'(x, y, z, t). \quad (10)$$

Following Kelly and Lermusiaux (2016), substituting Equations (8–10) into Equations (3–7) yields governing equations for the subtidal motions and ITs, respectively. The equations for the ITs are displayed as follows:

$$\frac{\partial u'}{\partial t} + \bar{u} \frac{\partial u'}{\partial x} + \bar{v} \frac{\partial u'}{\partial y} + \bar{w} \frac{\partial u'}{\partial z} + u' \frac{\partial \bar{u}}{\partial x} + v' \frac{\partial \bar{u}}{\partial y} + w' \frac{\partial \bar{u}}{\partial z} - f v' = -\frac{1}{\rho_0} \frac{\partial p'}{\partial x}, \quad (11)$$

$$\frac{\partial v'}{\partial t} + \bar{u} \frac{\partial v'}{\partial x} + \bar{v} \frac{\partial v'}{\partial y} + \bar{w} \frac{\partial v'}{\partial z} + u' \frac{\partial \bar{v}}{\partial x} + v' \frac{\partial \bar{v}}{\partial y} + w' \frac{\partial \bar{v}}{\partial z} + f u' = -\frac{1}{\rho_0} \frac{\partial p'}{\partial y}, \quad (12)$$

$$0 = -\frac{1}{\rho_0} \frac{\partial p'}{\partial z} + b', \quad (13)$$

$$\frac{\partial b'}{\partial t} + \bar{u} \frac{\partial b'}{\partial x} + \bar{v} \frac{\partial b'}{\partial y} + \bar{w} \frac{\partial b'}{\partial z} + u' \frac{\partial \bar{b}}{\partial x} + v' \frac{\partial \bar{b}}{\partial y} + w' \frac{\partial \bar{b}}{\partial z} = 0, \quad (14)$$

$$\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} = 0. \quad (15)$$

Multiplying Equation (11) by $\rho_0 u'$, Equation (12) by $\rho_0 v'$, and Equation

(14) by $\rho_0 b'/N^2$ ($N = \sqrt{g \bar{b}/\bar{z}}$), summing the results, and integrating in the vertical direction, we obtain the IT energy equation

$$\frac{\partial E}{\partial t} = -\nabla_H \cdot \mathbf{F} + A + P_S + P_B, \quad (16)$$

where

$$E = \int_{-h}^0 \frac{1}{2} \rho_0 \left(u'^2 + v'^2 + \frac{b'^2}{N^2} \right) dz, \quad (17)$$

$$\mathbf{F} = \int_{-h}^0 p' \mathbf{u}'_H dz \quad (18)$$

$$A = \int_{-h}^0 -\rho_0 \left[\mathbf{u}' \cdot (\bar{\mathbf{u}} \cdot \nabla) \mathbf{u}' + \frac{b'}{N^2} (\bar{\mathbf{u}} \cdot \nabla b') \right] dz, \quad (19)$$

$$P_S = \int_{-h}^0 -\rho_0 \mathbf{u}' \cdot (\mathbf{u}' \cdot \nabla) \bar{\mathbf{u}} dz, \quad (20)$$

$$P_B = \int_{-h}^0 -\rho_0 \left(\frac{b'}{N^2} (\mathbf{u}' \cdot \nabla \bar{b}) \right) dz. \quad (21)$$

Here, E is total IT energy containing both kinetic energy and available potential energy; $\mathbf{F}$ is IT energy flux; A is IT energy advection caused by the subtidal motions; P_S and P_B are IT energy production terms associated with the shear and stratification of the subtidal motions, respectively (Kelly and Lermusiaux, 2016); h is the total depth of water; and $\mathbf{u} = (u, v, w)$ is velocity vector. Following Nash et al. (2005), the pressure perturbation caused by ITs at a certain depth z is:

$$p' = - \int_z^0 \rho_0 b' d\bar{z} - p_{surf}, \quad (22)$$

where surface pressure p_{surf} can be inferred from the baroclinicity condition that the depth-averaged pressure perturbation must vanish.

The IT energy budget is completed by considering the generation (C) and dissipation (D) terms (Cao et al., 2026):

$$\frac{\partial E}{\partial t} = C - \nabla_H \cdot \mathbf{F} + A + P_S + P_B - D. \quad (23)$$

Here, the conversion rate C is calculated by multiplying the pressure perturbation p' by the vertical barotropic tidal velocity w_{bt} at the sea floor:

$$C = p'(-h) w_{bt}(-h). \quad (24)$$

The dissipation term is calculated as the budget residual (Kerry et al., 2014; Wang et al., 2018).

During data processing, the baroclinic velocity is obtained by subtracting barotropic velocity from the total velocity, both of which are directly output by the model. Subsequently, the M_2 IT components are extracted by applying a temporal Butterworth bandpass filter ([1.73 2.13] cpd) to the baroclinic velocity and density fields, while subtidal components are extracted by applying a 0.5 cpd lowpass filter to the total velocity and density fields.

2.3. IT incoherence

Under the influence of background field, energy of the M_2 ITs transfers from the coherent to the incoherent components. The coherent component is defined with a predictable amplitude and phase lag, whereas the incoherent component lies within the tidal frequency band but does not have a predictable amplitude and phase lag (Arbic, 2022). Therefore, the ITs are expressed as:

$$\mathbf{u}' = \mathbf{u}'_c + \mathbf{u}'_{ic} \quad (25)$$

The total M_2 ITs $\mathbf{u}'$ are extracted from the baroclinic currents with a fourth-order bandpass Butterworth filter with cutoff frequency of [1.73, 2.13] cpd (Zhao et al., 2010; Cao et al., 2022). The coherent component $\mathbf{u}'_c$ is calculated through the traditional harmonic analysis. Thereafter, the incoherence of the M_2 ITs can be assessed in terms of horizontal kinetic energy (HKE):

$$\gamma = \frac{\langle \text{HKE}_{ic} \rangle}{\langle \text{HKE} \rangle} = \frac{\langle \frac{1}{2} \int_{-h}^0 \rho |\mathbf{u}'_{ic}|^2 dz \rangle}{\langle \frac{1}{2} \int_{-h}^0 \rho |\mathbf{u}'|^2 dz \rangle} \quad (26)$$

where $\langle \rangle$ denotes time averaging.

2.4. Relative changes in mode-1 IT phase speed

The relative changes of phase speed can be used to assess contributions of processes leading to the incoherence of ITs (Zaron and Egbert, 2014; Buijsman et al., 2017; Savage et al., 2020; Cao et al., 2022). Considering that most of the IT energy is transported away from the LS by the first mode (Zhao, 2014), the formula for relative changes in mode-1 IT phase speed is used. Contributions of different processes are assessed following (Zaron and Egbert, 2014; Savage et al., 2020):

$$\frac{\delta c_p}{c_p} \approx \frac{1}{2} \frac{\delta c_1^2}{c_1^2} + \sqrt{1 - \frac{f^2}{\omega^2}} \frac{\delta |\bar{\mathbf{u}}_{11}|}{c_1} + \frac{1}{2} \frac{f^2}{\omega^2} \frac{\delta \bar{\zeta}_{11}}{f}. \quad (27)$$

In the equation, c_p is the mode-1 phase speed, δ denotes standard deviation, ω is the frequency of the M_2 ITs, and c_1 is the mode-1 eigenspeed for the eigenvalue problem with respect to the eigenfunction Φ_n (n is the mode number):

$$\frac{d^2 \Phi_n}{dz^2} + \frac{N^2}{c_n^2} \Phi_n = 0, \quad (28)$$

subject to boundary conditions $\Phi_n(0) = \Phi_n(-h) = 0$.

$$\bar{\mathbf{u}}_{11} = \int_{-h}^0 \bar{\mathbf{u}} \Phi_1^2 dz \text{ and } \bar{\zeta}_{11} = \frac{\partial \bar{\mathbf{u}}_{11}}{\partial x} - \frac{\partial \bar{\mathbf{u}}_{11}}{\partial y}. \quad (29)$$

The three terms on the righthand side of Equation (27) represent contributions of background stratification, advection by the subtidal current, and refraction due to background relative vorticity. The Matlab script provided by Samuel M. Kelly (<https://www.d.umn.edu/~smkelly/software.html>) is used to calculate the mode-1 eigenspeed and eigenfunction.

3. Results

3.1. Model validation

The background currents in the NE case, depicted in Fig. 1a, illustrate the climatology of the Kuroshio Current (Rudnick et al., 2011). Beginning east of Luzon Island, the Kuroshio Current flows northward across the LS with a core velocity of about 1 m/s, closely follows the eastern coast of Taiwan Island, and then shifts northeastward. This provides a stable baseline against which the influence of westward-propagating mesoscale eddies in the Philippine Sea are examined.

