

Scale-Dependent Budget and Vertical Pathway of Kinetic Energy in the Northwest Pacific

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ABSTRACT: The oceanic energy budget across mesoscale and submesoscale plays a crucial role in regulating large-scale ocean circulation and air–sea interactions. Although submesoscale baroclinic instability is known to inject energy near the surface and fuel an inverse energy cascade toward mesoscale eddies, the vertical pathway through which this energy is redistributed remains poorly understood. Using a submesoscale-permitting ocean general circulation model (OGCM) and a coarse-graining approach, we examine the spatial structure and scale dependence of the three-dimensional kinetic energy (KE) budget in the northwest Pacific, where energetic submesoscale motions exist. At submesoscales (~ 10 km), the vertical pressure work (PW) drives a substantial downward transport of KE from the mixed layer into the ocean interior, with a magnitude comparable to the baroclinic instability in enhancing submesoscale motions beneath the mixed layer. In the meanwhile, KE is gradually transferred upscale, except near intense surface fronts where forward energy cascade occurs. As the horizontal scale increases to mesoscale (~ 100 km), a considerable amount of KE converted via baroclinic instability is transferred upscale via inverse energy cascade, while the vertical PW continues to facilitate a downward energy transport toward greater depth, consistent with the barotropization tendency of mesoscale eddies. The results of this study reveal a tight coupling between the cross-scale energy cascade and vertical energy redistribution. Particularly, as submesoscale energy cascades to larger scales, it also penetrates deeper through PW, ultimately feeding interior mesoscale motions while reducing energy dissipation near the surface. This mechanism highlights the previously underappreciated role of PW in the evolution of the three-dimensional structure of submesoscale eddies.

KEYWORDS: Eddies; Energy transport; Ocean dynamics; Ocean models

1. Introduction

Ocean circulations involve the multiscale exchange of energy from basin scales to dissipation scales. The energy pathway across these spatial scales determines the spatial structure and temporal variability of circulations and impacts Earth climate (Ferrari and Wunsch 2009). At the heart of these energy processes lie the energetics of mesoscale and submesoscale eddies, which control how the energy injected at instability scales feeds back to large-scale circulations and how it is ultimately dissipated (Capet et al. 2008; Tulloch et al. 2011; McWilliams 2016). Understanding the energy budget and cascade at mesoscale and submesoscale is, therefore, essential for closing the oceanic energy budget and improving the representation of these processes in ocean climate models.

The mesoscale and submesoscale energy budgets reflect the combined effects of external forcing, instabilities, and interactions

among large-scale circulations, eddies, and turbulence. Wind forcing primarily injects energy at large scales, where it is stored mainly as available potential energy (Gill et al. 1974; Wunsch and Ferrari 2004). Through barotropic and baroclinic instabilities, the large-scale potential and kinetic energy (KE) are converted into eddy KE at submesoscale and mesoscale (e.g., Chen et al. 2014; Qiu et al. 2014; Sasaki et al. 2014, 2020a). A substantial fraction of this energy undergoes an inverse energy cascade toward larger horizontal scales and is ultimately dissipated by surface and bottom friction (Scott and Wang 2005; Scott and Arbic 2007; Schubert et al. 2020; Renault et al. 2016, 2023; Contreras et al. 2023). Another fraction of submesoscale energy cascades toward smaller scales through frontogenesis and wave–eddy interactions, where it is ultimately dissipated by turbulent processes (Barkan et al. 2017, 2021, 2024; Brüggemann and Eden 2015; Srinivasan et al. 2023). Via these processes, KE is redistributed by both spatial- and scale-dependent transfers that bridge the energy sources and sinks and maintain an overall closed energy budget.

To quantify eddy energetics in the ocean, a traditional way is to employ the Reynolds decomposition. By separating the flow into the mean and eddy components, one can derive the energy budget equations for them and thereby examine the contribution of budget terms (e.g., von Storch et al. 2012; Chen et al. 2014; Kang and Curchitser 2015). However, the scale dependence of

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energy budget terms remains unclear, since all the perturbations with a wide range of horizontal scales deviating from the mean state are lumped together as the “eddy” component. Alternatively, a scale-dependent energy budget can be obtained using the spectral approach, which diagnoses energy budget equations in wavenumber space via the Fourier transform (e.g., Capet et al. 2008; Wang et al. 2019; Dong et al. 2020). The spectral KE flux, obtained by integrating the nonlinear terms from a prescribed wavenumber to the largest wavenumber, quantifies the energy cascade across horizontal scales (Scott and Wang 2005; Klein et al. 2008; Sasaki and Klein 2012). Nevertheless, the spectral approach yields only spatially averaged results due to the limitation of the Fourier transform. Recently, a novel “coarse-graining” approach allows the computation of cross-scale KE flux on the same grid as the raw data and hence overcomes this issue (Aluie et al. 2018; Storer et al. 2022; Schubert et al. 2023; Buzzicotti et al. 2023; Wang et al. 2026). In this framework, scale-dependent energy budget equations can be derived to assess the contributions of individual budget terms for different spatial scales (Barkan et al. 2017; Aluie et al. 2018).

Despite the progress in the energetics analysis methods, most previous studies focused mainly on the horizontal pathways of energy budget since the budget terms are vertically integrated or averaged, in which the effect of vertical energy redistribution disappears (e.g., von Storch et al. 2012; Chen et al. 2014; Kang and Curchitser 2015). Consequently, the vertical pathways of energy between the surface and interior ocean remain less well understood. The vertical pressure work (PW)—associated with the vertical propagation of Rossby and gravity waves—is a key mechanism for the vertical energy redistribution in the ocean circulations and geostrophic turbulence (Charney 1971). It facilitates the vertical transfer of KE from regions of instability to depths where energy dissipation or dispersion occurs, such as surface and bottom boundary layers (Zhai and Marshall 2013; Matsuta and Masumoto 2023). In particular, pressure work transfers energy generated by baroclinic instability downward into the interior ocean in the subtropics and Antarctic Circumpolar Current, thereby playing an important role in balancing the wind energy input with bottom frictional dissipation (Zhai and Marshall 2013; Matsuta and Masumoto 2023). Vertical energy transport by pressure work also constitutes a substantial energy sink for upper-ocean KE (Zhan et al. 2016; Wang et al. 2018) and contributes to the intraseasonal variability within the ocean interior (Yang et al. 2021a). Although these studies establish PW as a primary pathway for vertical energy redistribution, their analyses have focused almost exclusively on mesoscale dynamics, largely due to limitations in model horizontal resolution. Consequently, the role and underlying mechanism of pressure work at submesoscales remain largely unexplored.

