



Response of near-inertial energy to a supercritical tropical cyclone and jet stream in the South China Sea: modeling study

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1 **ABSTRACT**

2 We used a well-validated three-dimensional ocean model to investigate the process of energetic
3 response of near-inertial oscillations (NIOs) to a tropical cyclone (TC) and strong background jet
4 stream in the South China Sea (SCS). We found that the NIO and near-inertial kinetic energy
5 (KENi) varied distinctly during different stages of the TC forcing, and the horizontal and vertical
6 transport of KENi was largely modulated by the velocity and vorticity of the jet stream. The KENi
7 reached its peak value within ~one-half the inertial period after the initial TC forcing stage in the
8 upper layer, decayed quickly by one-half in the next two days, and further decreased in a slower
9 rate during the relaxation stage of the TC forcing. Analyses of the KENi balance indicate that the
10 weakened KENi in the upper layer during the forcing stage was mainly attributed to the downward
11 KENi transport due to pressure work through the vertical displacement of isopycnal surfaces, while
12 upward KENi advection from depths also contributed to the weakening in the TC-induced
13 upwelling region. In contrast, during the relaxation stage as TC moved away, the effect of vertical
14 advection on KENi reduction was negligible and the KENi was chiefly removed by the outward
15 propagation of inertial-gravity waves, horizontal advection and viscous dissipation. Both the
16 outward wave propagation and horizontal advection by the jet stream provided the KENi source in
17 the far-field. During both stages, the negative geostrophic vorticity south of the jet stream
18 facilitated the vertical propagation of inertial-gravity waves.

19



20 **1 Introduction**

21 Near-inertial oscillations (NIOs), whose frequencies are close to the local inertial frequency
22 contain around half of the observed internal wave kinetic energy in the ocean (Simmons and
23 Alford, 2012). NIOs also greatly affect the kinetic energy budget in the deeper ocean as they
24 propagate downward from the surface and enhance the mixing by increasing vertical shear (Gill,
25 1984; Gregg et al., 1986; Ferrari and Wunsch, 2009; Alford et al., 2016).

26 Tropical cyclones (TCs), with the rapid change of wind stress, provide an important
27 generation mechanism for the NIOs. Observational studies related to a single storm or tropical
28 cyclone (Price, 1981; Shay and Elsberry, 1987; D'Asaro et al., 1995) showed that the NIOs related
29 to TCs can be a factor 2-3 larger than the background NIOs and last for more than 5 inertial periods
30 (IPs). Using a hurricane-ocean coupled model, Liu et al. (2008) estimated that the energy input of
31 tropical cyclones into the near-inertial currents was about 0.03 TW, about 10% of the total wind-
32 induced near-inertial energy (Watanabe and Hibiya, 2002; Alford, 2003). The input of wind energy
33 to the near-inertial band is also controlled by the translation speed of the TC (U_h) (Geisler, 1970;
34 Price, 1981). In fact, the NIOs are largely variable during the forcing and relaxation stage of the
35 TC forcing related to the intensity and translation speed of the TC. The variation is determined not
36 only by the different magnitude of input of wind energy, but also by the different dynamic
37 conditions that regulate the near-inertial kinetic energy (KENi) transport during these stages. The
38 variable response of NIOs during different stage of TC forcing is critical for understanding the
39 process and physics of NIOs.

40 Once generated, the characteristics of NIOs, in terms of the decay time scale, propagation
41 direction, and propagation speed are influenced by various mechanisms. In linear wave theory, the
42 β -effect leads to an equatorward propagation of the NIOs and reduces its decay time scale because



43 of the vertical propagation into deeper ocean (Gill, 1984; D'Asaro, 1989; Garrett, 2001).
44 Background flow fields also impose large influence on the evolution of NIOs. The influence of
45 background vorticity on NIOs has been observed by Weller (1982) and Kunze and Sanford (1984),
46 and analytical proofs by Kunze (1985), Young and Ben Jelloul (1997), and Danioux et al. (2015).
47 Numerical study using a primitive equation model with a turbulent mesoscale eddy field and
48 uniform wind forcing gave a similar conclusion (Danioux et al., 2008). Non-linear interactions
49 also provide a mechanism for increasing the vertical wave number, thus for larger vertical shear
50 and the dissipation and reducing the decay time scale (Davies and Xing, 2002; Zedler, 2009). In
51 addition to modifying the characteristics of the inertial-gravity waves, nonlinear advection related
52 to the front can transport NIOs away from the storm track and to higher latitudes (Zhai et al., 2004).
53 Recent observations and numerical studies in Gulf Stream, Kuroshio, and Japan Sea revealed the
54 role of vertical circulation on the generation and radiation of near inertial energy (Whitt and
55 Thomas, 2013; Nagai et al., 2015; Rocha et al., 2018; Thomas, 2019).

56 The South China Sea (SCS) is a region with frequent tropical cyclone occurrence, ~10.3
57 each year (Wang et al., 2007). Observations (Sun et al., 2011a,b; Xu et al., 2013) and numerical
58 studies (Chu et al., 2000) indicated that these TC events are sources for near-inertial energy bursts.
59 Additionally, the SCS circulation contains abundant energetic flow (e.g. Qu, 2000; Gan et al.,
60 2006; Gan et al., 2016a) and mesoscale features, such as a strong coastal jet off the Vietnamese
61 coast (e.g. Gan and Qu, 2008) and eddies (e.g. Chen et al., 2012). The distribution and evolution
62 of the TC-induced NIOs are susceptible to the influence of these background currents. (Sun et al.,
63 2011a; Sun et al., 2011b). However, the estimate of the total contribution of a TC to the KEni in
64 the SCS is difficult to obtain from observations or linear wave theory due to sparse spatial
65 observations and the non-homogenous nature of SCS circulation.



66 In this study, we apply a well-validated numerical model with specific China Sea
67 configurations to examine the response of KEni to a large TC and background stream jet over the
68 sloping topography in the SCS. A description of Typhoon Neoguri and the details of the numerical
69 model implementation are given in Section 2. In Section 3, the general characteristic response of
70 the near-inertial current and the energy fluxes during the TC forced stage and their later relaxation
71 stage as TC moved away from the concerned region are presented. Following Section 3, the KEni
72 equation is used to identify the dynamic processes of the vertical viscous dissipation, pressure
73 work, and nonlinearity during different phases of the NIOs.

74 **2 Typhoon Neoguri (2008) and the Ocean Model**

75 **2.1 Typhoon Neoguri (2008)**

76 Typhoon Neoguri formed east of the Philippines and entered the SCS on April 15. It first moved
77 west-northwest with an average translation speed of 5.8 m s^{-1} before it slowed down to 1.6 m s^{-1}
78 on April 16, based on the Joint Typhoon Warning Center (JTWC) best track data, (Fig. 1). On
79 April 17, Neoguri sped up to 3.8 m s^{-1} , turned more northward, and developed into a typhoon with
80 a maximum wind speed 51 m s^{-1} and a MSLP (minimum sea level pressure) at 948 hPa, at 1800
81 UTC on April 17 near the Xisha Islands. The Neoguri was a supercritical typhoon travelling with
82 a translation speed greater than the first baroclinic wave speed. After skirting Hainan Island on
83 April 18, Neoguri moved northward, weakened to a tropical storm, and further dissipated as it
84 moved farther inland. The NIO burst induced by Neoguri was shown by the clockwise (A_{cw}) and
85 counter-clockwise (A_{ccw}) rotary current amplitudes (m s^{-1}), from a current meter mooring at
86 Wenchang station, to the east off Hainan Island (Fig. 1c).



87 2.2 Ocean Model

88 We use the China Sea Multi-scale Ocean Modeling System (CMOMS) (Gan et al., 2016a; Gan et
 89 al., 2016b) in this study. CMOMS is based on the Regional Ocean Modeling System (ROMS)
 90 (Shchepetkin and McWilliams, 2005), and the model domain covers the northwest Pacific Ocean
 91 (NPO) and the entire China Seas (Bohai, Yellow Sea, East China Sea, and SCS) from
 92 approximately 0.95°N, 99°E in the southwest corner to the northeast corner of the Sea of Japan.
 93 The horizontal size of this grid array decreased gradually from ~10 km in the southern part to ~7
 94 km in the northern part of the domain. Vertically, we adopted a 30-level stretched generalized
 95 terrain-following coordinate (s).

96 The model was forced with 6-hourly actual wind speeds of typhoon Neoguri obtained from
 97 the Cross-Calibrated Multi-Platform (CCMP) dataset, with a horizontal resolution of 0.25° (Atlas
 98 et al. (2011), <ftp://podaac-ftp.jpl.nasa.gov/allData/ccmp/L3.0/flk>). Wind stress is calculated based
 99 on the bulk formulation by Fairall et al. (2003). The daily mean air temperature, atmospheric
 100 pressure, rainfall/evaporation, radiation, and other meteorological variables from April 15 to April
 101 18, 2008 from the NCEP/NCAR Reanalysis 1 were used to derive the atmospheric heat and fresh
 102 fluxes. External forcing of depth-integrated velocities (U , V), depth-dependent velocities (u , v),
 103 temperature, T , and salinity, S , at the lateral boundaries were obtained from the Ocean General
 104 Circulation Model for the Earth Simulator (OFES) (Sasaki et al., 2008). Open boundary conditions
 105 from Gan and Allen (2005) were applied at the open boundaries.

106 The model was spun up from January 1, 2005 with winter initial fields (temperature and
 107 salinity) obtained from the last three-year mean fields of a 25-year run that is initialized with the
 108 World Ocean Atlas 2005 (WOA05, Locarnini et al., 2006, Antonov et al., 2006) data, and forced
 109 by wind stress derived from climatological (averaged from 1988 to 2013) monthly Reanalysis of



110 10 m Blended Sea Winds released by the National Oceanic and Atmospheric Administration
111 (<https://www.ncdc.noaa.gov/oa/rsad/air-sea/seawinds.htm>). The dynamic configuration and
112 numerical implementation of the CMOMS system are described in detail in Gan et al. (2016a,
113 2016b).

114 We have thoroughly validated the CMOMS by comparing simulated results with those
115 obtained from various measurements and findings in previous studies. In particular, we have
116 validated the extrinsic forcing of time-dependent, three-dimensional current system in the tropical
117 NPO, transports through the straits around the periphery of the SCS, and corresponding intrinsic
118 responses of circulation, hydrography and water masses in the SCS (Gan et al., 2016a). We have
119 also validated the circulation of CMOMS by providing a consistent physics between the intrinsic
120 responses of the circulation and extrinsic forcing of flow exchange with adjacent oceans (Gan et
121 al., 2016b). The model is also validated with available ARGO temperature profiles (not shown),
122 observed sea surface temperature (SST) and currents from a time-series current meter mooring
123 during Neoguri, as described below.

124 Three-dimensional, hourly-mean dynamic, and thermodynamic variables from April 10 to
125 May 10, 2008 were used to examine the near-inertial oscillations in this study. Because the inertial
126 period (IP) in the SCS is larger than 32 hours (near 22°N), the error induced by the hourly model
127 output is <3%.