The IT energetics in the LS (120°E–122.5°E, 18.5°N–22°N) are examined using Equations (16–24) for the NE case to further validate the model (Fig. 1b). The IT energy budget is primarily balanced by the tidal conversion, energy flux divergence and local dissipation. The temporal mean tidal conversion of ITs integrated within the LS is 19.6 GW, which dominates the energy budget. A significant portion of this energy propagates out of the LS, resulting in an energy flux divergence of 10.4 GW. Of this, 4.5 GW propagating westward into the South China Sea and 5.2 GW propagating eastward into the Philippine Sea. The westward energy fluxes form a strong northern branch and a weak southern branch. The eastward-propagating ITs form a single beam that interferes with ITs from the Ryukyu Island chain, producing a complex inhomogeneous energy flux field. Within the LS, local dissipation of ITs accounts

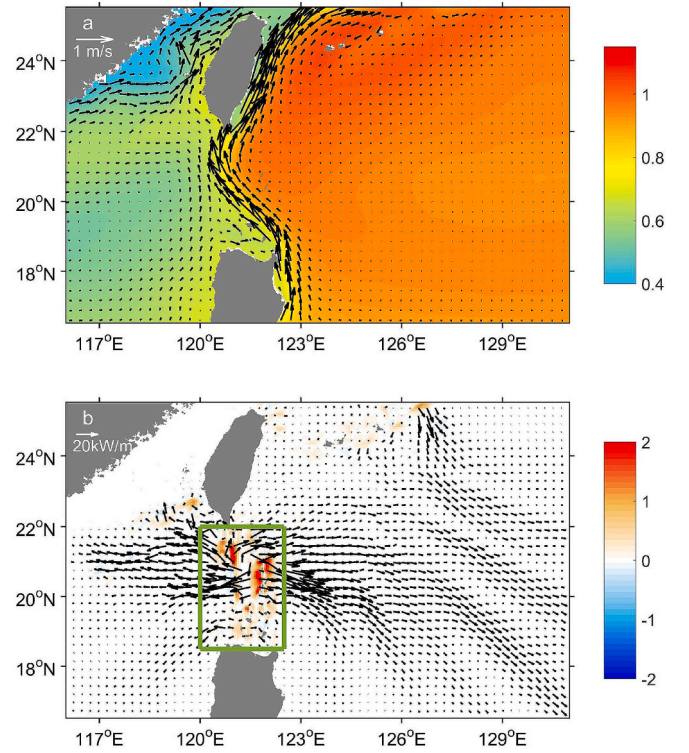


Fig. 1. (a) Model-simulated sea surface height (shading, unit: m) and currents (quivers) for the background ocean state. (b) The energy conversion rate (shading, unit: W/m²) and depth-integrated energy flux (quivers) of the M_2 ITs. The LS is denoted by the dark green box.

for 9.6 GW. These results are consistent with previous research (Kerry et al., 2014; Cao et al., 2022). Among the remaining terms, the buoyancy production, shear production and advection associated with subtidal motions, the Kuroshio Current in this case, are at least an order of magnitude smaller, with a combined contribution of only 0.39 GW. The temporal derivative of energy is close to 0.

3.2. Eddy evolution

Eddy evolution is examined by extracting subtidal motions with a fourth-order lowpass Butterworth filter. The cutoff frequency is set as 0.5 cpd (Zhang et al., 2016; Cao et al., 2022). Fig. 2a–2e show snapshots of the simulated sea level and surface velocity as the AE1 propagated through the Philippine Sea and impinging on the Kuroshio Current. Initially, the center of the AE1 was located at 22.5°N, 128°E and its maximum rotation speed was near 1.4 m/s. As the AE moved westward, it exhibited a slight equatorward deflection, which aligned with observations (Chelton et al. 2007; Cheng et al., 2017) and could be explained by the β -effect theory (Cushman-Roisin et al. 1990). On Day 40, the AE1, with a maximum velocity of approximately 1 m/s, intruded into the Kuroshio Current regime near 21.5°N. During its interaction with the Kuroshio Current, the AE1 elongated in the direction parallel to the axis of the Kuroshio Current and traveled northeastward along the east coast of the Taiwan Island. The evolution of the simulated CE1 is presented in Fig. 2g–2k. Initially centered at 20°N, 127.5°E, the CE1 moved westward with a slight poleward deflection. The edges of the CE1 began to impinge on the flank of the Kuroshio Current near 21°N. Afterwards, the CE1 was deformed to an elliptic shape with its major axis oriented northwest-southeast as it weakened. These evolutions closely resembled the simulations in Jan et al. (2017). The trajectories of AE2 and CE2 closely resembled those of AE1 and CE1, respectively, except during the last 10 days (Fig. 2f and 3l). During this interval, AE2 remained near 23°N while CE2 moved further north compared with CE1. The AE3

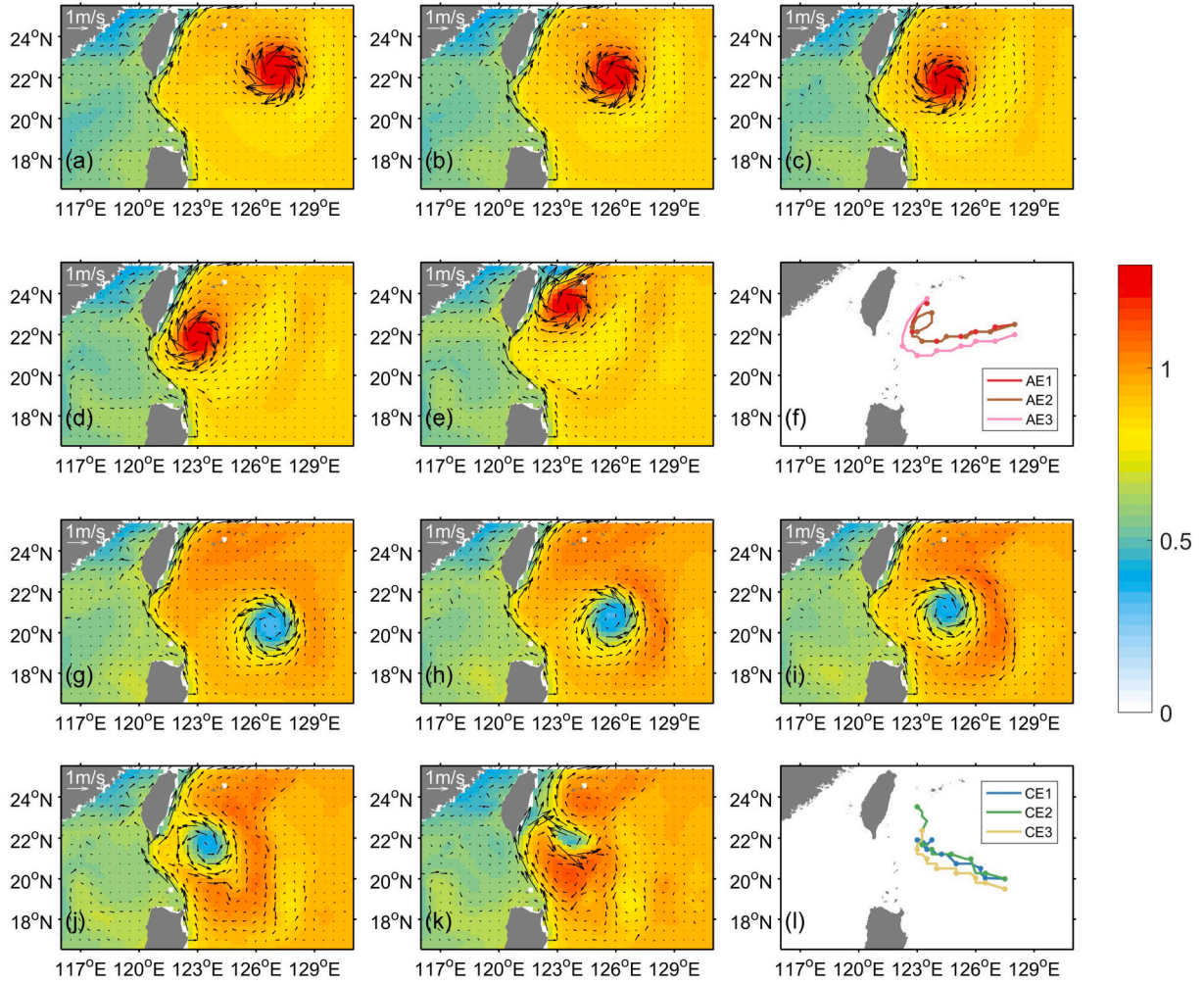


Fig. 2. Results of the simulated AE1: Sea level (shading, unit: m) and velocity (quivers) on days (a) 10, (b) 25, (c) 40, (d) 55, and (e) 70. (f) The trajectories of AEs in three cases, with dots indicating eddy centers at 10-day intervals. Panels (g-l) show corresponding CE results.

(CE3) exhibited a trajectory parallel to that of AE1 (CE1) but displaced 0.5° southward. However, by the end of the simulation, trajectories of the two AEs (CEs) nearly converged. Snapshots illustrating the evolution of these eddies are provided in the supporting information (Figs. S1-S4).