As a result, a critical question remains in our understanding of the submesoscale energy budget: To what extent does PW contribute to the vertical redistribution of submesoscale KE, and how is this process linked to the cross-scale energy cascade? This question is closely tied to the mechanisms through which submesoscale processes penetrate below the mixed layer and influence interior ocean dynamics. In this study, by employing the coarse-graining approach, we investigate the spatial structure

and scale dependence of the energy budget in the northwest Pacific using a submesoscale-permitting ocean general circulation model (OGCM). We reveal that PW plays a comparable role to baroclinic instability in enhancing submesoscale eddies below the mixed layer, where they reinforce the inverse energy cascade that transfers energy to mesoscale eddies. The vertical energy transport is closely coupled to the upscale horizontal energy cascade, reflecting the quasi-isotropic property of three-dimensional geostrophic turbulence (Charney 1971). Our results reveal a previously underappreciated vertical pathway of submesoscale energy cascade and highlight the importance of PW in the three-dimensional evolution of submesoscale eddies.

This manuscript is organized as follows: Section 2 introduces the OGCM dataset and the theoretical framework for the energy budget analysis using the coarse-graining approach. Section 3 presents the characteristics of the KE spectra. Section 4 shows the horizontal and vertical structures of the KE budget terms and their dependence on horizontal scales. Section 5 gives a more detailed analysis of the PW, focusing on the role of its vertical component in the energetics. Seasonal cycle of submesoscale KE budget is analyzed in section 6. Section 7 provides a summary and discussion of the results on understanding submesoscale energy pathways.

2. Methodology

a. OGCM

The submesoscale-permitting simulation named OFES2_NP30 was performed with the OGCM for the Earth Simulator, version 2 (OFES2; Sasaki et al. 2020b), which is based on the Modular Ocean Model, version 3 (MOM3; Pacanowski and Griffies 1999). Covering the domain of 100°E–70°W and 20°S–68°N, the model has a uniform horizontal resolution of 1/30° (~3 km) and 105 uneven vertical layers (twenty 5-m layers in the upper 100 m and 55 layers in the upper 500 m). Hence, it is capable of resolving mixed layer instability whose horizontal scale is approximately 30 km in the subtropical North Pacific (Sasaki et al. 2017). The 3-hourly atmospheric data from JRA-55 do version 1.1 (Tsujino et al. 2018) were employed to force the model. More details of OFES2_NP30 are described in Sasaki et al. (2020a). In this study, daily mean outputs including 3D currents, potential density, and sea surface height (SSH) in 2016 are used in analyses. Note that most near-inertial waves are filtered out after daily averaging; therefore, the submesoscale flows in the model data are mostly geostrophically balanced (Sasaki et al. 2020a).

This model is able to capture the spatiotemporal variation of large-scale circulations and meso/submesoscale eddies in the northwest Pacific, as demonstrated through the comparison with altimetry products and Argo observations (Sasaki et al. 2020a, 2022). Hence, we just conduct a brief validation for the results in 2016 to avoid repetition. Here, we use the multiobservation global ocean 3D temperature, salinity, height, geostrophic current, and mixed layer depth (MLD) (<https://doi.org/10.48670/moi-00052>) distributed by the Copernicus Marine and Environment Monitoring Service (CMEMS). The dataset contains 3D temperature, salinity, and geostrophic currents as well as SSH

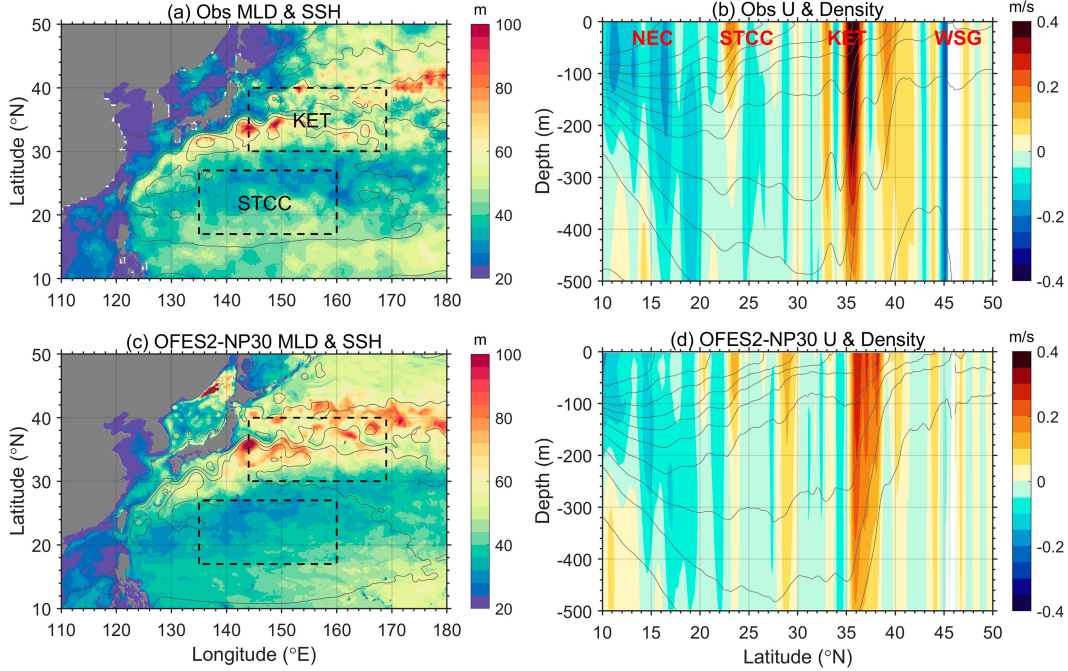


FIG. 1. Annual-mean SSH (contours with an interval of 0.25 m) and MLD (shadings; m) from (a) observations and (c) OFES2-NP30 in 2016. The two dashed rectangles indicate the KET and STCC regions. Meridional transect of zonal currents (shadings; m s^{-1}) and potential density (contours with an interval of 0.5 kg m^{-3}) along 150°E from (b) observations and (d) OFES2-NP30.