128 **3 Model result**

129 **3.1 Characteristic response to the TC**

130 The evolution of the response to the TC in the ocean with existence of a coastal jet in the
131 SCS is presented according to different stages of the TC forcing. During the pre-storm stage (PS),
132 before Neoguri entered the SCS on April 14, the wind stress was relatively weak (<0.1 Pa). A



133 prominent jet current separated from the Vietnamese coast flowing eastward near 16°N (Fig. 2a),
 134 and characterized the circulation in the western part of the SCS (Gan and Qu, 2008). The jet formed
 135 negative (positive) geostrophic vorticity (ζ_g) to the south (north), with the minimum (maximum)
 136 Rossby number (ζ_g/f) <-0.2 (>0.2) near 15.8°N (16.8°N). During the forced stage (FS, Fig. 2b)
 137 between April 15 and April 19, the SCS was under the direct influence of Neoguri, the wind forcing
 138 became significantly stronger (>0.1 Pa), the KE near the surface (10 m) intensified significantly
 139 ($>500 \text{ J m}^{-3}$) to the east of the TC, and the coastal jet was suppressed by southward flow.
 140 Meanwhile, a strong local divergence and upwelling formed in the surface and generated a strong
 141 cooling ($\sim 1.5^\circ\text{C}$) belt along the TC path that lasted for more than a week. The cooling zone radiated
 142 hundreds of kilometer away from the core of the TC. These features were well captured by the
 143 TC-induced temperature difference between April 19 and 14 from both simulated (Fig. 3a) and
 144 observed SST (Fig. 3b) (<http://podaac.jpl.nasa.gov/dataset/JPL-L4UHfnd-GLOB-MUR>). After
 145 the end of the FS on April 20 (Fig. 2c), the jet returned to its pre-storm intensity and shifted slightly
 146 northward (Fig. 2c) when the TC center approached the coast (Fig. 1a). Afterwards, during the
 147 relaxation stage (RS) after April 20 (Fig. 2d), the wind forcing from the TC decreased to <0.05 Pa.
 148 ———Rotary spectra shows that the near-inertial response of surface currents to the TC occurred
 149 near the local inertial frequency ($f = 0.028$ cph) at station Wenchang (112°E, 19.6°N) during the
 150 model simulation period (April 10 – May 5) (Fig. 4). The clockwise rotary spectra is calculated
 151 by:

$$152 \quad S_{cw} = 1/8(P_{uu} + P_{vv} - 2Q_{uv}), \quad (1)$$

153 where P_{uu} , P_{vv} and Q_{uv} are auto- and quadrature-spectra, respectively (Gonella, 1972). This
 154 simulated result is highly consistent with the observations in the lower frequency band. A relatively
 155 large discrepancy between the model output and observations at the higher frequency band could



156 have been caused by many reasons, such as the lack of mesoscale and sub-mesoscale processes in
 157 the atmospheric forcing field, and not resolving the oceanic subscale processes due to the limitation
 158 of current model resolution.

159 **3.2. Near-inertial response in the upper ocean**

160 We adopted the complex demodulation method successfully used in previous NIO studies
 161 (Gonella, 1972; Brink, 1989; Qi et al., 1995) to extract the inertial current signal. The simulated
 162 horizontal currents (\vec{u}_h) were analyzed for inertial currents (\vec{u}_i). The inertial currents contain
 163 clockwise (*cw*) and counter-clockwise (*ccw*) rotating components:

$$164 \quad u_i + iv_i = A_{cw} e^{-i(\phi_{cw} + ft)} + A_{ccw} e^{i(\phi_{ccw} + ft)}, \quad (2)$$

165 where u_i and v_i are the eastward and northward inertial currents at 10 m in the mixed layer, A and
 166 ϕ are the amplitude and phase of the rotary currents, respectively. Subscripts represent the
 167 clockwise (*cw*) and counter-clockwise (*ccw*) rotating direction, and f is the local Coriolis
 168 coefficient. To obtain the amplitude and phase, we performed harmonic analysis daily with each
 169 segment over one inertial period (IP). Then the rotary amplitude and phase were calculated
 170 following previous studies (Mooers, 1973, Qi et al., 1995, Jordi and Wang, 2008).

171 The time evolution of the daily rotary currents during the FS and RS in the surface layer
 172 varied spatially and was related to the intensity and translation speed of the TC. On April 15 during
 173 PS, Neoguri affected mainly the region south of 13°N, with a relatively fast translation speed (U_h
 174 $> 3C_l$, Fig. 1b) and weaker intensity ($V_{max} \sim 35 \text{ m s}^{-1}$). In most areas, *cw* rotary currents were strong
 175 ($A_{cw} > 0.1 \text{ m s}^{-1}$) yet decayed quickly after 3 days ($< 2 \text{ IP}$) (Fig. 5a), while the magnitudes of *ccw*
 176 currents were very small (Fig. 5b). After April 18, Neoguri moved into the region between 14°N
 177 and 18°N, where it intensified more than 40% but moved slower with $U_h \sim 2C_l$. Both the *cw* and
 178 *ccw* currents possessed larger intensities than in the southern region. The induced *cw* currents



displayed an obvious rightward bias, where the enhanced inertial currents extended to ~350 km to the right of the track and to ~<150 km to the left of the track. This extension of horizontal scale was related to the region with wind stress $|\tau| > 0.25$ Pa in Neoguri.

The maxima of the *ccw* component were located to the left of the TC's path where the wind vector (Fig. 6) rotated in the same direction as the ocean currents presented in Fig. 5. The connection between the right (left) bias of the *cw* (*ccw*) currents with the rotation direction of the wind vector is in agreement with the explanation of Price (1981). 2-3 IPs (>6 days) after the direct forcing, the *cw* currents remained significant (>0.2 m s⁻¹) in an area extending from 110°E to 116°E. In contrast, the *ccw* components, which rotated in the direction of the Earth's rotation, dissipated quickly after the wind forcing stopped within ~1 day. This short duration of the forced inertial motion is in agreement with previous studies (Jordi and Wang, 2008).

Besides the intensity and duration, we also looked at the frequency shift ($\delta\omega = \omega - f$) and the horizontal scale of the NIOs. The frequency shift from the local inertial frequency was estimated from the temporal evolution of the phase of the back rotary current: $\delta\omega = -\partial\phi/\partial t$. In the FS, the maximum frequency shift occurred near the jet (112°E to 115°E, 15°N to 16°N), where $\delta\omega \approx 0.08f$ ($\Delta\phi \approx \pi/4$, $\Delta t = 3$ days, $f = 4 \times 10^{-5}$ s⁻¹ at 16°N). The horizontal scale was estimated from the spatial variation of the rotary current by calculating the horizontal wave number in the meridional direction as $k_y = \partial\phi/\partial y$. The largest wave number $k_y \approx 3.1 \times 10^{-5}$ rad m⁻¹ was also found near the jet.

3.3 Characteristic near-inertial energy

Response in the upper layer

We focused on the area between 110-115°E and 13-19°N (box in Fig. 5a), defined as the forced region, where the strongest NIO was produced during FS of Neoguri. We calculated the



202 wind-induced near-inertial energy flux (or the wind work) using $\vec{\tau}_i \cdot \vec{u}_i$, where $\vec{\tau}_i$ is the band-passed
 203 near-inertial wind stress and \vec{u}_i is the near-inertial current at the surface (Silverthorne and Toole,
 204 2009). A 4th order elliptic band-pass filter (Morozov and Velarde, 2008) was applied to obtain
 205 near-inertial motion with a band ranging from $0.8f$ to $1.2f$, where f is the local Coriolis coefficient.
 206 The time series of domain-averaged $\vec{\tau}_i \cdot \vec{u}_i$ over the forced region reveals that significant energy
 207 input took place during the FS, with the peak value about $68 \times 10^{-3} \text{ W m}^{-2}$ on April 17 (Fig. 7).
 208 Under this large wind energy input, the area-averaged depth-integrated KEni (or *AKEni* hereafter)
 209 in the upper layer (0-30 m) increased significantly from its pre-storm value to a maximum ~ 1500
 210 J m^{-2} during the FS, with an increase rate of about $16 \times 10^{-3} \text{ W m}^{-2}$ (Fig. 7a). Despite the continuous
 211 positive wind energy flux, the *AKEni* in the upper layer plateaued, indicating that a large amount
 212 of the wind energy was either propagating out the forced region or was lost to the lower layers due
 213 to entrainment. The detailed mechanisms are discussed in the following sections. After the peak
 214 of FS, the wind work decreased significantly with small negative value around the end of the FS.
 215 The *AKEni* decreased to one half of its peak value within 2 days (decrease rate was about 7×10^{-3}
 216 W m^{-2}). After that, the wind work was almost negligible, and the decrease rate of *AKEni* became
 217 smaller ($\sim 0.8 \times 10^{-3} \text{ W m}^{-2}$).

218 ***Response at depths***

219 During the FS, the *AKEni* in the upper 200 m constituted $\sim 90\%$ of the total *AKEni* in the
 220 whole water column, while the *AKEni* between 30-200 m alone accounted for $\sim 30\text{-}50\%$ (Fig. 7b).
 221 The KEni in this mid-layer had a temporal evolution different from that in the upper layer. It
 222 reached its maximum on April 20, around one and a half days later, and was more than 80% of the
 223 peak value of the *AKEni* in the upper layer ($\sim 1000 \text{ J m}^{-2}$). Compared to the upper layer *AKEni*, the
 224 the mid-layer *AKEni* during FS increased slightly more slowly ($\sim 4 \times 10^{-3} \text{ W m}^{-2}$) while from April



225 20 to May 5 during RS it decreased much more slowly ($\sim 0.61 \times 10^{-3} \text{ W m}^{-2}$), and the *AKEni* became
 226 greater than that in the upper layer at the end of FS. The *AKEni* below 200 m was much smaller,
 227 but increased continuously from April 17 to 29, with a rate of about $0.62 \times 10^{-3} \text{ W m}^{-2}$, which was
 228 comparable to the *AKEni* rate of decrease in the 30-200 m layer. The *AKEni* in this deep layer
 229 became greater than that in the upper layer after April 25 and that in the layer 30-200 m after April
 230 29; the deep layer reached its maximum value >10 days later.

231 Spatially, two relatively large KEni patches below upper layer were located to north and
 232 south of $\sim 16^\circ\text{N}$ (Fig. 8a). Their horizontal scales, influenced by near-inertial waves, were much
 233 smaller compared with those in the upper layer. The region with relatively large KEni in the layer
 234 between 30-200 m located near the jet currents, with stronger value during FS than during RS (Fig.
 235 8 a,b). These results suggest that the KEni in this layer might have been determined by both vertical
 236 propagation of the near-inertial gravity wave and horizontal advection of KEni of the background
 237 current. Similar horizontal distribution also occurred below 200 m (Fig. 8 c,d). In contrast to the
 238 layer above, the relatively large value during RS on April 30 indicated a downward propagation
 239 of KEni into the deeper layer. Around the saddle zone west of the Xisha Islands, a relatively large
 240 KEni below 200 m aligned with the 1000 m isobath, and might reflect a topographic effect on the
 241 near-inertial wave.

242 **3.4 Vertical propagation of near-inertial energy**

243 It is clear that the distribution of the KEni was mainly controlled by the propagation of
 244 near-inertial wave energy both horizontally and vertically as well as by the background jet stream.
 245 In order to understand the KEni distribution in the deeper water inside the forced region and in the
 246 far field, we selected four different locations, marked as A1, A2, C1, and C2 in Fig. 8 (a-d), for
 247 the analysis of the KEni evolution during the FS and RS. Among them, A1 (113°E , 15.7°N) and



A2 (112.5°E, 16.9°N) are on the right side of the TC track, inside the forced region, and situated about 200 km apart from each other at the northern (A2) and southern (A1) sides of the jet, respectively. C1 (114.9°E, 16.9°N) and C2 (114.9°E, 18°N) are the corresponding stations in the far field where relatively strong KEni intensification occurred.

Forced region (stations A1 and A2)

South of the jet stream at station A1

The time series of the band-passed inertial velocity u_i as a function of depth shows that there was an upward phase propagation, in which u_i in the layers below 100 m was leading the upper 50 m (Fig. 9a). Accompanying this phase propagation was a downward propagation of surface KEni, which was represented by the lowering of the u_i maxima as a typical Poincaré wave (Kundu and Cohen, 2008). There were two phases of vertical energy propagation: 1) during FS, there was a rapid extension of the large u_i maxima to below 100 m from April 17 to 20, and 2) during RS, the center of the large u_i value descended from ~100 m to 280 m from April 25 to May 5. The vertical propagation velocity, C_{gz} , estimated from this downward transport, was ~17 m day⁻¹.