3.3. IT variability

3.3.1. IT energy budget within the LS

To investigate the influence of mesoscale eddies on the M_2 ITs, we examine IT energetics in the LS. We conduct calculation every 3 days (averaged over the first 50 h, i.e., four tidal cycles) for the results of the period from days 11 to 70.

The IT energetics in the NE case remain essentially stable. By contrast, the eddies induce temporal variability in the IT energy budget, particularly during the second half of the analysis period (Fig. 3). Attention is first paid to the dominant term, tidal conversion. Variations in conversion rates are generally more pronounced in AE cases than in CE cases, highlighting asymmetric influence of AEs and CEs on tidal conversion within the LS (Fig. 3a). In the AE1 case, tidal conversion peaks at 21.6 GW on day 55, representing a 2.1 GW (10.8%) increase relative to the NE case at the same time. Peak values in the AE2 and AE3 cases are even higher, corresponding to increases of 3.2 GW and 3.3 GW, respectively. These results indicate the role of eddy intensity and trajectory in modulating IT generation. In addition, the variation in the AE2 case precedes that in the AE1 case, which is attributable to the larger size of AE2. In contrast, variations in IT generation under the CEs are less

pronounced, with the largest change occurring in CE2 case, which is merely a 1.1 GW decrease compared to the NE case. The dynamics associated with more significant influence of AEs will be discussed in Section 4.1. The variations in energy flux divergence under the influence of AEs and CEs are generally symmetric, with increases observed in AE cases and decreases in CE cases (Fig. 3b). The largest variation in each case is approximately 1 GW, amounting to a 10% change relative to the NE case. Changes in dissipation show a pattern similar to that of generation: dissipation in the AE cases increases significantly, whereas in the CE cases it fluctuates slightly around the values of the NE case (Fig. 3c).

Next, we examine the terms associated with subtidal motions. The buoyancy production in the CE cases exhibits minor fluctuations around the NE-case values of ~ 0.9 GW. In contrast, the buoyancy production in AE cases changes more significantly. The largest changes occur in the AE2 case: a maximum increase of 0.4 GW and a maximum decrease of 0.3 GW relative to the NE case. Buoyancy production remains positive throughout the simulation period across all cases, indicating a consistent transfer of energy from subtidal motions to ITs. Shear production in the NE case is approximately -0.1 GW, indicating energy transfer from ITs to subtidal motions. This distributed energy transfer is driven by macroscopic geostrophic shear, which stands as an alternative mechanism to the localized critical-layer momentum deposition identified by Shakespeare (2023). The impingement of CEs onto the Kuroshio Current increases IT energy loss to subtidal motions, with the maximum loss of 0.7 GW observed in the CE2 case. The influence of AEs is more complex.

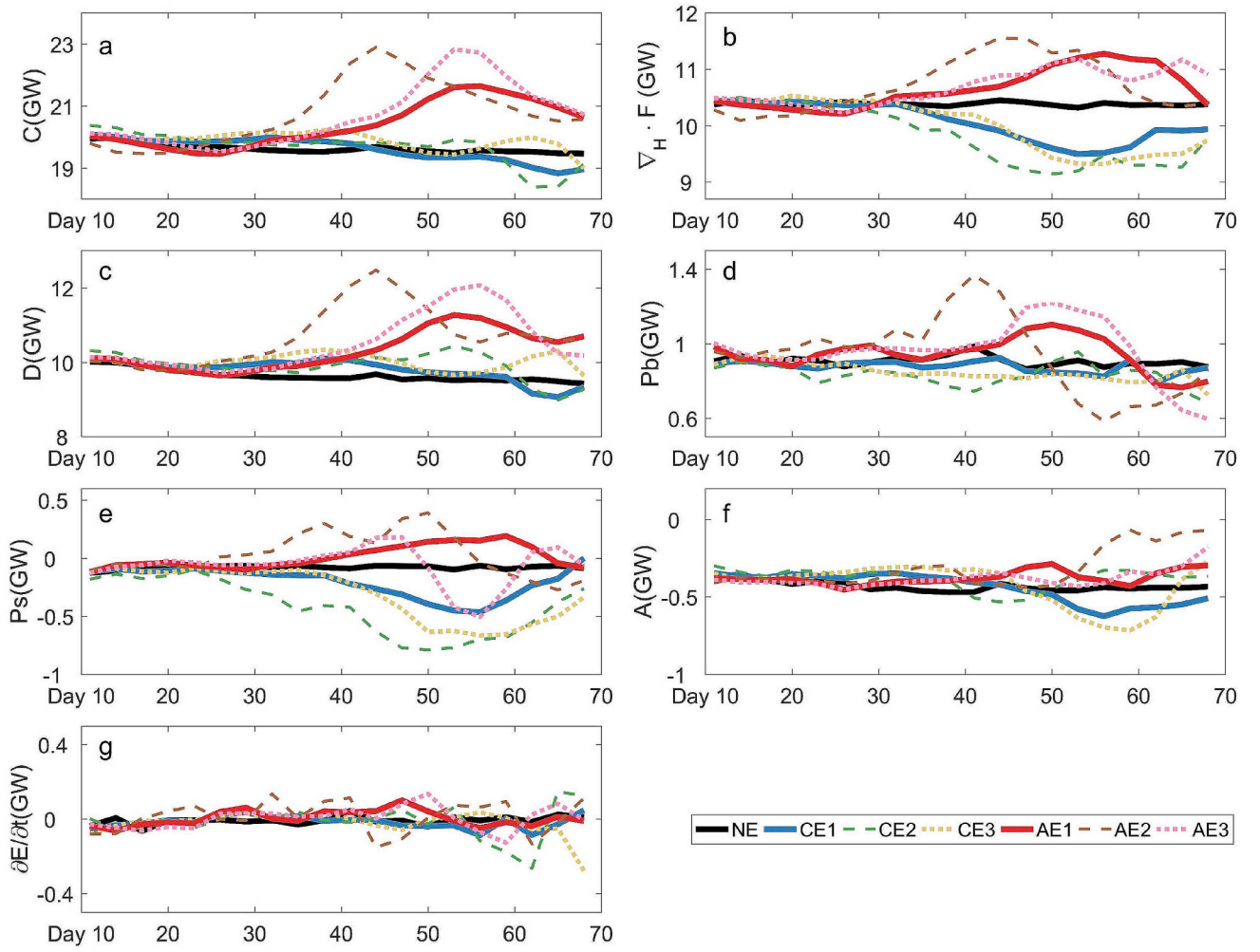


Fig. 3. The (a) tidal conversion C , (b) energy flux divergence $\nabla_H \cdot \mathbf{F}$, (c) dissipation D , (d) buoyancy production P_b , (e) shear production P_s , (f) advection A , and (g) partial derivative of energy $\partial E / \partial t$ of the M₂ ITs calculated every 3 days for the period from day 11 to 70.

The AEs alter the magnitude of IT energy loss and even reverse the direction of transfer, i.e., channeling energy from subtidal motions into ITs. Advection terms change slightly under the influence of eddies, except during the final days of the simulation in some of the cases. Details regarding the IT energy budget across all cases are summarized in Table S2.

3.3.2. IT propagation

Spatial distribution of the time-varying baroclinic energy flux is further examined. The comparison between the AE1/CE1 and NE cases is presented below as an example, with additional cases provided in the supporting information (Figs. S5 and S6). Fig. 4a-e display the difference in the M₂ baroclinic energy fluxes between the AE1 case and the NE case. During days 10 to 40 before the westward-moving AE enters into the Kuroshio Current regime, significant changes in the baroclinic

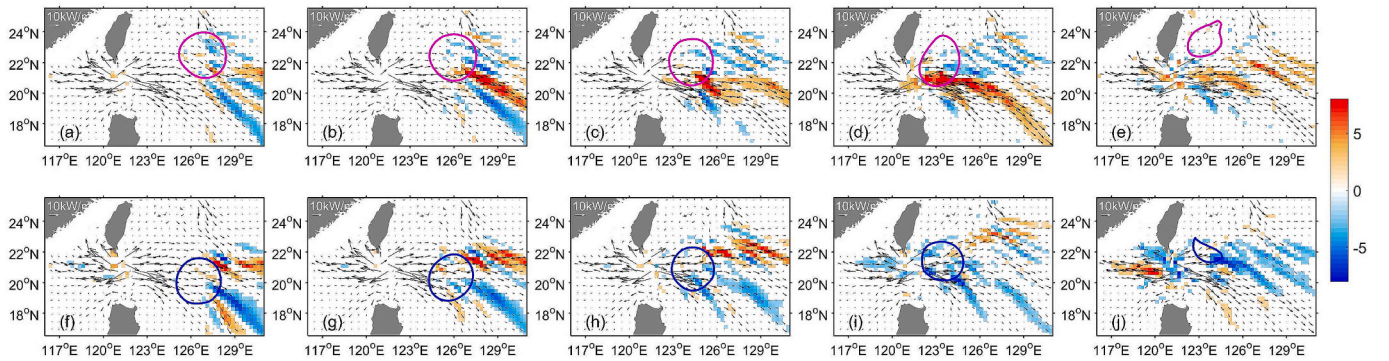


Fig. 4. Differences in the M₂ baroclinic energy fluxes between the NE and AE1 case (shading, unit: kW/m) on days (a) 10, (b) 25, (c) 40, (d) 55, and (e) 70. Differences with magnitude less than 2 kW/m are not shown for clarity. Vectors denote the energy fluxes in the AE1 case. Panels (f-j) show the corresponding differences between the NE and CE1 cases.