and MLD, which has a uniform horizontal resolution of 0.25° and 50 uneven layers from the surface to 5500 m. As shown in Figs. 1a and 1c, the simulated surface circulation and MLD are consistent with the observations. Notably, owing to the high resolution, the model successfully reproduces the narrow zonal band with shallow mixed layer along the Kuroshio Extension (KET), accompanied by a deep mixed layer on its northern and southern flanks (Sasaki et al. 2017; Ding et al. 2022). Vertical structure of the mean flow is examined from the meridional transect along 150°E , as shown in Figs. 1b and 1d. From south to north, there exist the North Equatorial Current (NEC), Subtropical Counter-current (STCC), KET, and Western Subarctic Gyre (WSG), with a strong meridional density gradient at the surface in the STCC and in deeper layers in the KET. Such density structures create favorable conditions for baroclinic instability and hence the generation of meso/submesoscale eddies (Qiu and Chen 2010; Sasaki et al. 2014; Qiu et al. 2014), which are also well simulated by OFES2-NP30 (Sasaki et al. 2020a, 2022).

b. Theoretical framework

In this study, we develop a scale-dependent KE budget with the coarse-graining approach (Aluie et al. 2018; Storer et al. 2022; Buzzicotti et al. 2023). In this approach, the spatially low-passed field is calculated by convolving the raw field with a given window function C :

$$\overline{A}(x, y) = A(x, y) * C(x, y), \quad (1)$$

where $*$ is the convolution operator. The top-hat window function C is expressed as

$$C(x, y) = \begin{cases} 4/(\pi L^2), & r < L/2, \\ 0, & \text{otherwise.} \end{cases} \quad (2)$$

The low-pass field includes only the signals at scales larger than L . Note that the filtering operation commutes with spatial derivations, i.e., $\nabla \overline{A} = \overline{\nabla A}$ (Aluie et al. 2018) (see Fig. S1 in the online supplemental material). Following Buzzicotti et al. (2023), the coarse and fine-scale KE are defined as

$$\text{KE}_L = \frac{1}{2} \rho_0 (\overline{u^2} + \overline{v^2}) \quad (3)$$

and

$$\text{KE}_f = \frac{1}{2} \rho_0 (\overline{u^2} - \overline{u}^2 + \overline{v^2} - \overline{v}^2), \quad (4)$$

respectively, where (u, v) is the horizontal velocity. The former represents KE at scales larger than L , while the latter contains only the KE at scales smaller than L , which is the focus of this study.

We derive the governing equation for the fine-scale KE. Starting from the primary momentum and incompressible continuity equations under the Boussinesq and hydrostatic approximations, that is,

$$\partial_t \mathbf{u}_H + \mathbf{u} \cdot \nabla \mathbf{u}_H + f \mathbf{k} \times \mathbf{u}_H = -\frac{1}{\rho_0} \nabla_H p + \mathbf{D}_H, \quad (5)$$

$$0 = -\frac{1}{\rho_0} \partial_z p - \frac{\rho g}{\rho_0}, \quad (6)$$

$$\nabla \cdot \mathbf{u} = 0, \quad (7)$$

where $\mathbf{u} = (u, v, w)$ is the 3D velocity vector, $\mathbf{u}_H = (u, v)$ is the horizontal velocity vector, f is the inertial frequency, p is the pressure, ρ is the density ($\rho_0 = 1025 \text{ kg m}^{-3}$ is the reference density), and $\mathbf{D}_H = (D_u, D_v)$ is the frictional term. Multiplying Eq. (5) by $\rho_0 \mathbf{u}_H$ and Eq. (6) by $\rho_0 w$ and summing them together, the total KE budget equation writes

$$\partial_t \text{KE} = -\nabla \cdot (\mathbf{u} \text{KE}) - \nabla \cdot (\mathbf{u} p) - \rho g w + \rho_0 \mathbf{u}_H \cdot \mathbf{D}_H, \quad (8)$$

where $\text{KE} = (1/2)\rho_0(u^2 + v^2)$ is the total KE. Terms on the right-hand side of Eq. (8) denote the tendency of total horizontal KE caused by advection, pressure work, buoyancy production, and frictional dissipation. Note that the KE budget equation can also be obtained via multiplying Eq. (5) by $\rho_0 \mathbf{u}_H$ (Aiki et al. 2016), in which the pressure term $-\mathbf{u}_H \cdot \nabla_H p$ appears on the right-hand side (see Fig. S2). It can be further rewritten as $-\nabla \cdot (\mathbf{u} p) - \rho g w$ based on Eqs. (6) and (7) (Capet et al. 2008), finally yielding Eq. (8). The reason for splitting $-\mathbf{u}_H \cdot \nabla_H p$ is that this term obscures the underlying physical processes, particularly the energy conversion between PE and KE via the vertical buoyancy flux. In contrast, $-\nabla \cdot (\mathbf{u} p)$ and $-\rho g w$ indicate the two distinct physical processes altering KE: $-\nabla \cdot (\mathbf{u} p)$ denotes a divergence of 3D KE flux due to pressure work, while $-\rho g w$ quantifies the conversion between PE and KE, since it appears on the rhs of both the KE and PE equations but with opposite signs (Aiki et al. 2016; Chen et al. 2014). A similar consideration has been taken in previous studies diagnosing the Lorenz energy cycle in the ocean (e.g., Capet et al. 2008; von Storch et al. 2012; Chen et al. 2014; Kang and Curchitser 2015).

Applying the filtering operation [Eq. (1)] to Eqs. (5) and (6), the governing equations for motions with scales larger than L are obtained, i.e.,

$$\partial_t \bar{\mathbf{u}}_H + \bar{\mathbf{u}} \cdot \nabla \bar{\mathbf{u}}_H + f \mathbf{k} \times \bar{\mathbf{u}}_H = -\frac{1}{\rho_0} \nabla_H \bar{p} + \bar{\mathbf{D}}_H - \nabla \cdot \bar{\tau}(\mathbf{u}_H, \mathbf{u}) \quad (9)$$

and

$$0 = -\frac{1}{\rho_0} \partial_z \bar{p} - \frac{\bar{\rho} g}{\rho_0}, \quad (10)$$

where $\bar{\tau}(a, b) = \overline{ab} - \bar{a}\bar{b}$ is the stress tensor which indicates the forcing of small-scale motions on large-scale motions (Aluie et al. 2018). Multiplying Eq. (9) by $\rho_0 \bar{\mathbf{u}}_H$ and Eq. (10) by $\rho_0 \bar{w}$ and summing them together, we get the coarse KE budget equation:

$$\partial_t \text{KE}_L = -\nabla \cdot [\bar{\mathbf{u}} \text{KE}_L + \rho_0 \bar{\mathbf{u}}_H \cdot \bar{\tau}(\bar{\mathbf{u}}_H, \bar{\mathbf{u}})] - \nabla \cdot \bar{\mathbf{u}} \bar{p} - \bar{\rho} g \bar{w} - \Pi + \rho_0 \bar{\mathbf{u}}_H \cdot \bar{\mathbf{D}}_H. \quad (11)$$