During the first phase, the KEni in the top 30 m and in the 30-200 m layer shared a similar rate of increase on April 17, indicating that the enhancement of KEni at 30-200 m was related to the entrainment between the upper and deep layer (Fig. 9c). While the KEni in the upper 30 m decreased quickly from April 18, it kept increasing at depths from 30-200 m, suggesting that other contributing mechanisms existed besides the entrainment. The KEni below 200 m also experienced notable intensification, with a smaller increasing rate than that found in the 30-200 m layer (Fig. 9c). Because the viscous effect is small in the deeper water, this enhancement of the KEni was most likely associated with the propagation of an inertial-gravity wave.

During the second phase, the KEni in the 30-200 m layer decreased significantly at station



271 A1, indicating the existence of either downward or horizontal energy transport. From the linearized
 272 inertial-gravity wave equation under the influence of background vorticity, C_{gz} can be obtained by

$$273 \quad C_{gz} = \frac{\omega^2 - f_{eff}^2}{\omega m}, \quad (3)$$

274 where $\omega \approx 1.08f$ is the frequency with maximum S_{cw} at 200 m (Fig. 9e); $f_{eff} = f + \zeta_g/2$ is the effective
 275 Coriolis coefficient; and $\zeta_g/f = 0.16$ at A1. m is the vertical wave number that we chose to be the
 276 first baroclinic mode under a two-layer approximation based on the stratification (blue line in Fig.
 277 10). From Eq. (3), C_{gz} was about 13.6 m day^{-1} , which was in the same range as the modeled C_{gz} .
 278 Consistent with the case of $(\omega - f_{eff})/f_{eff} < 0.1$ in Kunze (1985), the background vorticity in our case
 279 accounted for more than 90% of the modification of the magnitude of the wave dispersion property.
 280 Meanwhile, the KEni in the layer below 200 m did not increase notably, suggesting that other
 281 mechanisms besides vertical propagation of the near-inertial gravity wave might have been
 282 important in the evolution of KEni in water deeper than 200 m.

283 *North of the jet stream at station A2*

284 At location A2, strong u_i was mainly trapped in the water above 100 m, and below 100 m
 285 $u_i < 10 \text{ cm s}^{-1}$. It returned to its pre-storm magnitude after 5 IPs (Fig. 9b). The local upwelling
 286 related to the positive background vorticity and to the associated strong surface divergence might
 287 have caused the smaller vertical scale of u_i at A2 (Fig. 4b). The KEni was generally smaller than
 288 that at A1, and relatively large energy was found only in the ML (Fig. 9d). C_{gz} at A2, estimated
 289 from Eq. (3), was 1.2 m day^{-1} ($f_{eff} = 1.065f$, $\omega \approx 1.1f$, $f = 4.2 \times 10^{-5} \text{ s}^{-1}$, and $m = 2\pi/30 \text{ m}$), which was
 290 about one tenth of that at A1. This is consistent with the lack of a distinct pattern of vertical
 291 propagation of NIOs at this station, as shown in the band-passed u_i (Fig. 9b), and the presence
 292 (absence) of a near-inertial peak of S_{cw} at 10 m (200 m) (Fig. 9f).

293 *Far field region (stations C1 and C2)*



294 C1 and C2 are located ~400 km to the right of the forced region. During the FS, u_i (Fig. 11a,b) and
 295 KEni (Fig. 11c,d) in the upper layer were smaller than those at those stations in the forced region
 296 due to the weaker TC influence. Only a small downward propagation was discerned during the FS
 297 (Fig. 11a,b). However, notable intensification of the KEni occurred in the layers below the upper
 298 layer after April 23. At C1, the S_{cw} at 10 m had a small red shift, while the S_{cw} at 200 and 500 m
 299 displayed blue shifts with peaks near $1.07f$ (Fig. 11e). The difference between S_{cw} in the upper
 300 layer and in the layers below implies another source of KEni other than local inertial-gravity wave
 301 vertical propagation.

302 At C2, downward energy propagation appeared after April 23, reaching 100 m from the
 303 surface within 7 days, giving $C_{gz}=14.3 \text{ m day}^{-1}$ (Fig. 11b). Unlike C1, the intensification was
 304 mainly in the 30-200 m layer. The S_{cw} at both 10 m and 200 m had a broad energy band near the
 305 local f (Fig. 12f). Because $\zeta_g/f = 0.12$ and $f_{\text{eff}} = 1.06f$, C_{gz} estimated from Eq. (3) had an upward
 306 propagation, which cannot explain the downward propagation here. The linearized wave theory,
 307 with the consideration of Doppler drift due to background currents, does not seem to be valid in
 308 this location. We will discuss this issue in the next section.

309 **4 KEni Budget**

310 We utilized the KEni equation to provide a further analysis of the source of KEni in the water
 311 column. Because the horizontal component of near-inertial kinetic energy is significantly larger
 312 than the vertical component (Hebert and Moum, 1994), we used the horizontal component to
 313 represent the KEni. The KEni budget can be obtained from the horizontal momentum equation.
 314 Following Silverthorne and Toole (2009), the energy equation becomes:

$$315 \quad \underbrace{\frac{\partial KEni}{\partial t}}_{RATE} = \underbrace{-\bar{u}_i \cdot \langle \nabla_h p \rangle}_{PRES} - \underbrace{\rho_0 \bar{u}_i \cdot \langle \bar{u}_h \cdot \nabla_h \bar{u}_h \rangle}_{NLh} - \underbrace{\rho_0 \bar{u}_i \cdot \left\langle w \frac{\partial \bar{u}_h}{\partial z} \right\rangle}_{NLv} - \underbrace{\rho_0 \bar{u}_i \cdot \left\langle \frac{\partial}{\partial z} \left(v \frac{\partial \bar{u}_h}{\partial z} \right) \right\rangle}_{VVISC} \quad (4)$$



where $AKEni$ is the area-averaged depth-integrated near-inertial energy; \bar{u}_i and \bar{u}_h are the near inertial velocity vector and horizontal velocity, respectively; p is pressure; ρ_0 is the reference density; ∇_h is the horizontal gradient operator; w is the vertical velocity; ν is the viscosity coefficient; and the angle bracket represents band-passed filtering on the near-inertial band. The $PRES$ term on the right side of equation represents the pressure work on the $AKEni$, which is associated with the inertial-gravity wave propagation. NL_h and NL_v represent the horizontal and vertical divergence of energy flux that include the effects of 1) the advection of $AKEni$ due to background currents and 2) the straining of the wave field due to the background shear currents. Zhai et al. (2004) found that the geostrophic advection of $AKEni$ contributed most of the NL_h and was the main mechanism for transporting the NIOs in the absence of baroclinic dispersion of inertial-gravity waves. It was also found to be more important than the dispersive processes along the Gulf Stream or shelf-break jet. $VVISC$ is the vertical viscous effect. As before, we integrate the $KEni$ vertically in three layers: the upper layer (0-30 m), the subsurface layer (30-200 m), and the deep layer (>200 m). In the following sections, the $AKEni$ budget is considered in entire forced region (Fig. 5) as well as at the specific stations along the jet stream.

4.1. Mean balance

Figure 12 shows the time series of the $AKEni$ budget over the entire forced region defined in Fig. 5. The time-averaged horizontal distributions of each term are presented in Fig. 13. During the FS, the increase of $AKEni$ in the upper layer was mainly attributed to the wind energy input because the $VVISC$ term was one order larger than the other terms, with a maximum of $30 \times 10^{-3} \text{ W m}^{-2}$ on April 17 (Fig. 12a). The time-integrated $VVISC$ during the FS was $2.15 \times 10^3 \text{ J m}^{-2}$ (Table 1). Stronger $VVISC$ in the upper 30 m occurred in the region between 14°N and 18°N (Fig. 13a) along the TC track with a rightward bias, similar to the distribution of current intensity (Fig. 5c). Like



the wind work during the FS (Fig. 7a), *VVISC* became negative after April 19, indicating the *AKEni* removal by negative wind work. The influence of *VVISC* extended to the 30-200 m layer, and provided a positive energy flux ($\sim 1 \times 10^3 \text{ J m}^{-2}$) in this layer (Figs. 12b, 13b). The effect of *VVISC* in the deep layer was negligible (Fig. 12c).

Shortly (~ 1 day) after the large injection of *KEni* into the upper layer during the FS, the *PRES* became significant (Fig. 12a, Table 1) and its horizontal distribution resembled that of *VVISC* (Fig. 13a, d), suggesting that *PRES* radiated the *KEni* out of the forced region. It provided a negative *KEni* flux in the upper layer ($-0.65 \times 10^3 \text{ J m}^{-2}$), which was largely compensated by the positive flux in deeper layers (Table 1). This suggests that, during the FS, the main role of the pressure work was to transport the *KEni* from the upper layer to the deep layers, and $< 15\%$ of the *KEni* was horizontally propagated outside the forced region.

During the RS, the *VVISC* was relatively small in the upper layer and it accounted for one third of the *AKEni* removal in the layers below (Table 1). The *PRES* became a major sink for *AKEni* in the ML ($-0.85 \times 10^3 \text{ J m}^{-2}$) and subsurface layer ($-0.16 \times 10^3 \text{ J m}^{-2}$), but was the major source in the water below 200 m. The *AKEni* loss due to the horizontal wave propagation outside the forced region was $\sim -0.42 \times 10^3 \text{ J m}^{-2}$, accounting for about 40% of the total loss in the whole water column.

Nonlinear advection terms had an important influence in the top 200 m but made little contribution to the *AKEni* budget in the water below 200 m (Fig. 12, Table 1). The horizontal effects of NL_h and NL_v in these layers were mainly limited to a smaller region, as compared to the *VVISC* and the *PRES*; and their relatively large values occurred near the slope and the jet (Figs. 13).



361 In the upper layer, NL_h advected the KEni from the source region; NL_h had positive and
 362 negative values on the eastern and western sides of the TC track, respectively (Fig. 13g). Similar
 363 features, but with much weaker amplitude, were found in the layers below (Fig. 13h,i). During the
 364 FS, in the 30-200 m layer, the domain-averaged NL_h was positive ($0.21 \times 10^3 \text{ J m}^{-2}$), indicating a
 365 possible extraction of the KEni from background flows. NL_v was a strong energy sink in the upper
 366 200 m ($\sim 0.64 \times 10^3 \text{ J m}^{-2}$). The TC wind field generated a strong surface horizontal divergence and
 367 upwelling around $16\text{--}17^\circ\text{N}$ (Fig. 4b). As a result, the smaller KEni in the lower layer was advected
 368 to the surface east of the Xiasha Islands. This lower KEni generated a negative gradient with
 369 ambient water and resulted in the strong eastwards transport of KEni in the eastward jet current.
 370 As a result, a positive NL_h center located around the area with the strongest negative NL_v , and a
 371 negative NL_h center lay to the west of the positive maximum of NL_h . During the RS, NL_h became
 372 negative for all layers and provided $\sim 1/3$ of the total KEni loss in the water column ($-0.35 \times 10^3 \text{ J}$
 373 m^{-2}), while NL_v over the whole water column was significantly reduced.