energy fluxes appear both within the eddy and to its east. For instance, on day 10, the variation in the magnitude of M_2 baroclinic energy flux exceeds 2 kW/m in many areas, which represents a relative change of over 26%. The CE also causes changes in the M_2 baroclinic energy flux (Fig. 4f-4j). While it is difficult to directly compare the results of the AE1 and CE1 cases due to their different paths, some insights can be drawn based on the patterns on day 40 when the AE1 and CE1 move to similar locations. In the AE1 case, the ITs weaken in the northern part of the AE's wake but strengthen in the southern part. Conversely, in the CE1 case, there is an increase in the north but a decrease in the south. This difference in the effects of AE1 and CE1 is consistent with previous ray tracing analyses (Guo et al., 2023; Wang and Legg, 2023). In the latter half of the period, variations in IT energy flux emerge not only in the Philippine Sea, but also in the LS and South China Sea, which can be attributed to changes in energetics within the LS. In addition, these flux variations exhibit site dependence. For example, in the CE1 case on day 70, the westward beam displays inhomogeneity: its northern portion weakens while the southern portion strengthens.

A common feature in the differences in IT energy flux between the eddy cases and NE case is that positive and negative values form alternative beams. This pattern resembles the beam-shaped hot and cold spots of energy flux found in the ideal experiments where mode-one ITs are refracted as they pass through eddies (Dunphy and Lamb, 2014; Wang and Legg, 2023). In our simulation, the M_2 ITs propagating away from the LS are comprised of low-mode components (Zhao, 2014; Alford et al., 2015). Their interaction with eddies will induce energy refraction similar to the ideal experiments in Dunphy and Lamb (2014). However, in our setting, the pattern is more complex. This complexity may be due to the M_2 ITs branch originating from the Ryukyu Island chain and propagating southeast. They could interfere with the eastward branch from the LS, forming a geographically inhomogeneous baroclinic energy

flux field (Wang et al., 2018). The modulation of mesoscale eddies further enhances this geography variation.

3.3.3. IT incoherence

The variability in generation and propagation of ITs increases their incoherence (Dunphy and Lamb, 2014; Kerry et al., 2016; Guo et al., 2023). The incoherence of ITs is quantified according to Equation (26). For the NE case, the incoherence is evaluated over the final 60 days of the simulation. For the eddy cases, the calculation is divided into two periods –the propagation of eddies across the Philippine Sea (day 10–40), and their subsequent impingement on the Kuroshio Current (day 40–70). For the AE2 case, however, the dividing point is day 35 owe to its earlier impingement on the Kuroshio Current.

In the NE case, the M_2 ITs display a high degree of coherence, which is unsurprising given the relatively steady background stratification and currents in this case (Fig. 5e). In the AE1 case, the M_2 ITs display significant incoherence in the Philippine Sea during the first stage (Fig. 5a). Specifically, the percentage of incoherent ITs exceeds 50% in the wake and along the path of the AE. The mean incoherence in the Philippine Sea (122.5°E–131°E, 16.5°N–25.5°N) is 19.7%. However, the ITs in the LS and South China Sea (116°E–120°E, 16.5°N–25.5°N) exhibit minimal variation, with mean incoherence values of merely 1.7% and 2.7%, respectively. During the second stage, high incoherence values appear near the Taiwan Island where the AE impinges upon the Kuroshio Current (Fig. 5b). Unlike the first stage, the ITs in the LS and South China Sea gain incoherence, reaching 7.1% and 5.0%, respectively. The CE also increases the incoherence of ITs. The temporal variation and spatial distribution of incoherent ITs resemble those observed in the AE1 case (Fig. 5c and 5d). The percentage of incoherent ITs in the LS and South China Sea in the second stage is even higher, reaching 11.6% and 8.8%, respectively. Similar incoherence patterns in the other cases are shown

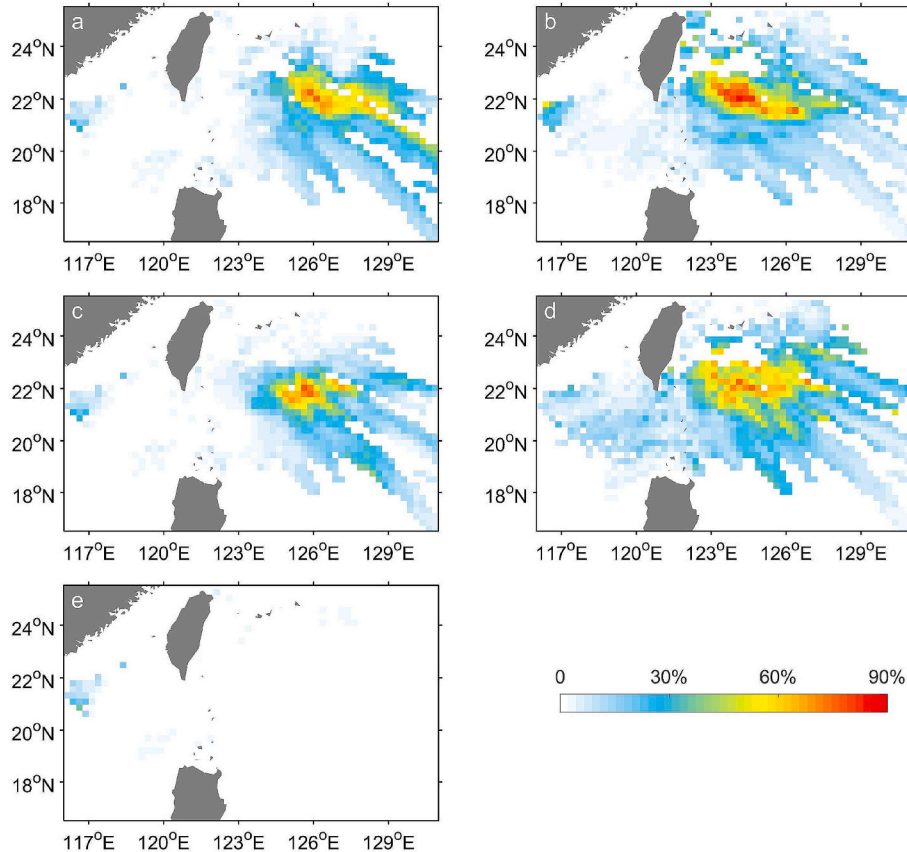


Fig. 5. Incoherence of the M_2 IT averaged over the periods (a,c) from day 10 to 40 and (b,d) from day 40 to 70 for the (a,b) AE1 and (c,d) CE1 cases. In the (e) NE case, the incoherence is calculated over the final 60 days of the simulation. Note that only areas with HKE greater than 1 kJ/m² are shown for clarity.

in Figs. S7 and S8. The time-averaged incoherence for the South China Sea, LS and Philippine Sea in each case is presented in Table S3.

4. Discussion

4.1. Asymmetric influence of AEs and CE on tidal conversion

In examination of the results, the first key finding is that AEs substantially enhance the tidal conversion, while CEs exert only minor influence. To investigate the asymmetric influence of AEs and CEs, we compare the amplitudes of bottom pressure perturbation [$P'(-h)$] and vertical barotropic tidal velocity [$W_{bt}(-h)$], and their phase difference (θ), which are commonly used to calculate the conversion rate following Zilberman et al. (2011):

$$<C> = 0.5P'(-h)W_{bt}(-h)\cos(\theta). \quad (30)$$

For illustration, results are shown for the 50-hour period beginning on Day 55 in the NE, AE1, and CE1 cases, when the difference in conversion rate between NE and AE1 is maximal (Fig. 6). The spatial distributions of P' closely resembles with that of the conversion rate. Its difference between the NE and AE1 cases mirrors the difference in conversion rate. This indicates that the bottom pressure perturbation exerts a stronger influence than either bottom vertical barotropic tidal velocity or their phase difference in determining the conversion rate.