Compared with the total KE budget equation [Eq. (8)], an additional term Π arises on the right-hand side, which denotes the cross-scale KE flux:

$$\Pi = -\rho_0 [\bar{\tau}(\mathbf{u}, \mathbf{u}) \cdot \nabla \bar{\mathbf{u}} + \bar{\tau}(\mathbf{u}, v) \cdot \nabla \bar{v}]. \quad (12)$$

The positive Π indicates energy transfer from scales larger than L to those smaller than L (forward energy cascade) and vice versa (inverse energy cascade). Moreover, the term $\rho_0 \bar{\mathbf{u}}_H \cdot \bar{\tau}(\bar{\mathbf{u}}_H, \bar{\mathbf{u}})$ represents the energy transport by motions at scales smaller than L .

Additionally, one can also get the filtered KE equation by applying the filter to Eq. (8), that is,

$$\partial_t \overline{\text{KE}} = -\nabla \cdot \overline{\mathbf{u} \text{KE}} - \nabla \cdot \overline{\mathbf{u} p} - \overline{\rho g w} + \overline{\rho_0 \mathbf{u}_H \cdot \mathbf{D}_H}, \quad (13)$$

where $\overline{\text{KE}} = (1/2)\rho_0(\overline{u^2} + \overline{v^2})$ is the filtered KE that is different from coarse KE [Eq. (3)]. Subtracting Eq. (11) from Eq. (13) yields the fine-scale KE budget equation:

$$\partial_t \text{KE}_f = \text{ADV} + \text{PW} + \text{BP} + \Pi + D. \quad (14)$$

The terms on the right-hand side are

$$\text{ADV} = -\nabla \cdot [\bar{\mathbf{u}} \overline{\text{KE}} - \bar{\mathbf{u}} \text{KE}_L - \rho_0 \bar{\mathbf{u}}_H \cdot \bar{\tau}(\bar{\mathbf{u}}_H, \bar{\mathbf{u}})], \quad (15)$$

$$\text{PW} = -\nabla \cdot (\overline{\mathbf{u} p} - \bar{\mathbf{u}} p), \quad (16)$$

$$\text{BP} = -g(\overline{\rho w} - \bar{\rho} \bar{w}), \quad (17)$$

$$D = \rho_0(\overline{\mathbf{u}_H \cdot \mathbf{D}_H} - \bar{\mathbf{u}}_H \cdot \bar{\mathbf{D}}_H). \quad (18)$$

Because of the minor contribution of vertical momentum flux (Srinivasan et al. 2023), the cross-scale KE flux can be simplified as

$$\Pi \approx -\rho_0 [\bar{\tau}(u, u) \bar{u}_x + \bar{\tau}(u, v)(\bar{u}_y + \bar{v}_x) + \bar{\tau}(v, v) \bar{v}_y]. \quad (19)$$

Here, advection (ADV) and PW terms represent the transport of fine-scale KE, whereas buoyancy production (BP), Π , and D act as the sources and sinks for KE. Note that only Π denotes the KE flux across the scale L , while other budget terms are within scales smaller than L . In this study, the coarse-graining approach is performed following Schubert et al. (2020) and Wang et al. (2026). Specifically, we first transform the geographical coordinates into Cartesian coordinates and then regrid model outputs onto a regular grid with a 3-km spacing. This treatment enables the use of a constant convolution kernel throughout the domain, thereby improving computation efficiency. Land points are assigned with zero velocity as suggested by Buzzicotti et al. (2023). Subsequently, the filtering operation is conducted as denoted by overbars in Eqs. (15)–(18). To obtain a scale-dependent KE budget, these terms are calculated using the outputs of OFES2_NP30 by choosing a series of L (12, 18, 24, 30, 42, 54, 72, 102, 150, 210, 300, and 450 km). We do not calculate the fine-scale KE budget for L larger than 450 km because this study focuses on submesoscale and mesoscale processes. Also, the computation is expensive for convolution filtering with a large spatial window.

Although the scale-dependent KE budget is, in principle, closed within the coarse-graining framework, we are unable to completely close the KE budget in practice due to the following