374 **4.2 Role of the jet stream**

375 We further show the distinct $AKEni$ balance in the southern and northern sides of the jet stream.
 376 During the FS, $VVISC$ at A1 on the southern side of the jet stream was the dominant $AKEni$ source
 377 in both the upper layer and the 30-200 m layer (Fig. 14a,b), consistent with the large vertical scale
 378 on the southern side of the jet due to local negative vorticity. The enhancement of near-inertial
 379 currents in the upper layer and the concurrent current divergence resulted in the vertical oscillation
 380 of isopycnals (pressure) below the upper layer at this station. During this pumping process, the
 381 $AKEni$ in the upper layer was partly transported downward by the $PRES$ and partly by the NL_v .
 382 During the RS, the $PRES$ became the main factor in the $AKEni$ budget. It changed from source to
 383 sink in the 30-200 m layer, because of less downward KEni flux from the upper layer. In the deeper



384 layer, the negative *PRES* indicated that there was a near-inertial wave propagating away from this
385 location. The positive *AK_{Eni}* flux provided by *NL_h* weakened the effect of the negative *PRES*.

386 During the FS at A2 on the northern side of the jet stream, *VVISC* in the upper layer (Fig.
387 14d) had slightly larger magnitude than that at A1. However, it greatly decreased to $3 \times 10^{-3} \text{ W m}^{-2}$
388 below the ML (Fig. 14e), which suggested that the smaller vertical scale on the northern side of
389 the jet limited the deep penetration of the wind energy in this location. Compared to A1, the *NL_v*
390 was much stronger in the upper layer, and about one half of the lost energy was compensated for
391 by the *NL_h*. The *PRES* was negligible compared to that at A1 (Fig. 14d-f). During the RS, the *PRES*
392 in the ML became a notable sink after April 21 and was accompanied by a positive *NL_h* (Fig. 14d).
393 This suggests that the strong jet increased the *AJ_{Eni}* through either advection or wave propagation
394 due to *PRES* as a result of jet-NIO interaction at this station. In the deeper layer, the *PRES* provided
395 a positive *AK_{Eni}* flux. From the spectral analysis, the wave at 500 m had a large blue shift of
396 $>0.15f$ (Fig. 9f) that cannot be explained by the background vorticity alone. The wave likely
397 originated from the northern latitude.

398 In the far field at stations C1 and C2, where there was no direct wind forcing from the TC,
399 surface forcing (*VVISC*) was relatively small during the FS (Fig. 14g,j). Therefore, horizontal
400 transport of energy is needed to sustain the *K_{Eni}* intensification at these two locations (Fig. 11c,d).
401 At C1, which was on the southern side of the jet stream (Fig. 8) and had a negative background
402 vorticity, the *PRES* was the main source of *AK_{Eni}* in both subsurface and deep layers (Fig. 14h,i).
403 The existences of the blue shift near the local inertial frequency (Fig. 11e) and of the negative
404 background vorticity suggest the presence of a southward propagating near-inertial wave towards
405 C1 from northern region. Because C2 lies near the northeastward turning point of the jet (Fig. 8),
406 the nonlinear effect became significant in the 30-200 m layer where the jet was strongest (Fig.



14k). After the enhancement of *AKEni* in the subsurface layer, the *PRES* further transported the *KEni* downwards and became the major source for the increase of *AKEni* in the deep layer after April 27 (Fig. 14l). The northeast current advected the lower frequency NIO from the lower latitude towards the higher latitude, C2, which explains the red shift of the NIO at this location (Fig. 11f).

5 Summary

TCs force the ocean to form NIOs. The response of NIOs is largely associated with the different forcing stages of the TCs and background flow. Due to spatiotemporally limited measurements, our understanding of the process and mechanism that govern the NIO response is mainly based on theories that are constrained by idealized assumptions. In this study, we utilize a well-validated circulation model to investigate the characteristic response of *KEni* to a moderately strong TC (Neoguri) with observed strong *KEni* and to a unique background circulation.

The near-inertial currents in the upper layer strengthened significantly during the TC forced stage and displayed a clear rightward bias due to stronger wind forcing and the resonance between the wind and the near-inertial currents. The distribution of near-inertial currents and the associated rotary spectra showed that the propagation patterns of NIOs varied greatly from location to location and were closely linked to the influences of the background jet.

We calculated the *KEni* balance to diagnose spatiotemporally varying responses and processes of the near-inertial signals in terms of different forcing stages of the TC. Results show that during the forcing period, the vertical viscous term, which represents the wind work and entrainment at the base of the upper layer, was the *KEni* source in the upper layer. Around 0.5 IP after the maximum TC forcing, the pressure work became the main *KEni* sink in the upper layer, transporting *KEni* in the ML into the deeper layers through inertial pumping (upwelling). The



430 upwelling, caused by the TC-enhanced divergence, also advected smaller KEni from deeper layers
431 to weaken the KEni in the upper layer.

432 During the TC relaxation stage, the loss of KEni in the forced region of the whole water
433 column were caused by the vertical viscous term, the pressure work, and horizontal advection
434 effects (Table 1). However, these effects acted differently in different layers. The viscous effect
435 mainly occurred inside the water column, but decreased to near zero in the upper layer after the
436 direct impact of Neoguri. The pressure work mainly transported the KEni out of the forced region
437 horizontally and out of the upper layer vertically. It was strongest on the southern side of the jet,
438 where the negative background vorticity located. The horizontal nonlinear effect also contributed
439 greatly to the KEni balance near the jet region. It acted as a major sink of KEni by horizontally
440 advecting the NIO away from the forced region. For locations away from forced region, both near-
441 inertial wave propagation and horizontal advection contributed to the intensification of the KEni.

442 We examined the NIOs processes and underlying dynamics in response to different stage
443 of the TC in the semi-enclosed SCS under influence of unique and strong basin-wide circulation.
444 Unlike similar study in the SCS, this study enriches our understanding of the spatiotemporal
445 variability of TC-induced NIOs and provides a useful physical guidance for future process-
446 oriented field experiment in the SCS as well as in other subtropical marginal seas that are
447 frequently affected by the TC.

448

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455



References

- Alford, M. H., 2003: Improved global maps and 54-year history of wind-work on ocean inertial motions. *Geophys. Res. Lett.*, 30 (8), 1424, doi:10.1029/2002GL016614.
- Alford, M. H., M. F. Cronin, and J. M. Klymak, 2012: Annual cycle and depth penetration of wind-generated near-inertial internal waves at ocean station papa in the northeast pacific. *J. Phys. Oceanogr.*, 42, 889-908.
- Alford, M.H., MacKinnon, J.A., Simmons, H.L. and Nash, J.D., 2016: Near-inertial internal gravity waves in the ocean. *Annual review of marine science*, 8(1), 95-123.
- Antonov, J. I., R. A. Locarnini, T. P. Boyer, A. V. Mishonov, and H. E. Garcia, 2006. *World Ocean Atlas 2005, Volume 2: Salinity*. S. Levitus, Ed. NOAA Atlas NESDIS 62, U.S. Government Printing Office, Washington, D.C., 182 pp.
- Atlas, R., R. N. Homan, J. Ardizzone, S. M. Leidner, J. C. Jusem, D. K. Smith, and D. Gombos, 2011: A cross-calibrated, multiplatform ocean surface wind velocity product for meteorological and oceanographic applications. *Bull. Amer. Meteor. Soc.*, 92, 157-174, doi:10.1175/2010BAMS2946.1.
- Brink, K. H., 1989: Observations of the response of thermocline currents to a hurricane. *J. Phys. Oceanogr.*, 19, 1017-1022.
- Chen, G., J. Gan, Q. Xie, X. Chu, D. Wang, and Y. Hou, 2012. Eddy heat and salt transports in the South China Sea and their seasonal modulations, *J. Geophys. Res.*, 117, C05021, doi:10.1029/2011JC007724.
- Chu, P. C., J. M. Veneziano, C. Fan, M. J. Carron, and W. T. Liu, 2000: Response of the south china sea to tropical cyclone ernie 1996. *J. Geophys. Res.*, 105 (C6), 13 991-14 009.



- 479 D'Asaro, E. A., 1989: The decay of wind-forced mixed layer inertial oscillations due to the β -
 480 effect. *J. Geophys. Res.*, 94 (C2), 2045-2056.
- 481 D'Asaro, E. A., C. C. Eriksen, M. D. Levine, P. Niiler, C. A. Paulson, and P. V. Meurs, 1995:
 482 Upper-ocean inertial currents forced by a strong storm. part I: Data and comparisons with linear
 483 theory. *J. Phys. Oceanogr.*, 25, 2909-2937.
- 484 Danioux, E., P. Klein, and P. Riviere, 2008: Propagation of Wind Energy into the Deep Ocean
 485 through Mesoscale Eddy Field. *J. Phys. Oceanogr.*, 38, 2224-2241.
- 486 Danioux, E., Vanneste, J. and Bühler, O., 2015: On the concentration of near-inertial waves in
 487 anticyclones. *Journal of Fluid Mechanics*, 773.
- 488 Davies, A. M. and J. Xing, 2002: Influence of coastal fronts on near-inertial internal waves.
 489 *Geophys. Res. Lett.*, 29 (23), 2114, doi:10.1029/2002GL015904.
- 490 Ferrari, R., and C. Wunsch, 2009: Ocean circulation kinetic energy: Reservoirs, sources, and sinks.
 491 *Annual Review of Fluid Mechanics*, 41, 253-282.
- 492 Gan J., Z. Liu and L. Liang, 2016a: Numerical modeling of intrinsically and extrinsically forced
 493 seasonal circulation in the China Seas: A kinematic study, *J. Geophys. Res. (Oceans)*, 121 (7),
 494 4697-4715, doi: 10.1002/2016JC011800.
- 495 Gan, J., Z. Liu and C. Hui, 2016b: A three-layer alternating spinning circulation in the South China
 496 Sea, *J. Phys. Oceanogr.* doi:10.1175/JPO-D-16-0044.
- 497 Gan, J. and J. S. Allen, 2005: On open boundary conditions for a limited-area coastal model off
 498 Oregon. part 1: Response to idealized wind forcing. *Ocean Modelling*, 8, 115-133.
- 499 Gan, J., H. Li, E. N. Curchister, and D. B. Haidvogel, 2006: Modeling South China Sea circulation:
 500 Response to seasonal forcing regimes. *J. Geophys. Res.*, 111 (C06034),
 501 doi:10.1029/2005JC003298.



- 502 Gan, J., and T. Qu, 2008: Coastal jet separation and associated flow variability in the southwest
 503 South China Sea. *Deep-Sea Res. I*, doi:10.1016/j.dsr. 2007.09.008.
- 504 Garrett, C., 2001: What is the “near-inertial” band and why is it different from the rest of the
 505 internal wave spectrum? *J. Phys. Oceanogr.*, 31, 962-971.
- 506 Geisler, J.E., 1970: Linear theory of the response of a two layer ocean to a moving hurricane.
 507 *Geophysical and Astrophysical Fluid Dynamics*, 1, 249-272.
- 508 Gill, A. E., 1984: On the behavior of internal waves in the wakes of storms. *J. Phys. Oceanogr.*,
 509 14, 1129-1151.
- 510 Gonella, J., 1972: A rotary-component method for analysing meteorological and oceanographic
 511 vector time series. *Deep-Sea Research*, 19, 833-846.
- 512 Gregg, M. C., E. A. D'Asaro, T. J. Shay, and N. Larson, 1986: Observations of persistent mixing
 513 and near-inertial internal waves. *J. Phys. Oceanogr.*, 16, 856-885, 21.
- 514 Hebert, D. and J. N. Moum, 1994: Decay of a near-inertial wave. *J. Phys. Oceanogr.*, 24, 2334-
 515 2351.
- 516 Jordi, A. and D.-P. Wang, 2008: Near-inertial motions in and around the Palamos submarine
 517 canyon (NW Mediterranean) generated by a severe storm. *Continental Shelf Research*, 28, 2523-
 518 2534.
- 519 Kundu, P. K. and I. M. Cohen, 2008: *Fluid Mechanics*, 872 pp. Academic, San Diego.
- 520 Kunze, E., 1985: Nearinertial wave propagation in geostrophic shear. *J. Phys. Oceanogr.*, 15, 544-
 521 565.
- 522 Liu, L. L., W. Wang, and R. X. Huang, 2008: The mechanical energy input to the ocean induced
 523 by tropical cyclones. *J. Phys. Oceanogr.*, 38, 1253-1266.