Based on Equation (30), we further derive the variation in conversion rate:

$$\begin{aligned} \Delta C &= 0.5(P' + \Delta P')(W_{bt} + \Delta W_{bt})\cos(\theta + \Delta\theta) - 0.5P'W_{bt}\cos\theta \\ &= 0.5[(W_{bt}P'\cos\theta\cos\Delta\theta - W_{bt}P'\cos\theta) + W_{bt}\Delta P'\cos\theta\cos\Delta\theta \\ &\quad + P'\Delta W_{bt}\cos\theta\cos\Delta\theta + \Delta W_{bt}\Delta P'\cos\theta\cos\Delta\theta - W_{bt}P'\sin\theta\sin\Delta\theta \\ &\quad - W_{bt}\Delta P'\sin\theta\sin\Delta\theta - P'\Delta W_{bt}\sin\theta\sin\Delta\theta - \Delta W_{bt}\Delta P'\sin\theta\sin\Delta\theta] \end{aligned} \quad (31)$$

The spatial patterns of the eight terms on the right-hand side of Equation (31) for the AE1 and CE1 cases are presented in Figs. 7 and 8, respectively. In the AE1 case, the Term 2 ($W_{bt}\Delta P'\cos\theta\cos\Delta\theta$) exhibits close resemblance to the distribution of ΔC (Fig. 6e) and its magnitude is substantially larger than that of the other terms. While the localized variation of Term 3 ($0.5\Delta W_{bt}P'\cos\theta\cos\Delta\theta$) is non-negligible at the ΔC hotspot, its domain-integrated value is only 0.2 GW, which is one order of magnitude smaller than that of Term 2 (2.3 GW). This confirms the dominant role of bottom pressure perturbation in modulating IT generation in the AE1 case. However, in the CE1 case, the changes in most terms are minor compared with the AE1 case. Among them, Terms 2 and 5 exhibit relatively larger integrated values, yet they reach only -0.11 GW and 0.15 GW, respectively. These two terms largely compensate each other, leading to a suppressed net change in the total conversion rate for the CE1 case. Further inspection on equation (31) reveals that all terms depend on the perturbation of phase difference $\Delta\theta$. Under a simple assumption of small phase perturbations as is the case here ($\Delta\theta \approx 0$, hence $\cos\Delta\theta \approx 1$ and $\sin\Delta\theta \approx \Delta\theta$), we find that Term 1 vanishes; Terms 2–4 are driven by amplitude perturbations (Term 2 by $\Delta P'$, Term 3 by ΔW_{bt} , and Term 4 by changes in both amplitudes); Term 5 is driven by perturbations in the phase difference $\Delta\theta$; and the remaining Terms 6–8 represent higher-order cross-interactions.

We next examine the processes influencing changes in pressure perturbation. According to Equation (22), the changes in bottom

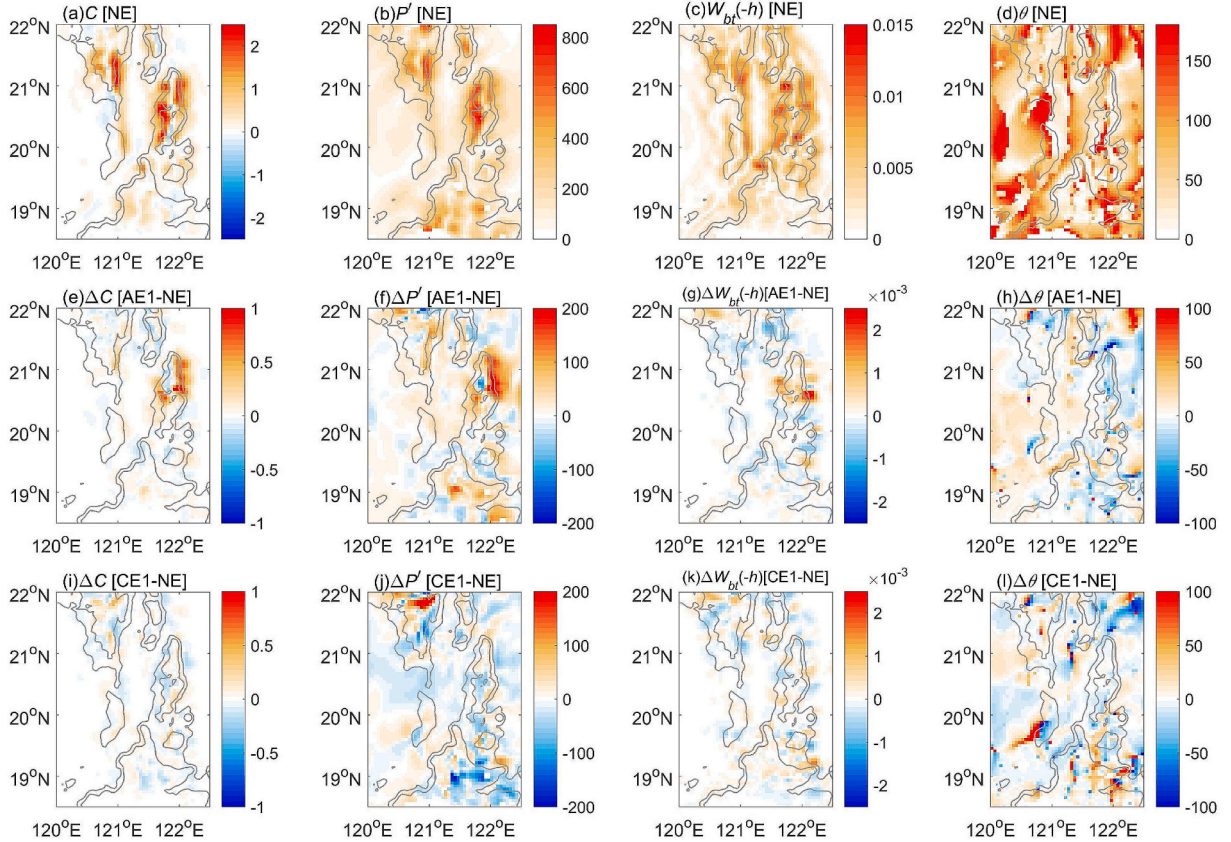


Fig. 6. (a) Conversion rate (C), (b) amplitude of bottom pressure perturbation (P'), (c) amplitude of bottom vertical barotropic tidal velocity ($W_{bt}(-h)$), and (d) their phase difference (θ) for the NE case. Their differences with the AE1 case are shown in panels (e–h), and differences with the CE1 case in panels (i–l). Grey lines indicate the 1000 m and 2000 m depth contours.