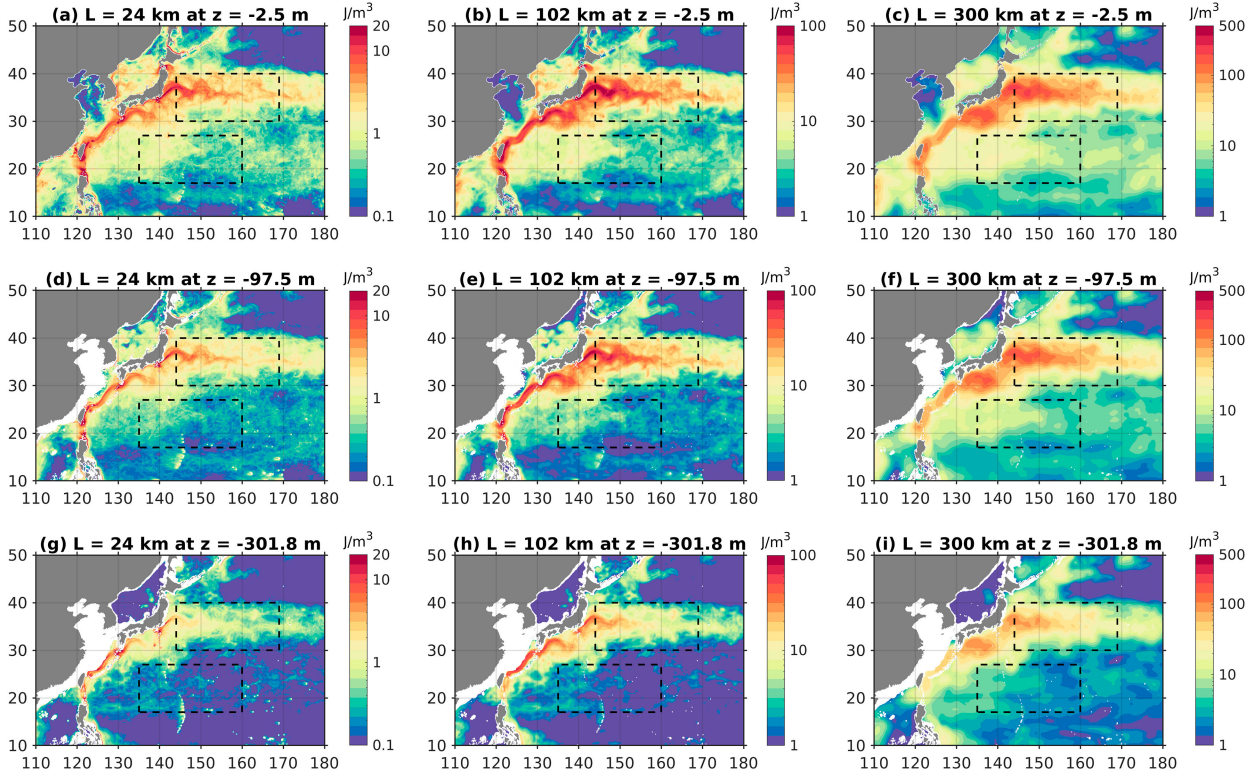


FIG. 2. Fine-scale KE (shadings; J m^{-3}) at scales of 24, 102, and 300 km at different depths. The two dashed rectangles in each panel indicate the KET and STCC regions. Note that the range of color bar is different for each column.

reasons. First, viscosity terms are not output from the OFES2_NP30 because of the storage limitation, preventing the direct computation of the KE dissipation term [Eq. (18)]. Second, the computation is conducted using the variables interpolated onto the same grid rather than on the raw B grid of the OGCM. While the approach is computationally convenient, it may introduce additional numerical errors. The computed residual of Eq. (14) is shown in Fig. S3, which includes not only KE dissipation but also forcing and energy transport due to viscosity terms and surface fluxes.

3. Fine-scale KE and KE spectra

Figure 2 shows the annually averaged fine-scale KE at scales of 24, 102, and 300 km at varying depths. At a scale of 24 km (submesoscale), fine-scale KE is high along the Kuroshio and KET at the surface (Fig. 2a). Although STCC is another region with active submesoscale motions (e.g., Qiu et al. 2014; Sasaki et al. 2014; Jing et al. 2021), KE_f there is approximately one order of magnitude smaller than that in the KET. As depth increases, KE_f decreases to a magnitude of 0.1 J m^{-3} or even lower in most areas (Fig. 2g). However, in the Kuroshio and KET regions, it remains nonnegligible at a depth of 301.8 m, which is well below the mixed layer.

Results for a scale of 102 km (mesoscale) resemble those for $L = 24$ km despite the higher magnitude of fine-scale KE (Figs. 2b,e,h). Note that submesoscale KE is also included in

the results for $L = 102$ km, as the fine-scale KE here accounts for the KE with scales smaller than the given L . However, when the scale further increases to 300 km, the fine-scale KE decays more slowly in the vertical direction, with fine-scale KE at $z = -301.8$ m comparable to that at $z = -97.5$ m (Figs. 2f,i). An exception is found in the Japan Sea, where the fine-scale KE at $z = -301.8$ m is one order of magnitude lower than at $z = -97.5$ m. These results suggest that the vertical structure of fine-scale KE is not only scale dependent but also regionally dependent.

We further examine the scale dependence of KE by examining the KE spectrum. With the coarse KE, the KE spectrum is calculated as

$$\text{KE}(L^{-1}) = \frac{d\{\text{KE}_L\}}{dL^{-1}} = -L^2 \frac{d\{\text{KE}_L\}}{dL}, \quad (20)$$

where $\{\cdot\}$ denotes a horizontal average in a given domain (Storer et al. 2022; Buzzicotti et al. 2023). Here, $d\{\text{KE}_L\}$ at L is approximated by the difference of KE_L at two adjacent scales, while the mean of the two scales is used for computing L^2 . This yields the KE spectrum within a scale range from 15 to 375 km. It has been illustrated that the calculated KE spectrum by Eq. (20) is consistent with that computed with the Fourier analysis (Sadek and Aluie 2018; Storer et al. 2022).

The KE spectra for the KET and STCC regions are shown in Fig. 3. Although the KE is higher in the KET region than in the STCC region, the two spectra share some similarities:

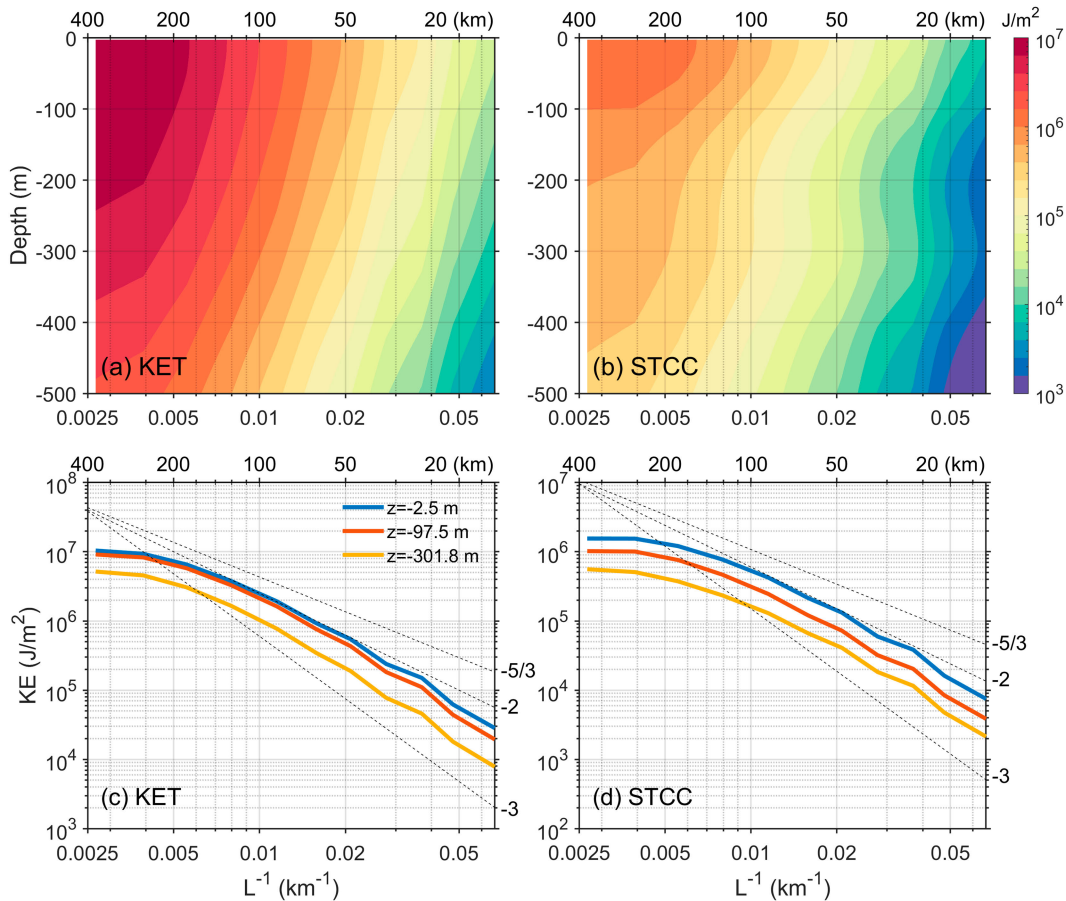


FIG. 3. KE spectra (shadings; J m^{-2}) estimated with the coarse-graining approach for the (a) KET and (b) STCC regions. (c),(d) The spectra at three selected depths are shown, with dashed lines indicating slopes of $-5/3$, -2 , and -3 .

First, the spectral density decreases with horizontal scale at all depths in both regions (Figs. 3a,b), consistent with the results derived from the shipboard ADCP measurements (Qiu et al. 2017); second, the KE spectra have slopes between -3 and -2 , which becomes slightly steeper with the increase of depth (Figs. 3c,d). These features are also found in the KE spectra for other subtropical regions, such as the North Atlantic (Callies et al. 2015; Ajayi et al. 2021). According to Sasaki and Klein (2012) and Wang et al. (2026), the peaks in the KE spectra for the two regions appear at a scale larger than 400 km, which lie beyond the scale range considered in this study.

Despite these similarities, differences are also found in the spectra for the two regions. Compared with that in the STCC region, the KE in the KET region is more deep reaching (Figs. 3a,b). Namely, the KE at the same scale decays more slowly in the vertical direction in the KET region than in the STCC region, which is attributed to the shallower background current in the latter (Figs. 1b,d). Additionally, in the STCC region, the KE in the submesoscale band ($15 < L < 50$ km) decays nonmonotonically with the increase of depth (Fig. 3b). These features are consistent with observational results reported by Qiu et al. (2017, 2022).

4. Scale-dependent fine-scale KE budget

a. Horizontal distribution

Annual-mean fine-scale KE budgets for $L = 24$ and 102 km are presented in Figs. 4 and 5, respectively, which are vertically integrated within 0–50 m. For submesoscale ($L = 24$ km; Fig. 4), it is evident that the fine-scale KE budget is a predominant balance between PW and BP, both of which are enhanced along the Kuroshio and KET (Figs. 4b,d). Here, BP is the primary source of fine-scale KE, indicating that baroclinic instability is responsible for the generation of submesoscale flows (e.g., Qiu et al. 2014; Sasaki et al. 2014, 2017; Jing et al. 2021). The generated fine-scale KE is mainly balanced by PW, whose role will be analyzed later in detail in section 5. Additionally, in the KET region, BP also exhibits a meridional asymmetry, with higher values on the northern side of KET (35° – 40° N, Fig. 4d), which has not been clearly revealed by previous studies using the spectral approach (Sasaki et al. 2014, 2017). McWilliams et al. (2019) found that the baroclinic instability is stronger at the northern side of the Gulf Stream north wall, due to the asymmetry of frontogenesis on two sides of the meandering current.

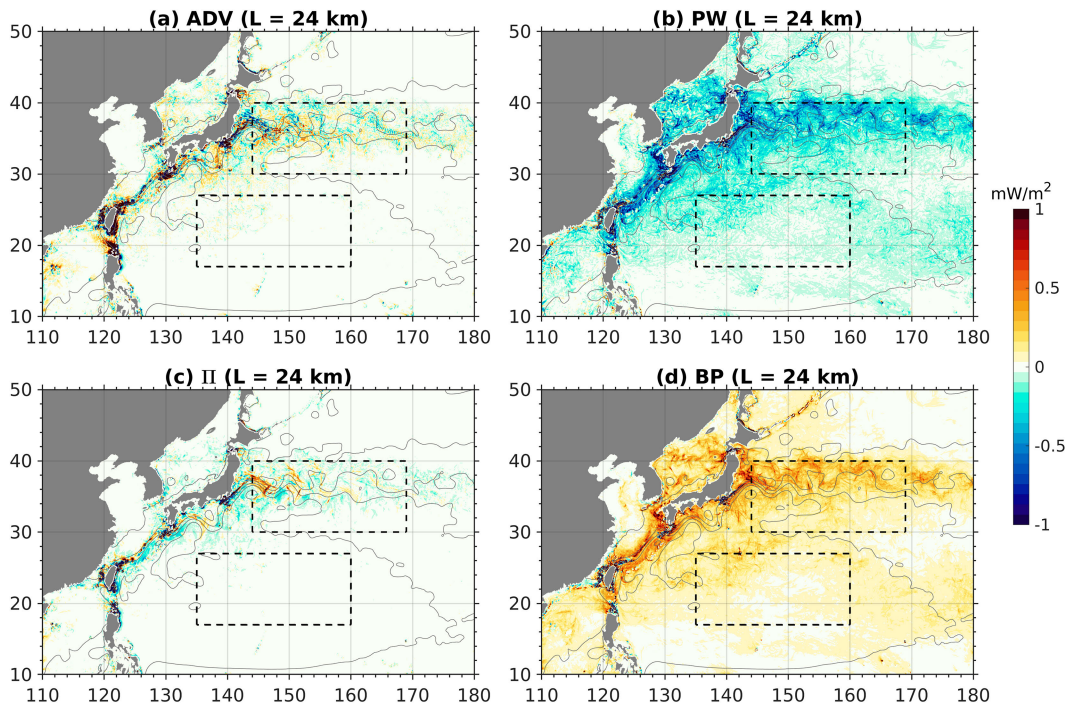


FIG. 4. Vertically integrated (upper 50 m) fine-scale KE budget for $L = 24$ km: (a) ADV, (b) PW, (c) cross-scale KE flux (Π), and (d) BP. In each panel, the contours indicate annual-mean SSH with an interval of 0.25 m, and two dashed rectangles in each panel denote the KET and STCC regions.