524 Locarnini, R. A., A. V. Mishonov, J. I. Antonov, T. P. Boyer, and H. E. Garcia, 2006. World Ocean
 525 Atlas 2005, Volume 1: Temperature. S. Levitus, Ed. NOAA Atlas NESDIS 61, U.S. Government
 526 Printing Office, Washington, D.C., 182 pp.

527 Mooers, C. N. K., 1973: A technique for the cross spectrum analysis of pairs of complex- valued
 528 time series, with emphasis on properties of polarized components and rotational invariants. Deep-
 529 Sea Research, 20, 1129-1141.

530 Morozov, E. G. and M. G. Velarde, 2008: Inertial oscillations as deep ocean response to hurricanes.
 531 Journal of Oceanography, 64(4), 495-509, doi:10.1007/s10872-008-0042-0.

532 Nagai, T., Tandon, A., Kunze, E. and Mahadevan, A., 2015: Spontaneous generation of near-
 533 inertial waves by the Kuroshio Front. J. Phys. Oceanogr., 45, 2381-2406.

534 Price, J., 1981: Upper ocean response to a hurricane. J. Phys. Oceanogr., 11(2), 153-175.

535 Qi, H., R. A. de Szoeke, and C. A. Paulson, 1995: The structure of near-inertial waves during
 536 ocean storms. J. Phys. Oceanogr., 25, 2853-2871. 22.

537 Rocha, C.B., Wagner, G.L. and Young, W.R., 2018: Stimulated generation: extraction of energy
 538 from balanced flow by near-inertial waves. Journal of Fluid Mechanics, 847, 417-451.

539 Sasaki, H., M. Nonaka, Y. Masumoto, Y. Sasai, H. Uehara, and H. Sakuma, 2008: An eddy-
 540 resolving hindcast simulation of the quasiglobal ocean from 1950 to 2003 on the Earth Simulator.
 541 In High Resolution Numerical Modelling of the Atmosphere and Ocean, K. Hamilton and
 542 W. Ohfuchi (eds.), chapter 10, pp. 157–185, Springer, New York.

543 Shay, L. K. and R. L. Elsberry, 1987: Near-inertial ocean current responses to hurricane Frederic.
 544 J. Phys. Oceanogr., 17, 1249-1269.



545 Shchepetkin, A. F. and J. C. McWilliams, 2005: The regional ocean modeling system: A split-
 546 explicit, free-surface, topography following coordinates ocean model. *Ocean Modelling*, 9, 347-
 547 404.

548 Silverthorne, K. E. and J. M. Toole, 2009: Seasonal kinetic energy variability of near-inertial
 549 motions. *J. Phys. Oceanogr.*, 39, 1035-1049.

550 Simmons, H. L. and M. H. Alford, 2012: Simulating the long-range swell of internal waves
 551 generated by ocean storms. *Oceanography*, 25 (2), 30-41, doi:[http://dx.doi.org/10.5670/](http://dx.doi.org/10.5670/oceanog.2012.39)
 552 [oceanog.2012.39](http://dx.doi.org/10.5670/oceanog.2012.39).

553 Sun, L., Q. Zheng, D. Wang, J. Hu, C.-K. Tai, and Z. Sun, 2011a: A case study of near-inertial
 554 oscillation in the south china sea using mooring observations and satellite altimeter data. *Journal*
 555 *of Oceanography*, 67, 677-687, doi:10.1007/s10872-011-0081-9.

556 Sun, Z., J. Hu, Q. Zheng, and C. Li, 2011b: Strong near-inertial oscillations in geostrophic shear
 557 in the northern south china sea. *Journal of Oceanography*, 67, 377-384, doi:10.1007/s10872-011-
 558 0038-z.

559 Thomas, L. N., 2019: Enhanced radiation of near-inertial energy by frontal vertical circulations.
 560 *J. Phys. Oceanogr.*, 49, 2407-2421.

561 Wang, G., J. Su, Y. Ding, and D. Chen, 2007: Tropical cyclone genesis over the South China Sea.
 562 *Journal of Marine Systems*, 68, 318–326.

563 Watanabe, M. and T. Hibiya, 2002: Global estimate of the wind-induced energy flux to the inertial
 564 motion in the surface mixed layer. *Geophys. Res. Lett.*, 29, 1239, doi:10.1029/2001GL04422. 23.

565 Whitt, D.B. and Thomas, L.N., 2013. Near-inertial waves in strongly baroclinic currents. *J. Phys.*
 566 *Oceanogr.*, 43, 706-725.



567 Xu, Z., B. Yin, Y. Hou, and Y. Xu, 2013: Variability of internal tides and near-inertial waves on
568 the continental slope of the northwestern South China Sea. *J. Geophys. Res.*, 118, 1-15,
569 doi:10.1029/2012JC008212.

570 Young, W. R. and M. B. Jelloul, 1997: Propagation of near-inertial oscillations through a
571 geostrophic flow. *Journal of Marine Research*, 55, 735-766.

572 Zedler, S. E., 2009: Simulations of the ocean response to a hurricane: Nonlinear processes. *J. Phys.*
573 *Oceanogr.*, 39, 2618-2634.

574 Zhai, X., R. J. Greatbatch, and J. Sheng, 2004: Advective spreading of storm-induced inertial
575 oscillations in a model of the northwest Atlantic Ocean. *Geophys. Res. Lett.*, 31 (L14315),
576 doi:10.1029/2004GL020084.



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Table 1. Time integrated KEni budget (unit: $\times 10^3 \text{ J m}^{-2}$) during the FS and RS.



List of Figures

Figure 1 (a) Track of typhoon Neoguri (2008) from JTWC; blue triangles represent ARGO drifter locations and the red circle represents Wenchang where there were ADCP observations. The date is marked by month/day along the track. TS: tropical storm, STS: strong tropical storm, T: typhoon, ST: strong typhoon, and Super T: super typhoon. (b) translation speed (U_h , unit: m s^{-1}) and the 1st baroclinic wave speed (C_1 , unit: m s^{-1}) along the TC track; (c) clockwise (A_{cw} , green line) and counter-clockwise (A_{ccw} , blue line) rotary current amplitude (m s^{-1}) from current measurement at Wenchang. The x-axis is time marked as month/day.

Figure 2 Daily mean KE (J m^{-3} , color contour) and current vectors (arrows) at 10 m (a) on April 14 of the pre-storm stage (PS), (b) on April 18 during the strongest wind forcing of the forced stage (FS), (c) on April 20 after the end of the FS, and (d) on April 30 during the relaxation stage (RS). The grey contours are the 200 m, 500 m, and 1000 m isobaths. The magenta line represents TC track. Yellow triangle on April 18 represents the TC location. The TC was located beyond the plotting domain during the other three days, as shown in Figure 1a. The velocity magnitudes $< 0.2 \text{ ms}^{-1}$ are not shown in the vectors.

Figure 3: ΔSST ($\text{SST}_{\text{April 19}} - \text{SST}_{\text{April 14}}$) from (a) model results and (b) GHRSSST JPL MUR satellite products. The pink curve refers to the trajectory of the TC Neoguri.

Figure 4 Rotary spectra of clockwise component (upper 10 m) at Wenchang (112°E , 19.6°N) from model simulations (red) and observations (blue).

Figure 5 Time series, represented by color bar, of daily (a) clockwise and (b) counter-clockwise rotary current vectors from April 14 to 30 during different stages of the TC forcing, signifying the response of the current to the local wind rotation. For the clockwise (counter-clockwise)



component, only currents with magnitude larger than 0.2 (0.05) m s^{-1} are shown. The black box represents the forced region.

Figure 6 Time series of 6-hourly wind stress vectors during the forced-stage (FS) from April 15-20.

Figure 7 Time series of (a) the area-averaged wind energy flux into the near-inertial band (unit: 10^{-3} W m^{-2}) and (b) depth-integrated KE_{ni} (J m^{-2}) in the forced region for different layers.

Figure 8 Daily averaged KE_{ni} (KJ m^{-2}) of layers (a, b) 30-200 m and (c,d) below 200 m on (a, c) April 20 during FS, and (b,d) April 30 during RS. The thick red arrows show the location of the jet stream (Fig. 2), while the orange curve arrows indicate regions with relative vorticity $\zeta > 0$, and the blue curve arrows indicate regions with $\zeta < 0$ induced by the jet. Stations A1 and A2 are on the right side of the TC track at the northern (A2) and southern (A1) sides of the jet, respectively. Stations C1 and C2 are corresponding stations in the far field. Station B is located in the upstream of the jet stream.

Figure 9 Time series of (a, b) u_i (m s^{-1}), (c, d) KE_{ni} (J m^{-2}), and (e, f) rotary spectra (cw component) at locations A1 (a,c,e) and A2 (b,d,f).

Figure 10 Time-averaged N^2 (s^{-2}) from April 15 to May 5 at locations A1 (red) and A2 (blue).

Figure 11 As in Fig. 9, except for locations C1 (a, c, e) and C2 (b, d, f).

Figure 12 Time series of area-averaged, depth-integrated KE_{ni} budget for (a) 0-30 m, (b) 30-200 m, and (c) >200 m in the forced region. Terms represent (unit: $\times 10^{-3} \text{ W m}^{-2}$): (a-c) vertical viscous effect (VVISC), (d-f) divergence of energy flux (PRES), (g-i) horizontal non-linear interaction (NL_h), and (j-l) vertical non-linear interaction (NL_v). The vertical lines separate the pre-storm stage, FS and RS during the TC forcing.



Figure 13 Horizontal distribution of time-averaged (April 15-May 5) depth-integrated KEni budget in different layers: 0-30 m (left column), 30-200 m (middle), and >200 m (right). The terms re

Figure 14 Time series of KEni budget at locations: A1 (a-c), A2 (d-f), C1 (g-i), and C2 (j-l) in layers: 0-30 m (left column), 30-200 m (middle column), and >200 m (right column).

presented are (unit: $\times 10^{-3} \text{ W m}^{-2}$): (a-c) VVISC, (d-f) PRES, (g-i) NL_h , and (j-l) NL_v .



<i>TERM</i>	<i>RATE</i>		<i>VVISC</i>		<i>PRES</i>		<i>NLh</i>		<i>NLv</i>	
Phase	FS	RS	FS	RS	FS	RS	FS	RS	FS	RS
0-30 m	1.15	-0.97	2.15	0.00	-0.65	-0.85	-0.05	-0.10	-0.29	-0.01
30-200m	1.11	-0.44	0.96	-0.19	0.30	-0.16	0.21	-0.18	-0.35	0.10
>200 m	0.23	0.36	-0.02	-0.09	0.26	0.59	0.00	-0.07	-0.01	-0.07
Column	2.51	-1.02	3.10	-0.27	-0.09	-0.42	0.16	-0.35	-0.66	0.02

Table 1. Time integrated KEni budget (unit: $\times 10^3 \text{ J m}^{-2}$) during the FS and RS.