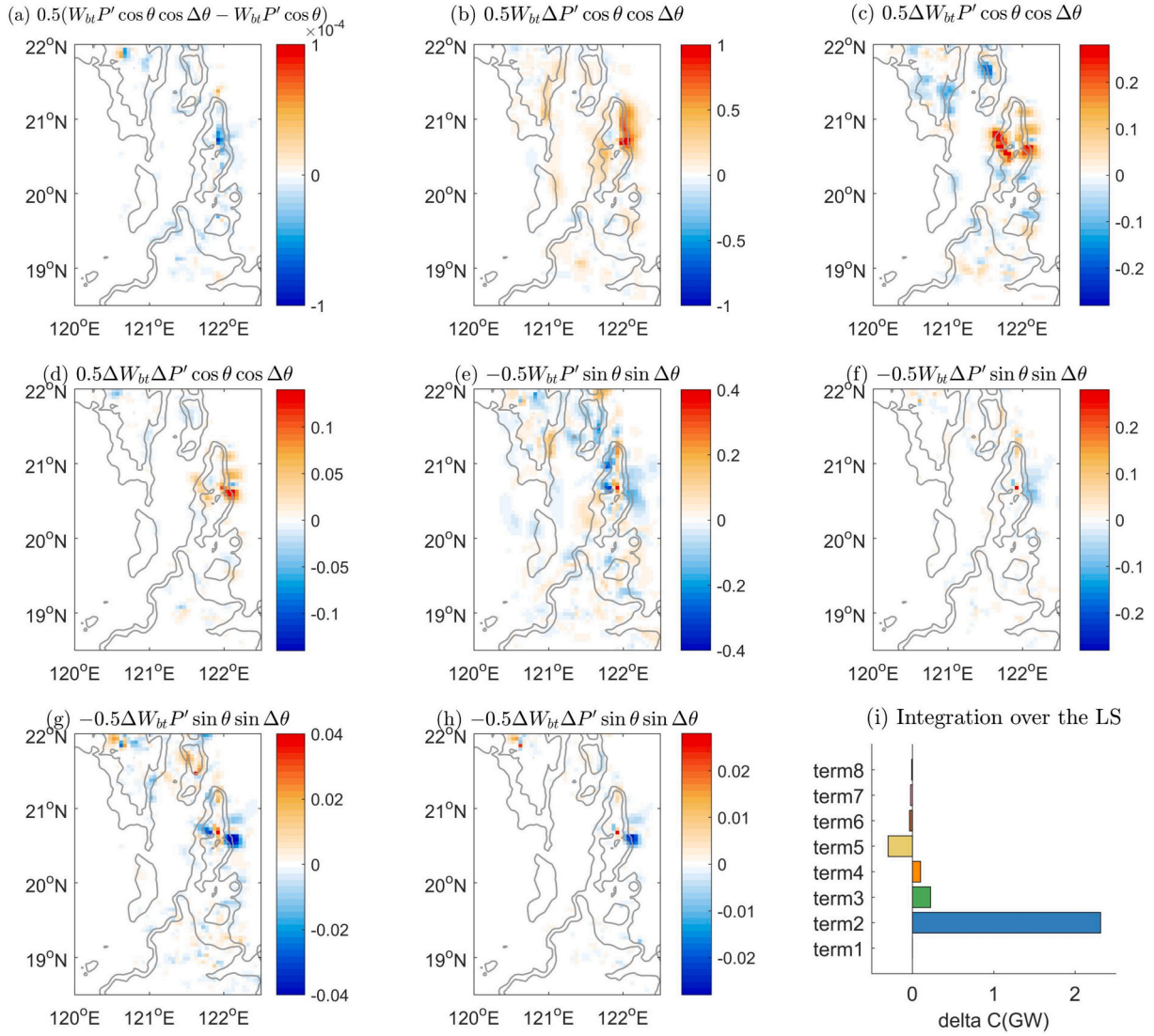


Fig. 7. Spatial distribution of the eight terms contributing to variation of conversion rates in the AE1 case: (a) $0.5(W_{bt}P'\cos\theta\cos\Delta\theta - W_{bt}P'\cos\theta)$, (b) $0.5W_{bt}\Delta P'\cos\theta\cos\Delta\theta$, (c) $0.5\Delta W_{bt}P'\cos\theta\cos\Delta\theta$, (d) $0.5\Delta W_{bt}\Delta P'\cos\theta\cos\Delta\theta$, (e) $-0.5W_{bt}P'\sin\theta\sin\Delta\theta$, (f) $-0.5W_{bt}\Delta P'\sin\theta\sin\Delta\theta$, (g) $-0.5\Delta W_{bt}P'\sin\theta\sin\Delta\theta$, and (h) $-0.5\Delta W_{bt}\Delta P'\sin\theta\sin\Delta\theta$. Integration of each term over the LS is presented in panel (i). Grey lines indicate the 1000 m and 2000 m depth contours.

pressure perturbation are given by

$$\frac{\partial p'}{\partial t}\bigg|_{z=-h} = - \int_{-h}^0 \rho_0 \frac{\partial b'}{\partial t} dz - \frac{\partial p_{surf}}{\partial t} \quad (32)$$

Results from the NE and eddy experiments show that surface pressure p_{surf} remains nearly identical, directing our attention to variations in b' . Based on Equation (14):

$$\frac{\partial b'}{\partial t} = -\bar{u} \frac{\partial b'}{\partial x} - \bar{v} \frac{\partial b'}{\partial y} - \bar{w} \frac{\partial b'}{\partial z} - u' \frac{\partial \bar{b}}{\partial x} - v' \frac{\partial \bar{b}}{\partial y} - w' \frac{\partial \bar{b}}{\partial z} \quad (33)$$

The magnitude of each term during the 50-hour period beginning on Day 55 in cases NE, AE1, and CE1 are compared in Fig. 9. The sixth term on the right-hand of the equation ($-w' \frac{\partial \bar{b}}{\partial z}$) is consistently the largest, and its difference between the eddy and NE cases exceeds that of other terms. In particular, the difference in the sixth term between AE1 and NE exhibits a pronounced value on the east side of the eastern ridge (Fig. 9l), at least one order of magnitude greater than the other terms. This location coincides with the increase in the amplitude of bottom pressure perturbation (P') and conversion rate (ΔC). This demonstrates that the vertical gradient of buoyancy associated with subtidal motions drives

the changes in b' , which in turn modifies P' . The vertical velocity of IT included in this term indicates that variation in b' not only drive changes in the ITs through P' but are also, in turn, influenced by it. Fig. S9 shows how eddies modify the subtidal component of vertical buoyancy gradient as they propagate westward. Once the perturbation reaches the LS, the ITs therein are correspondingly changed (Fig. S10).

The above findings suggest that the AE-CE asymmetry in the changes of IT conversion arises from a combination of factors. First, the vertical gradient of buoyancy associated with subtidal motions ($\frac{\partial \bar{b}}{\partial z}$) can be altered by the eddies. Open-ocean observations (Chaigneau et al., 2011; Zhang et al., 2014) have long established that CEs and AEs exhibit asymmetrical vertical core configurations. When they approach the LS, they modify the background stratification differently. Second, this change is further complicated by the asymmetric, non-linear interaction with the Kuroshio Current. For instance, Jan et al. (2017) demonstrated that when a CE interacts with the Kuroshio Current, the pycnocline slopes across the stream weakens. Conversely, an AE tends to strengthen the pycnocline. Finally, the vertical velocity of ITs included in the $-w' \frac{\partial \bar{b}}{\partial z}$ term implicates a wave-current feedback mechanism.

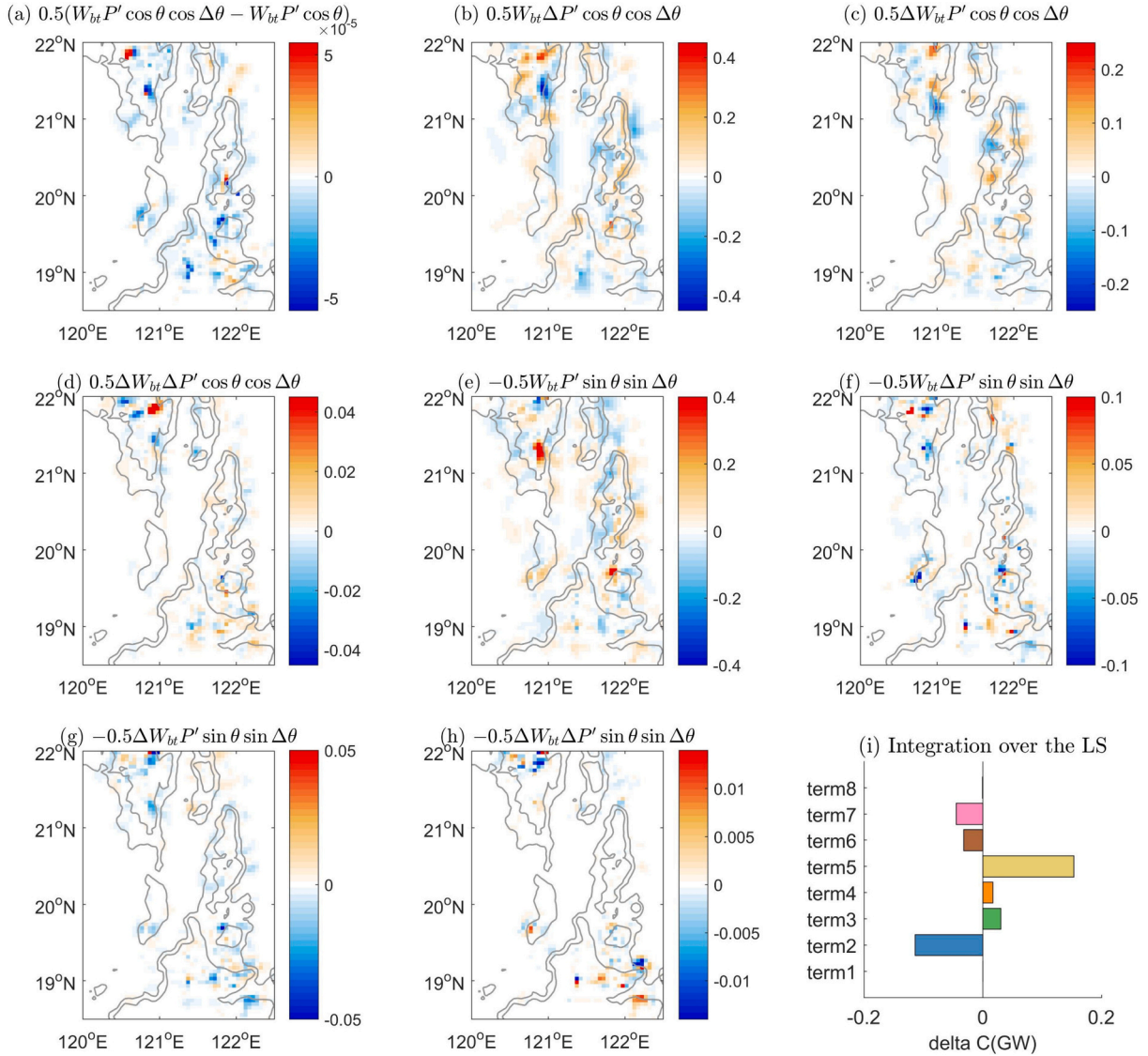


Fig. 8. Same as Fig. 7 but for the CE1 case. Note that the x-axis scale in panel (i) differs from that in Fig. 7.

4.2. Two stages of IT modulation by eddies

Results of this study indicate that eddies influence ITs originating from the LS differently in two stages: first during their propagation through the Philippine Sea, and subsequently after they impinge upon the Kuroshio Current.