Compared with BP, the cross-scale KE flux is only significant within a narrow band along the Kuroshio and KET (Fig. 4c). Additionally, the forward energy cascade is enhanced on the left side of the Kuroshio axis, which has been reported in Wang et al. (2026). A similar phenomenon is also found in other western boundary current regions (Contreras et al. 2023). The remarkable density and current gradient along the north wall of the western boundary current in the Northern Hemisphere favors intense frontogenesis and thereby the forward energy cascade (Klymak et al. 2016; McWilliams et al. 2019; Wang et al. 2026). ADV is comparable to Π in the open ocean, with alternating positive and negative values along the KET but much lower than BP and PW. The above results indicate that only a small fraction of fine-scale KE at submesoscales corresponds to the cross-scale energy cascade and nonlocal transport by the background current in the KET region.

The fine-scale KE budget for scales below a selected mesoscale (102 km) is presented in Fig. 5. It should also be noted that these budget terms account for the contributions from all resolved motions with scales smaller than L . Whereas, given the relative low level of submesoscale KE_f and its associated budget terms (Figs. 2 and 4), the budget terms in Fig. 5 are largely attributed to mesoscale processes. Differing from that for submesoscale, the four terms have almost comparable magnitudes in the upper ocean, particularly in the KET region. At this scale, KE_f is also generated by BP (Fig. 5d), while the cross-scale KE fluxes are largely negative along the Kuroshio, indicating a dominant inverse energy cascade (Fig. 5c). Specifically, the KE is transferred to motions at a scale larger than 102 km, agreeing

with (Yang and Liang 2018), who suggested an opposite role of barotropic and baroclinic instability in the EKE budget in the KET region using a mesoscale-resolving OGCM. As for the PW, widespread negative values are found almost in the whole domain, with a maximum in the KET region (Fig. 5b). In contrast, the ADV exhibits a distinct feature from that for $L = 24$ km, as the positive values are dominant in the KET region, indicating a KE input due to the nonlocal transport by the Kuroshio (Fig. 5a).

b. Vertical structure

The vertical structure of the fine-scale KE budget terms is examined based on the zonally averaged results within 150°–160°E (Fig. 6). BP acts as the primary source of KE_f at both mesoscale and submesoscale (Figs. 6g,h). It not only maximizes in the mixed layer but can also penetrate a greater depth in the KET region where a deep density front exists. Such a structure of BP has been found for mesoscale eddies in previous studies (e.g., Zhai and Marshall 2013; Yang and Liang 2018; Matsuta and Masumoto 2023). More recent observations (Yu et al. 2019; Zhang et al. 2021) and numerical simulations (Sasaki et al. 2017; Ajayi et al. 2020) confirmed a similar phenomenon for submesoscales as well, explaining why submesoscale KE spreads over a larger vertical extent compared with the MLD (Qiu et al. 2014; Capet et al. 2016; Zhang et al. 2021).

The deep-reaching vertical structure of BP differs from the prediction by the traditional mixed layer instability described by the Eady model, in which the perturbations are restricted

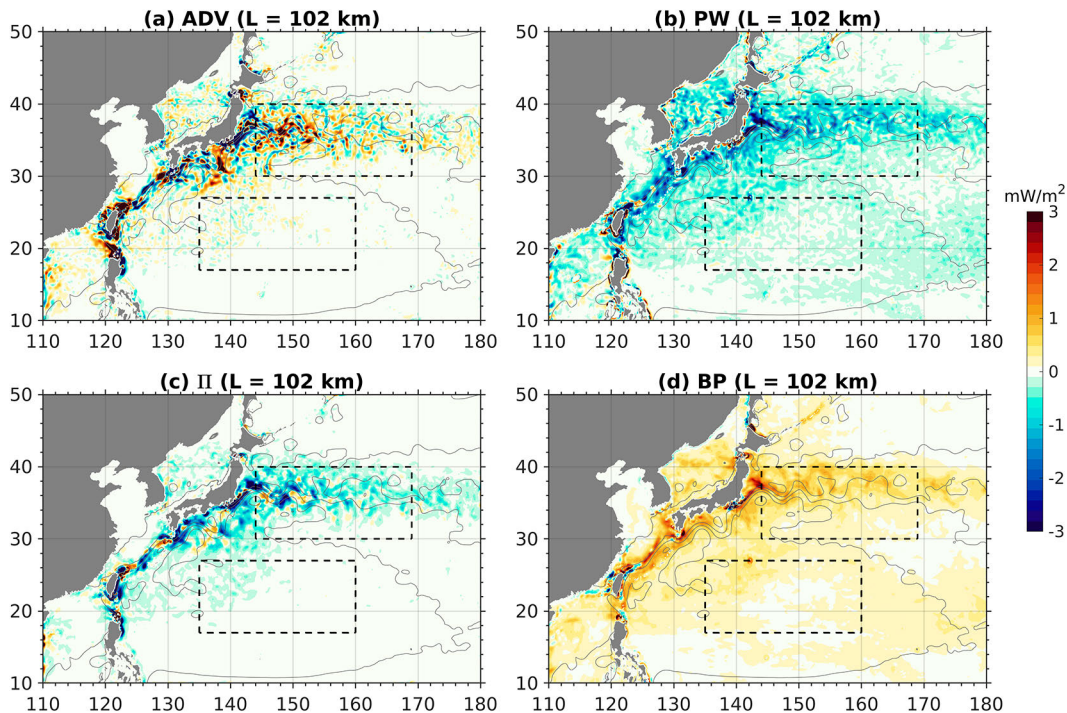


FIG. 5. As in Fig. 4, but for $L = 102$ km.

within the mixed layer and the largest growth rates appear near its upper and lower boundaries (Boccaletti et al. 2007; Fox-Kemper et al. 2008; Feng et al. 2025a). Instead, it is attributed to the Charney-type baroclinic instability characterized by a vertically decaying structure from the sea surface (Capet et al. 2016; Feng et al. 2025a; Zhang et al. 2025). The vertical scale of instability, i.e., the Charney depth, can be larger than the MLD in the western boundary current regions (Feng et al. 2025a,b).

Similar to the horizontal distribution, the energy source term, BP, is largely balanced by PW within the mixed layer. PW decreases KE_f in the mixed layer and increases it in the deep region for the two selected scales (Figs. 6c,d). Beneath the mixed layer, PW and BP play a comparable role in intensifying eddies. A similar vertical structure of PW has also been found in the equatorial Pacific (Wang et al. 2018) and northern South China Sea (Yang et al. 2013). An exception is found for $L = 102$ km in the KET region, where comparable positive and negative values of PW exist alternately beneath the mixed layer (Fig. 6d).