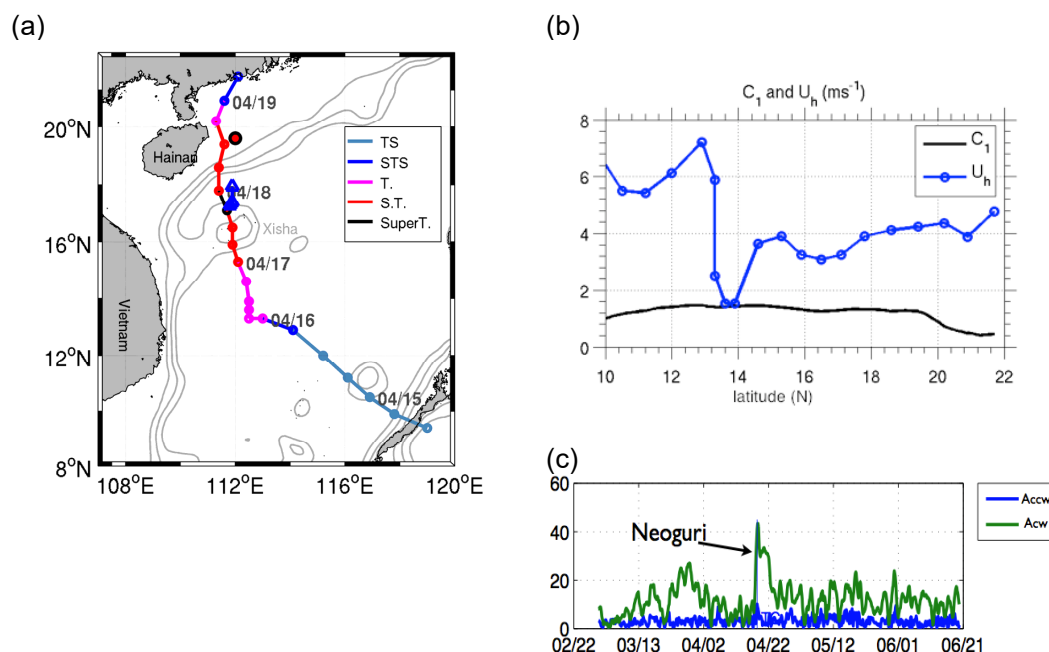


Figure 1 (a) Track of typhoon Neoguri (2008) from JTWC; blue triangles represent ARGO drifter locations and the red circle represents Wenchang where there were ADCP observations. The date is marked by month/day along the track. TS: tropical storm, STS: strong tropical storm, T: typhoon, ST: strong typhoon, and Super T: super typhoon. (b) translation speed (U_h , unit: m s^{-1}) and the 1st baroclinic wave speed (C_1 , unit: m s^{-1}) along the TC track; (c) clockwise (A_{cw} , green line) and counter-clockwise (A_{ccw} , blue line) rotary current amplitude (m s^{-1}) from current measurement at Wenchang. The x -axis is time marked as month/day.

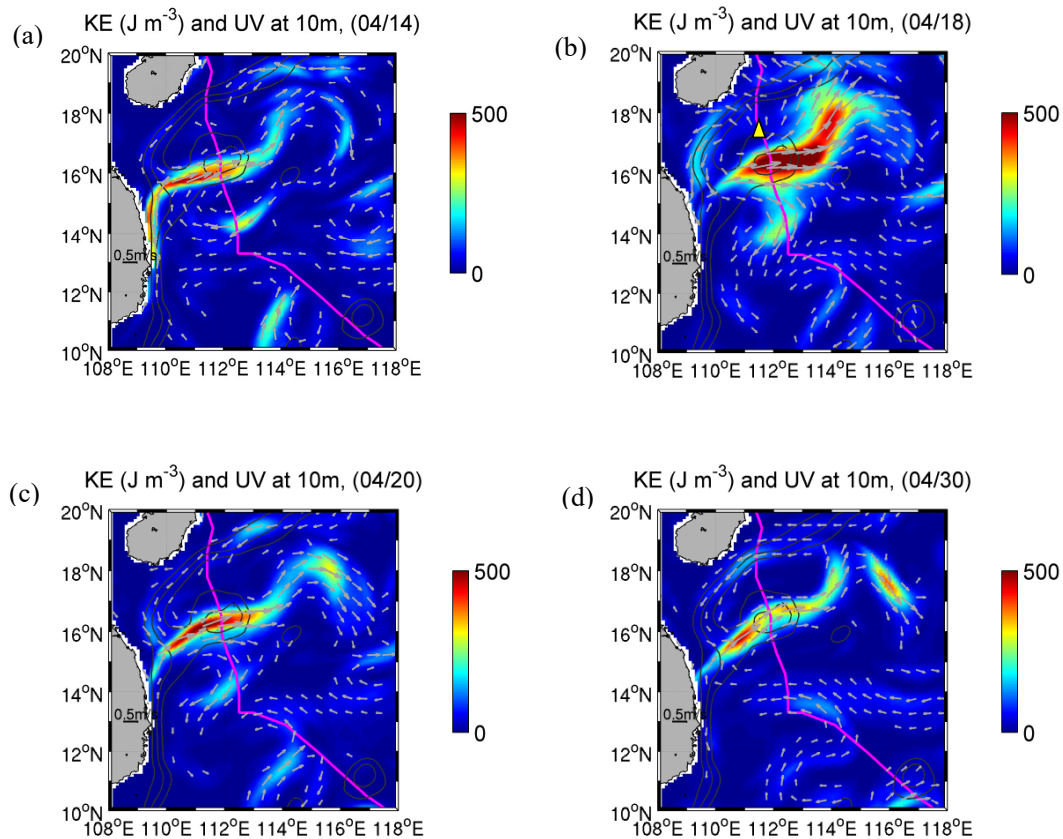


Figure 2 Daily mean KE (J m^{-3} , color contour) and current vectors (arrows) at 10 m (a) on April 14 of the pre-storm stage (PS), (b) on April 18 during the strongest wind forcing of the forced stage (FS), (c) on April 20 after the end of the FS, and (d) on April 30 during the relaxation stage (RS). The grey contours are the 200 m, 500 m, and 1000 m isobaths. The magenta line represents TC track. Yellow triangle on April 18 represents the TC location. The TC was located beyond the plotting domain during the other three days, as shown in Figure 1a. The velocity magnitudes $< 0.2 \text{ ms}^{-1}$ are not shown in the vectors.

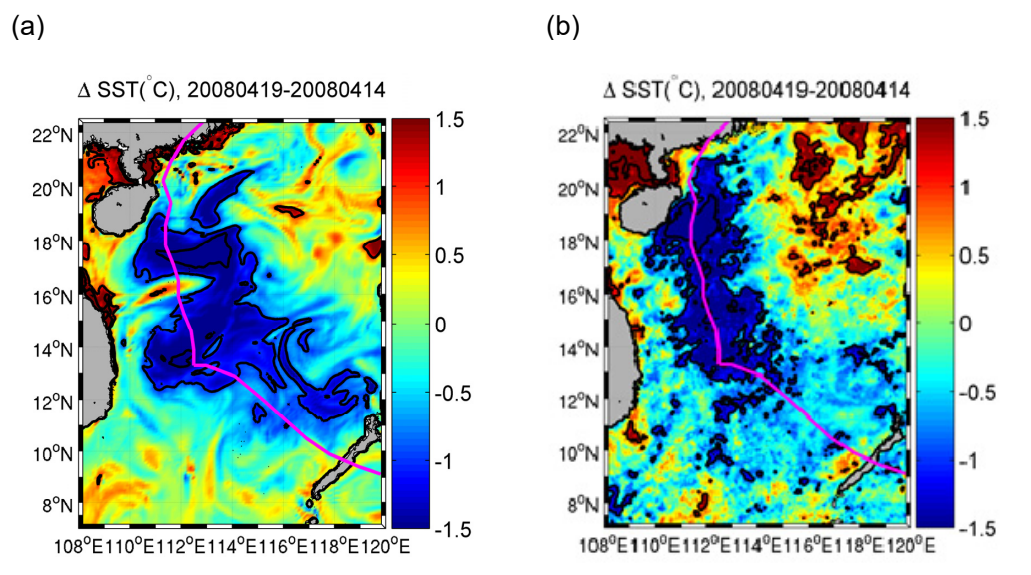


Figure 3 ΔSST ($\text{SST}_{\text{April 19}} - \text{SST}_{\text{April 14}}$) from (a) model results and (b) GHRSSST JPL MUR satellite products. The pink curve refers to the trajectory of the TC Neoguri.

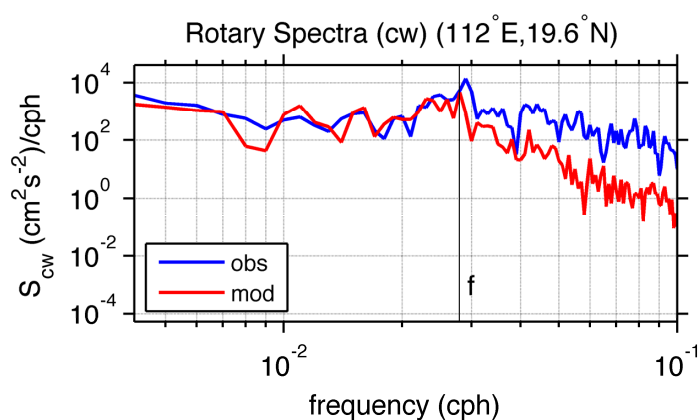


Figure 4 Rotary spectra of clockwise component (upper 10 m) at Wenchang (112°E, 19.6°N) from model simulations (red) and observations (blue).

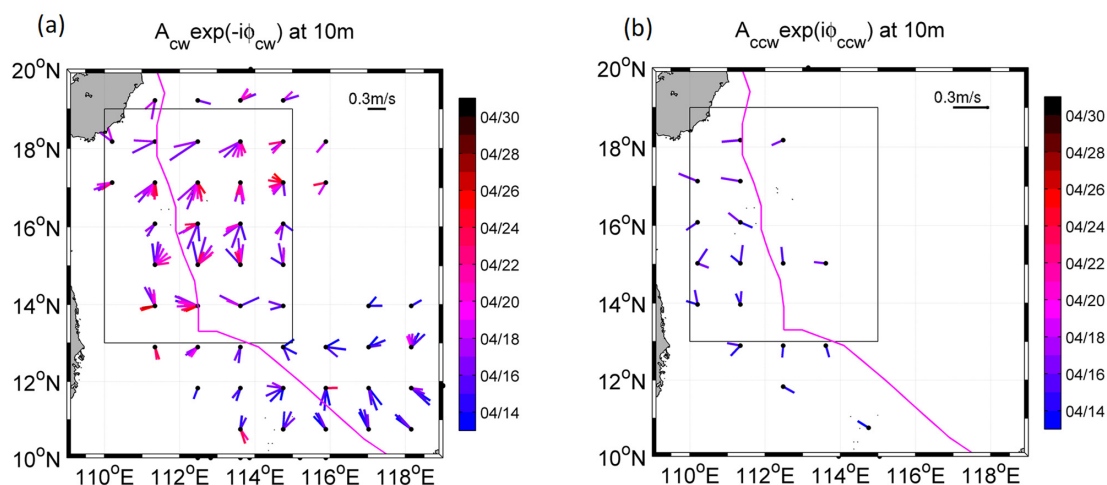


Figure 5 Time series, represented by color bar, of daily (a) clockwise and (b) counter-clockwise rotary current vectors from April 14 to 30 during different stages of the TC forcing, signifying the response of the current to the local wind rotation. For the clockwise (counter-clockwise) component, only currents with magnitude larger than 0.2 (0.05) m s^{-1} are shown. The black box represents the forced region.