In the first stage, IT energetics within the LS remain stable, which accounts for the strong coherence of IT in the LS and the South China Sea. However, in the Philippine Sea, the M_2 ITs exhibit significant variation as they propagate through the mesoscale eddies. The variation of ITs is influenced by multiple processes such as background stratification, advection by the subtidal current, and background relative vorticity, which can be assessed by their contribution to relative changes in the phase speed of mode-1 M_2 IT (Zaron and Egbert, 2014; Savage et al., 2020; Cao et al., 2022). Their relative contributions calculated following Equation (27) are illustrated in Fig. 10. Given that the background fields in the experiments are steady, except for the westward propagating eddies, it is reasonable to find that high values appear along the paths of the eddies. The largest contribution is due to the advection by the eddy-induced currents, exceeding 90% in most areas. The effect of relative vorticity is an order of magnitude smaller than that of advection, a contrast widely documented in previous studies (Zaron and

Egbert, 2014; Savage et al., 2020). Changes in stratification contributes the least among the three terms. The relative significance of stratification and currents associated with eddies is consistent with previous analyses based on the ray-tracing model (Guo et al., 2023).

During the second stage, as the eddies enter into the Kuroshio Current regime, significant changes are observed in the IT energetics within the LS. Concurrently, the incoherence of ITs increases in both the LS and the South China Sea. Jan et al. (2017) showed that the arrivals of eddies east of the Taiwan Island can have important remote effects on the Kuroshio Current in the Luzon Strait. The Kuroshio Current can modulate the generation and propagation of ITs, thereby increase their incoherence (Jan et al., 2012; Kerry et al., 2014; Varlamov et al., 2015; Xu et al., 2021; Cao et al., 2022; 2026). It is therefore suspected that the Kuroshio Current acts as an intermediate link, transmitting the influence of mesoscale eddies when they only temporally extend to, or even do not reach, the LS. To quantify the evolution of the Kuroshio Current, we calculate the LS transport (LST), which is widely used to characterize the variation and intrusion of the Kuroshio Current (Yang et al., 2013b; Nan et al., 2015). Fig. 11 shows that in the NE case, the transport remains relatively stable throughout the simulation with a mean value of 4.8 Sv. In the eddy cases, the LST closely mirrors the NE case during the first 40 days; however, as the eddies impinge upon the Kuroshio Current, the

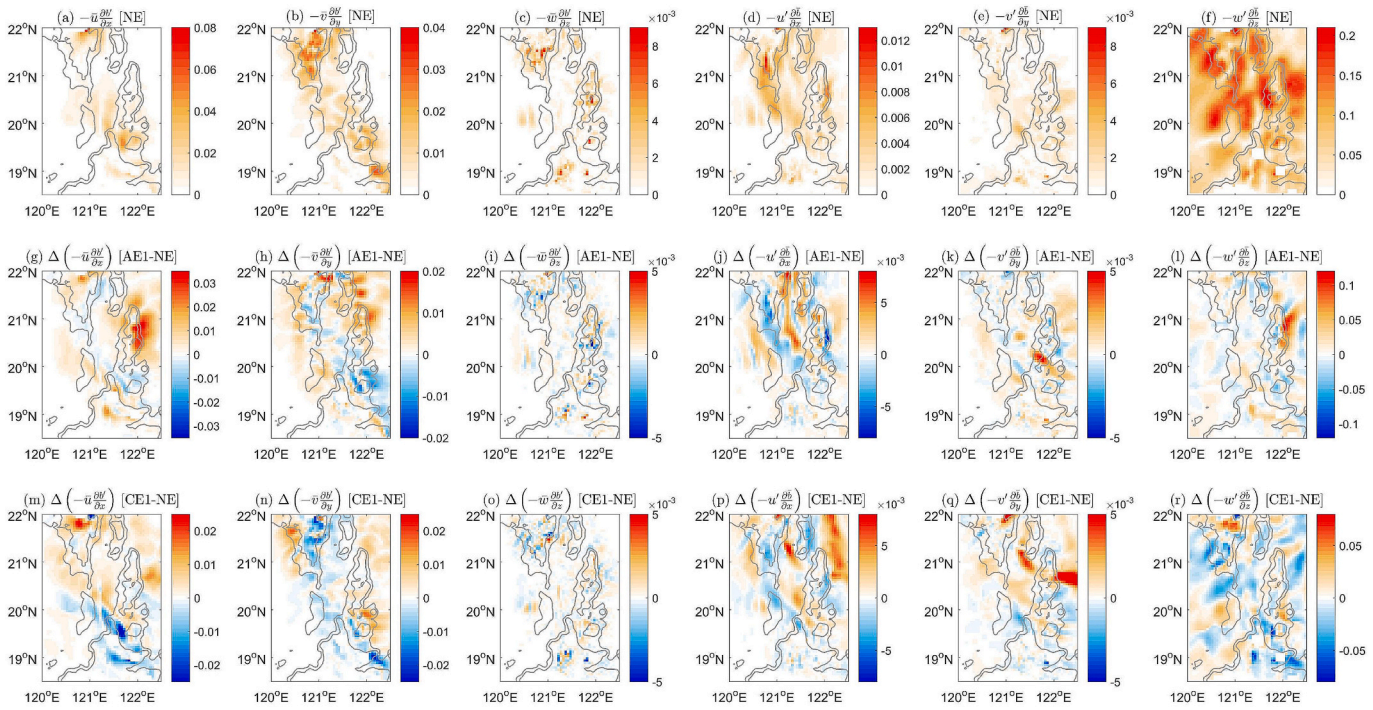


Fig. 9. Spatial distribution of the six terms contributing to variation of buoyancy associated with tidal motions (b) in the NE case: (a) $-\bar{u}\frac{\partial \bar{b}}{\partial x}$, (b) $-\bar{v}\frac{\partial \bar{b}}{\partial y}$, (c) $-\bar{w}\frac{\partial \bar{b}}{\partial z}$, (d) $-u'\frac{\partial \bar{b}}{\partial x}$, (e) $-v'\frac{\partial \bar{b}}{\partial y}$, and (f) $-w'\frac{\partial \bar{b}}{\partial z}$. Their differences with the AE1 case are shown in panels (g-l), and differences with the CE1 case in panels (m-r). Grey lines indicate the 1000 m and 2000 m depth contours. Note that the colorbar range varies with panel.

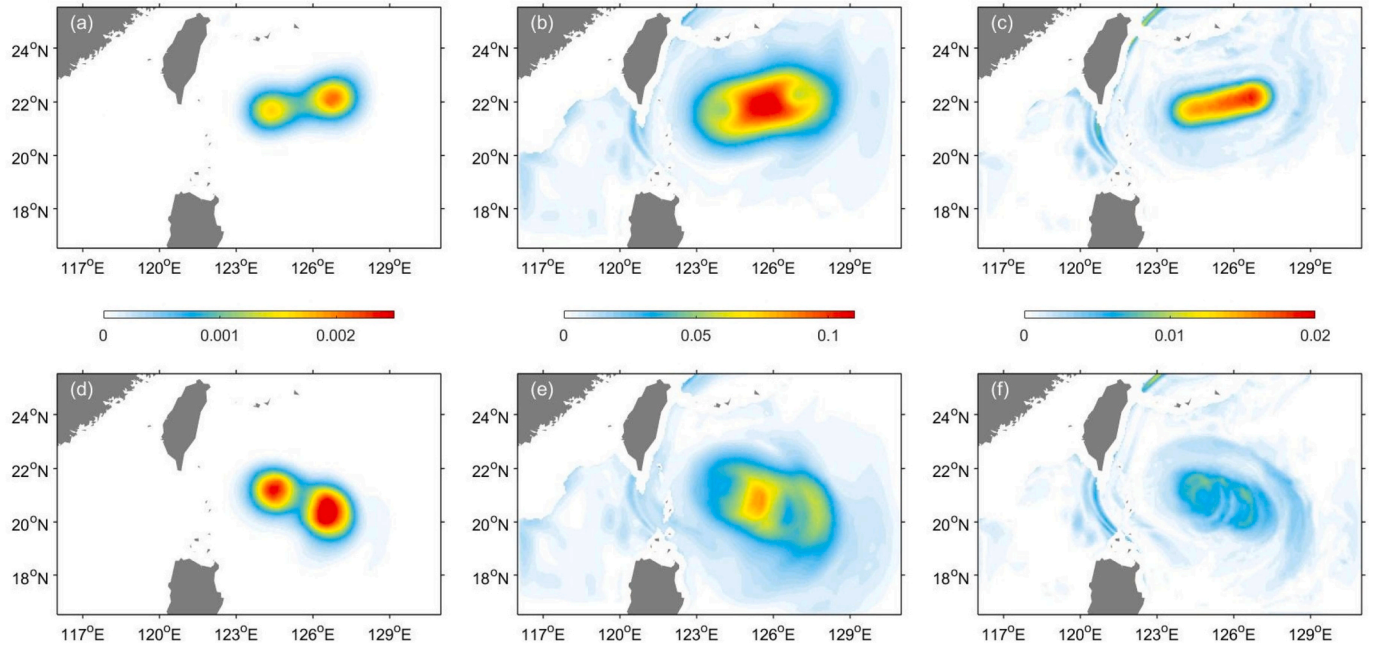


Fig. 10. Contributions of the (a,d) stratification terms $\frac{1}{2} \frac{\partial^2 \bar{b}}{c_1^2}$, (b,e) advection term $\sqrt{1 - \frac{f^2}{\omega^2}} \frac{\partial \bar{b}_{11}}{c_1}$, and (c,f) vorticity term $\frac{1}{2} \frac{f^2}{\omega^2} \frac{\partial^2 \bar{b}_{11}}{f}$ (Equation (27)) to the relative variation in phase speed of mode-1 M_2 ITs during the period from day 10 to 40 for the (a-c) AE1 and (e-f) CE1 cases. Note that each column employs a distinct color bar scale.

LST responds distinctively, exhibiting opposite trends with a maximum decrease of 0.72 Sv in the CE1 case and a maximum increase of 0.72 Sv in the AE1 case. On a finer structural scale, the 20.6°N sectional panels reveal distinct responses in the Kuroshio Current's core and surrounding stratification. In the AE1 case, a maximum velocity core exists at approximate 120.9°E on day 40. The core moves westward during the Kuroshio Current-eddy interaction, reaching 120.5°E by day 70. On day

40, a weak northward flow is observed in the upper layer east of 122°E, which strengthens by day 55 and dissipates by day 70. This flow is associated with the northward velocity in the western half of the AE1. Meanwhile, the pycnocline shows a complex pattern of variation dependent on its location. In the CE1 case, contrary changes are observed. The maximum velocity core moves eastward, and a weak southward flow associated with the CE is present on days 40 and 55.

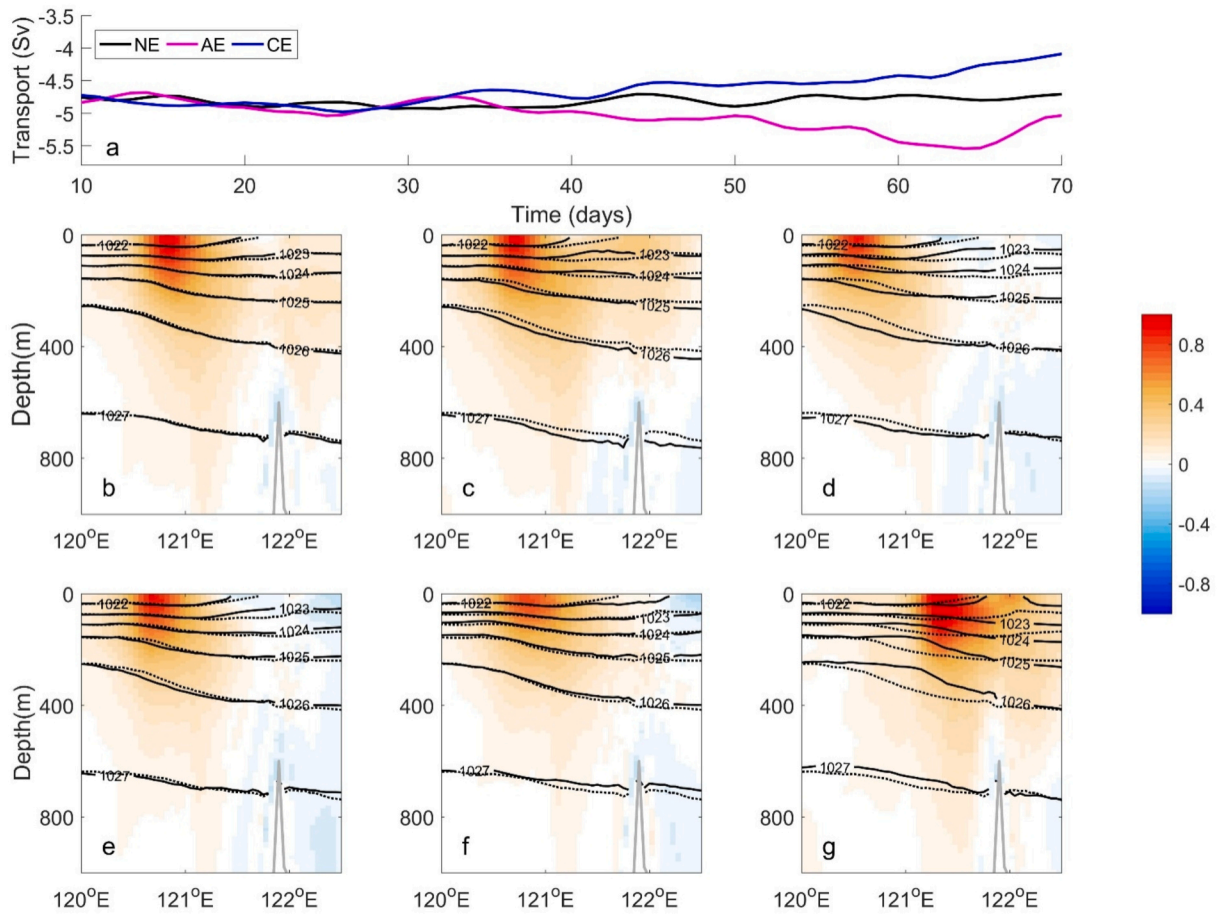


Fig. 11. (a) variations in LST for the NE, AE1, and CE1 cases. Meridional velocity (shading, unit: m/s) of the subtidal flow in the (b-d) AE1 and (e-g) CE1 cases on Days (b and e) 40, (c and f) 55 and (d and g) 70. The vertical profiles of isopycnal depths are superposed, with solid lines representing the eddy cases and dashed lines representing the NE case.

Additionally, the pycnoclines exhibit variations that are opposite to those in the AE1. These results confirm that the subtidal flow and stratification change as the eddy-Kuroshio Current interaction progresses. Xu et al. (2021) reveals that different intrusion paths of the Kuroshio Current significantly alter IT generation. Although the Kuroshio Current maintains a leaping path in this study, its associated flow and stratification vary under the influence of AEs/CEs. In this process, variations in tidal conversion, combined with energy exchange with subtidal motions, collectively enhance IT incoherence.

5. Summary

In this study, we investigate the influence of mesoscale eddies on the M_2 ITs based on numerical simulation. These idealized experiments situated to the LS provide a relatively clean environment to quantify the variation of ITs as an AE/CE propagates east of the LS and subsequently impinges on the Kuroshio Current.

The first key finding is the asymmetric influence of AEs and CE on tidal conversion. AEs induce more pronounced changes in the conversion rate, whereas CEs exert only minor effects. Further analysis reveals that variations in the conversion rate are primarily controlled by changes in bottom pressure perturbation, which is most strongly influenced by the product of the vertical gradient of buoyancy associated with subtidal motions and vertical velocity of IT.

Another noteworthy finding is that mesoscale eddies influence ITs differently in two stages. During the first stage, as the mesoscale eddies propagate in the Philippine Sea, they modulate the propagation of ITs through direct interaction and induce incoherence in ITs within the

Philippine Sea. The percentage of incoherent ITs could exceed 50% along the path of eddies. The eddy modulates IT propagation through multiple processes, of which the advection by the eddy-induced currents contributes up to 90%, a more substantial impact than background stratification and relative vorticity. As the eddies impinge on the Kuroshio Current and move northward, changes are evident in the energetics of ITs within the LS. The incoherence of ITs increases not only in the Philippine Sea but also in the LS and South China Sea. These variations in the second stage are attributed to changes in the Kuroshio Current during its interaction with eddies.

Overall, this study demonstrates that mesoscale eddies in the Philippine Sea not only affect the propagation of ITs but also modulate their energetics through interactions with the Kuroshio Current, both of which contribute to the increased incoherence in ITs. Compared to our idealized configuration, the generation of IT incoherence would likely be substantially amplified in a realistic ocean setting due to several concurrent mechanisms. For instance, the inclusion of the full tidal spectrum (e.g., M_2 , S_2 , K_1 , O_1) would introduce multi-constituent tidal interactions (Guo et al., 2020), broadening the frequency band of incoherent energy. Furthermore, replacing a single idealized eddy with a continuous, fully developed mesoscale eddy field and surface sub-mesoscale frontogenesis would enhance the scattering and dissipation of ITs, leading to much higher temporal volatility of ITs. Consequently, our idealized estimates of incoherence represent a conservative lower bound, and the real-world IT field around the LS is expected to be significantly more complex. These findings have significant implications for estimating ITs from altimeter measurements. While constructing an instantaneous IT model with a time window as short as 75 days has been

proven feasible with SWOT data (Zhao, 2024), our study demonstrates that even a single mesoscale eddy can rapidly induce substantial, spatially extensive incoherence in ITs within this time window. This means that a significant portion of IT energy remains incoherent and escape the 75-day model. Understanding this mechanism is therefore vital for future SWOT algorithms to properly correct incoherent ITs to study submesoscale motions.

CRedit authorship contribution statement

Zheng Guo: Writing – original draft, Visualization, Formal analysis. **Anzhou Cao:** Writing – review & editing, Methodology, Funding acquisition, Conceptualization. **Xu Chen:** Writing – review & editing. **Jinbao Song:** Writing – review & editing, Funding acquisition. **Xinyu Guo:** Writing – review & editing.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.pocean.2026.103808>.

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