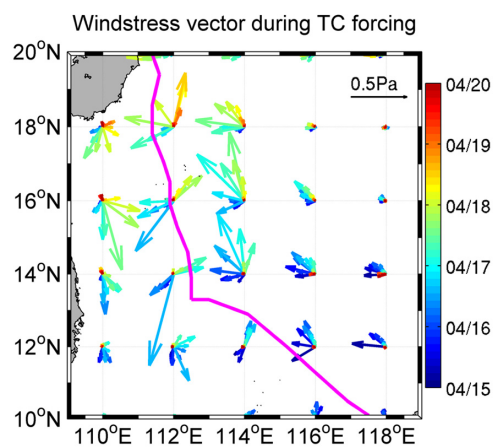


Figure 6 Time series of 6-hourly wind stress vectors during the forced-stage (FS) from April 15-20.

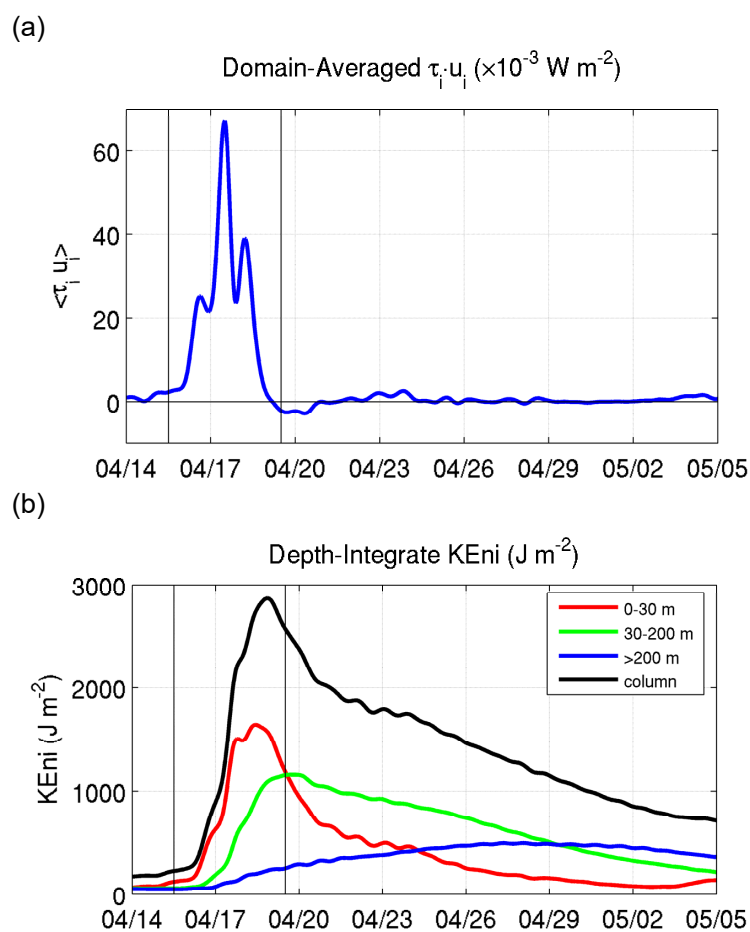


Figure 7 Time series of (a) the area-averaged wind energy flux into the near-inertial band (unit: 10^{-3} W m^{-2}) and (b) depth-integrated KEni (J m^{-2}) in the forced region for different layers.

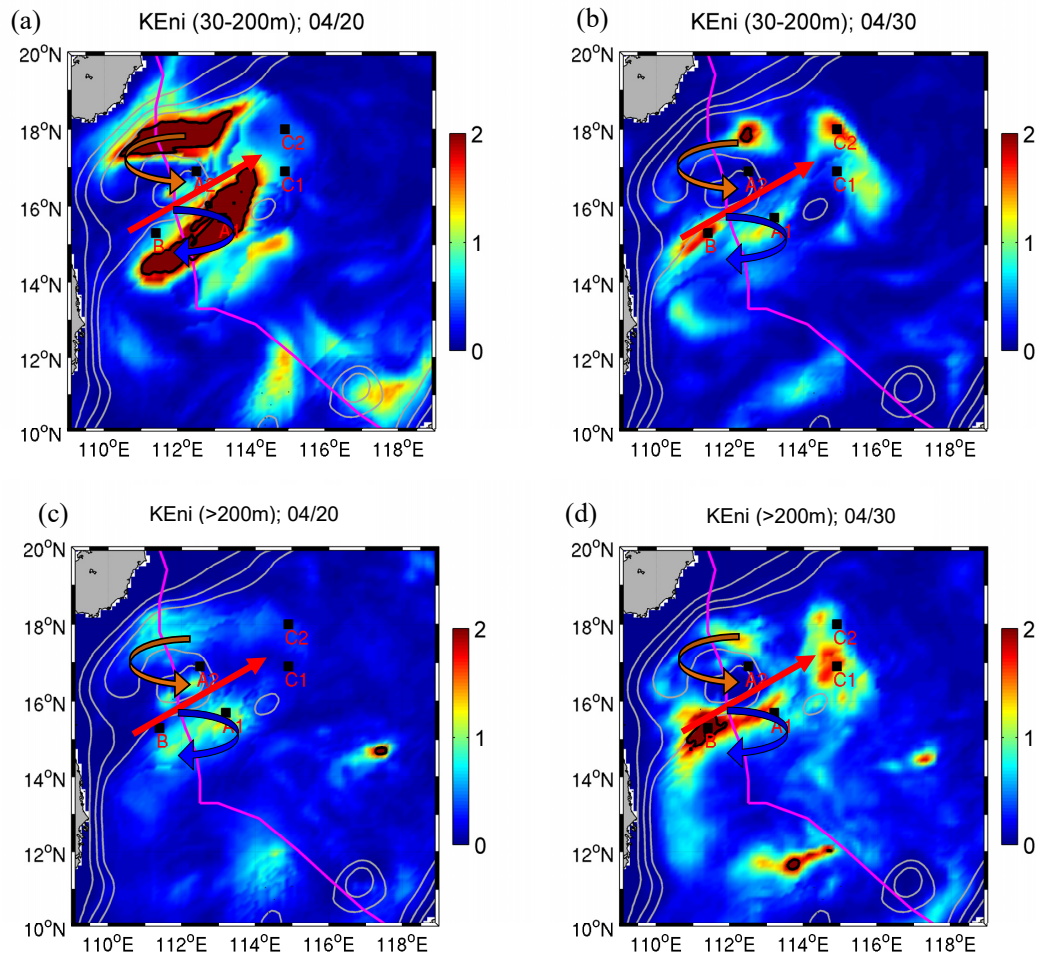


Figure 8 Daily averaged KEni (KJ m^{-2}) of layers (a, b) 30-200 m and (c,d) below 200 m on (a, c) April 20 during FS, and (b,d) April 30 during RS. The thick red arrows show the location of the jet stream (Fig. 2), while the orange curve arrows indicate regions with relative vorticity $\zeta > 0$, and the blue curve arrows indicate regions with $\zeta < 0$ induced by the jet. Stations A1 and A2 are on the right side of the TC track at the northern (A2) and southern (A1) sides of the jet, respectively. Stations C1 and C2 are corresponding stations in the far field. Station B is located in the upstream of the jet stream.

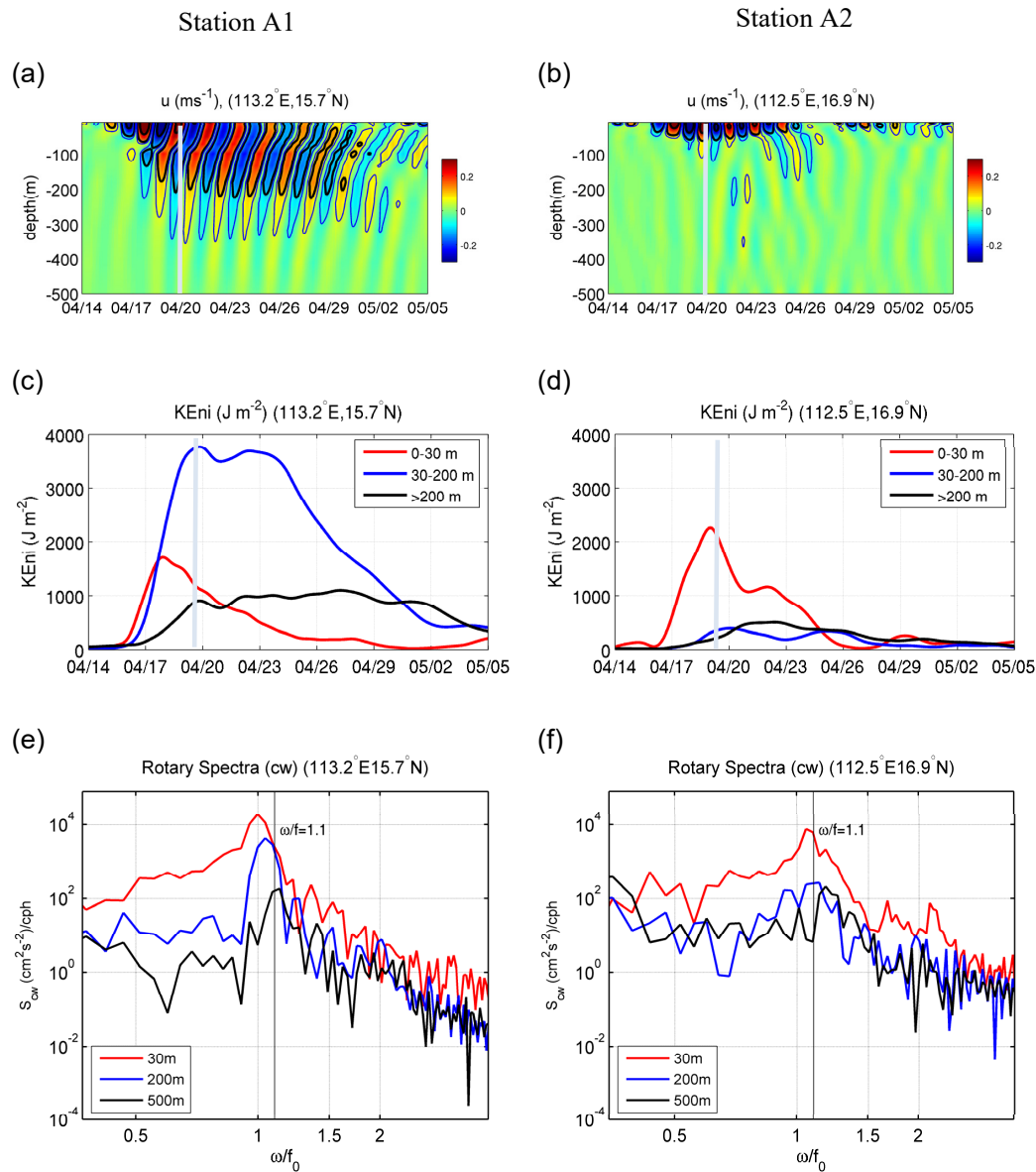


Figure 9 Time series of (a, b) u_i (m s⁻¹), (c, d) KE_{ni} (J m⁻²), and (e, f) rotary spectra (cw component) at locations A1 (a,c,e) and A2 (b,d,f).

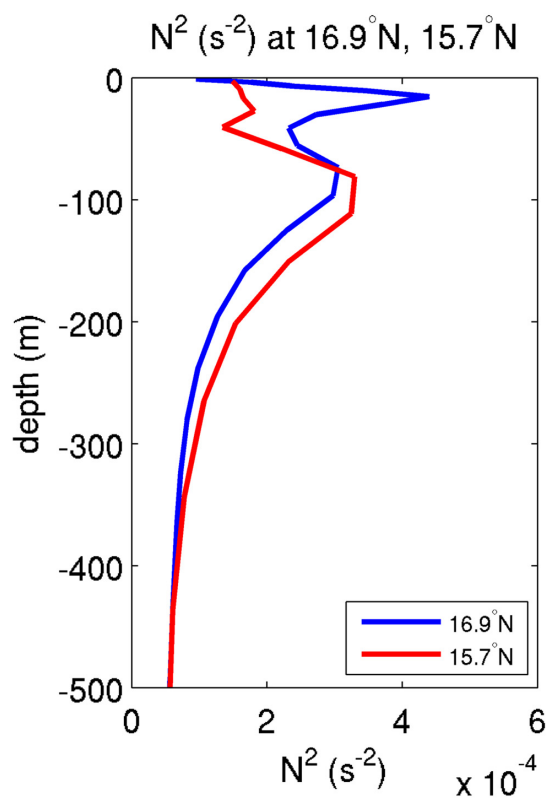


Figure 10 Time-averaged N^2 (s^{-2}) from April 15 to May 5 at locations A1 (red) and A2 (blue).

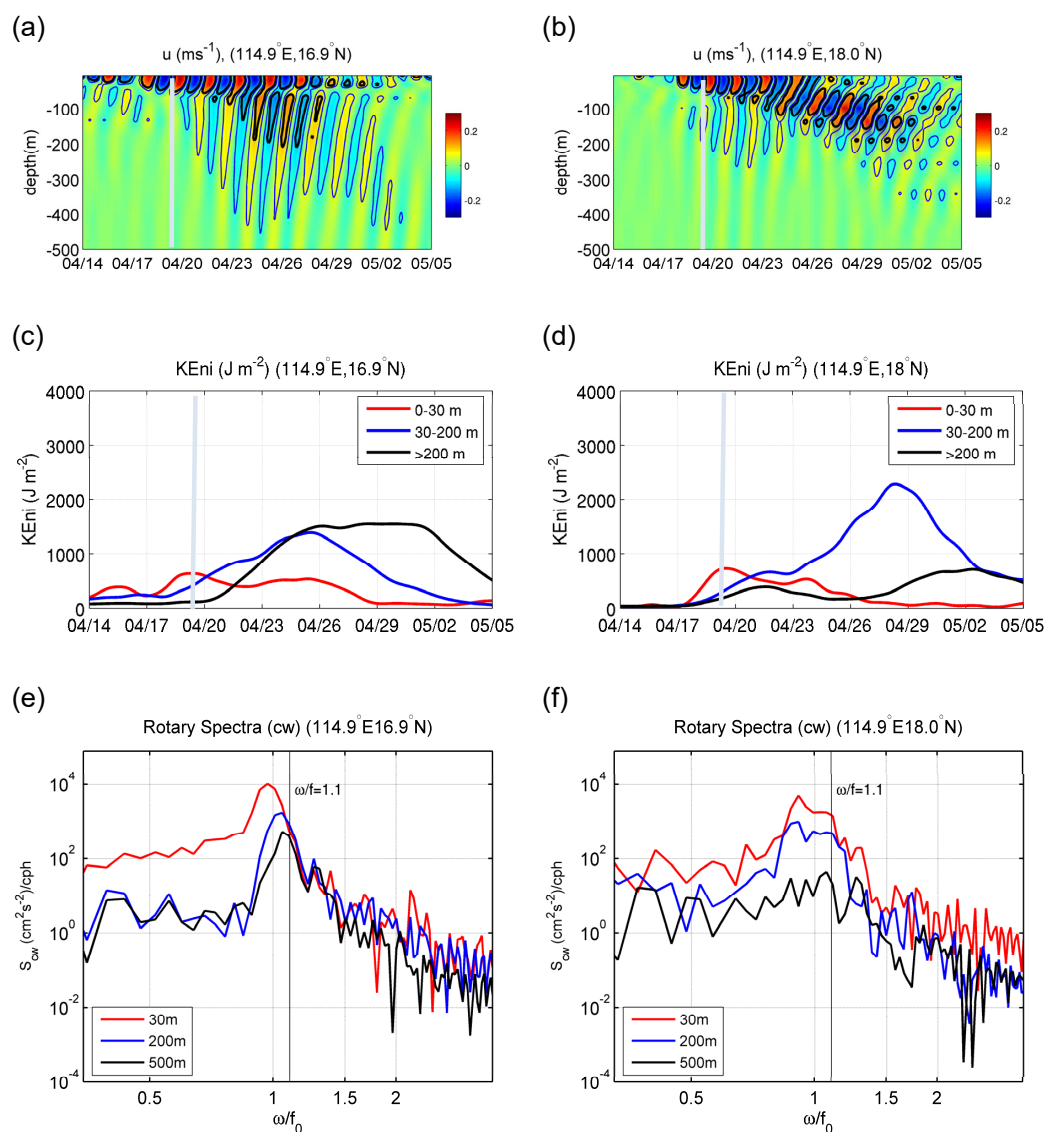


Figure 11 As in Fig. 9, except for locations C1 (a, c, e) and C2 (b, d, f).

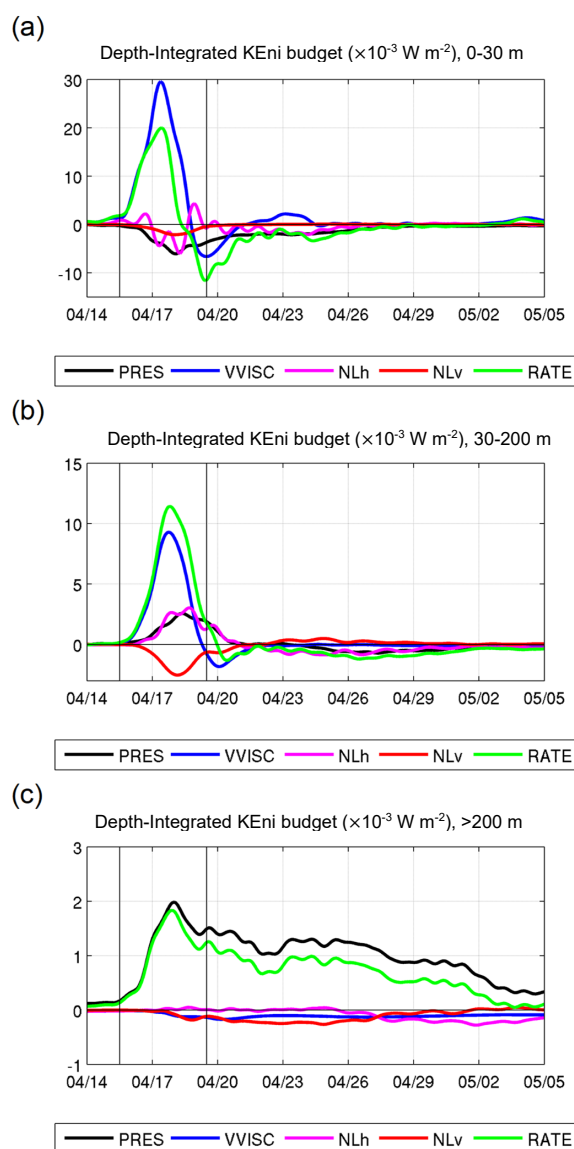


Figure 12 Time series of area-averaged, depth-integrated KEni budget for (a) 0-30 m, (b) 30-200 m, and (c) >200 m in the forced region. Terms represent (unit: $\times 10^{-3} \text{ W m}^{-2}$): (a-c) vertical viscous effect (VVISC), (d-f) divergence of energy flux (PRES), (g-i) horizontal non-linear interaction (NL_h), and (j-l) vertical non-linear interaction (NL_v). The vertical lines separate the pre-storm stage, FS and RS during the TC forcing.

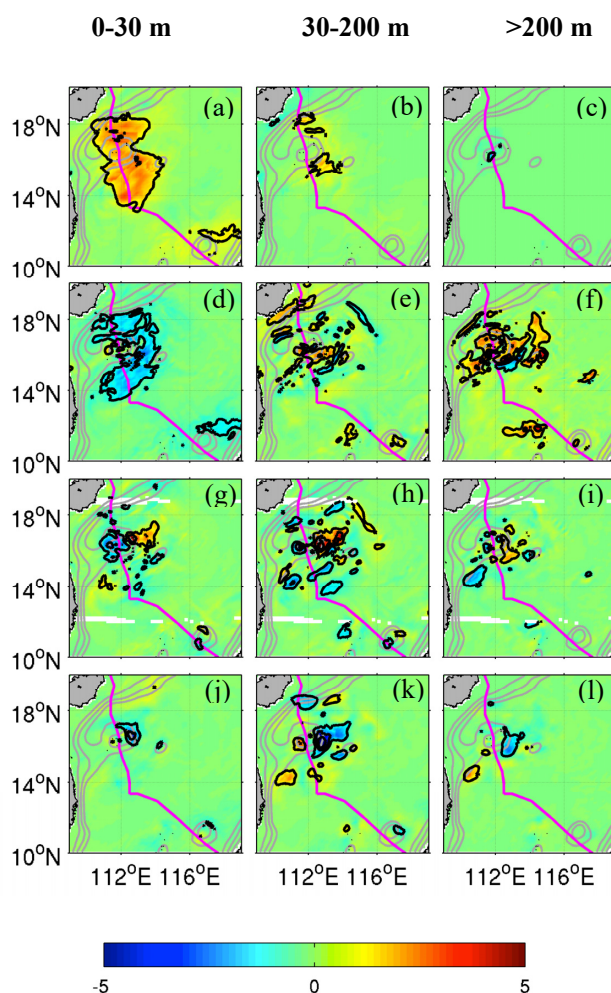


Figure 13 Horizontal distribution of time-averaged (April 15-May 5) depth-integrated KEni budget in different layers: 0-30 m (left column), 30-200 m (middle), and >200 m (right). The terms represented are (unit: $\times 10^{-3} \text{ W m}^{-2}$): (a-c) *VVISC*, (d-f) *PRES*, (g-i) *NL_h*, and (j-l) *NL_v*.

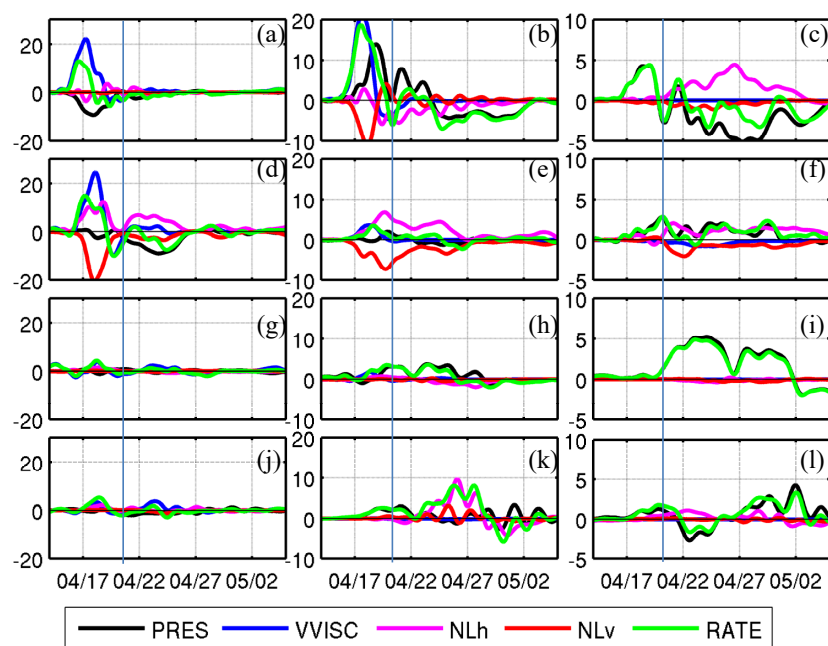


Figure 14 Time series of KEni budget at locations: A1 (a-c), A2 (d-f), C1 (g-i), and C2 (j-l) in layers: 0-30 m (left column), 30-200 m (middle column), and >200 m (right column).