On the other hand, the energy cascade, Π , and advective KE transport, ADV, are significant only in the KET region (30° – 40° N; Figs. 6a,b,e,f). For $L = 24$ km, the forward energy cascade appears near the surface within a depth range of approximately 0–50 m, while below that, the inverse energy cascade dominates. The result also confirms the enhancement of forward energy cascade around the north wall of the KET (35° – 40° N) (Fig. 4c), where a strong meridional gradient of density exists near the sea surface (Fig. 6e). For $L = 102$ km, Π is negative and comparable to BP in the KET region (Figs. 6f,h), indicating that a considerable amount of KE generated via

baroclinic instability is transferred upscale to larger horizontal scales. Nevertheless, outside of the KET region, the inverse energy cascade is weak, consistent with the previous estimates based on the spectral approach (Sasaki and Klein 2012; Sasaki et al. 2017) and structure function method (Qiu et al. 2022).

c. Scale dependence

The scale dependence of fine-scale KE budget terms is further examined by horizontally averaging them for different scales in the KET and STCC regions (Fig. 7). In the KET region, the ADV is mostly positive and enhances as the scale increases (Fig. 7a). For scales larger than 100 km, its magnitude decays from the sea surface to deeper layers, indicating a significant KE source due to advection near the surface related to the KET. As for PW, its vertical structure differs for $L < 60$ km and $L > 60$ km (Fig. 7b). A vertical dipole pattern occurs for $L < 60$ km, with the zero crossing depth increasing with scale. The result again demonstrates an important role of PW in accumulating energy for submesoscale motions beneath the mixed layer. For $L > 60$ km, PW is negative within 0–500 m, indicating a transport of fine-scale KE out of the source region (Yang et al. 2021a; Pang et al. 2022; Hao et al. 2022; Wang et al. 2025). A more detailed analysis of the PW term is provided in section 5.

In contrast, BP is positive for all scales considered in this study (Fig. 7d) and appears at a shallower depth for smaller scales, a typical feature of Charney-type instability (Roullet et al. 2012). Additionally, BP varies nonmonotonically with the increase of scale, reaching its maximum value within 200–250 km depending on the depth. By calculating the cospectrum of wb , previous studies found that wb is positive (negative) at high (low) wavenumbers (Roullet et al. 2012; Wang et al. 2019;

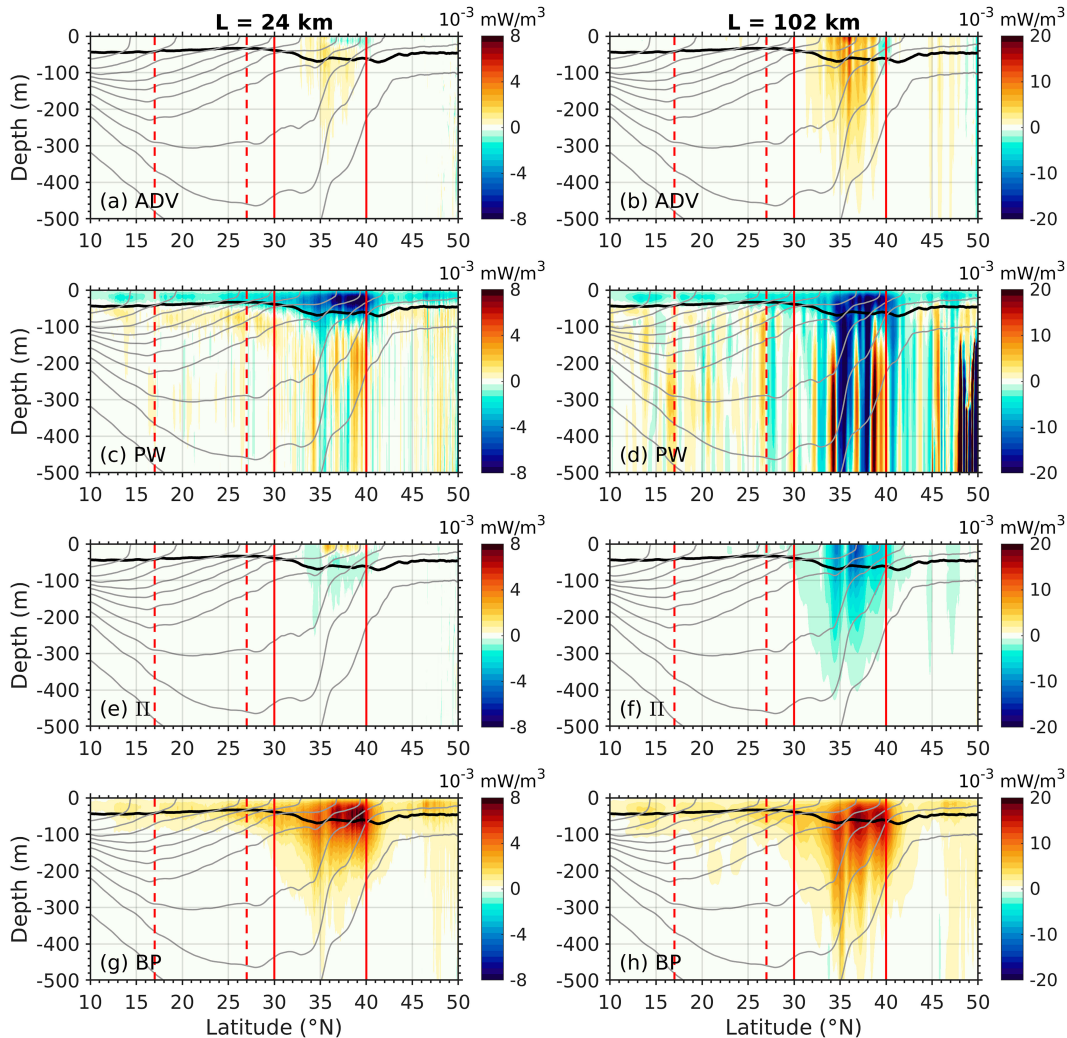


FIG. 6. Zonally averaged (150° – 160° E) fine-scale KE budget (shadings; mW m^{-3}) for (a),(c),(e),(g) $L = 24$ km and (b),(d),(f),(h) 102 km. In each panel, the black curve denotes zonally averaged MLD, and gray contours indicate potential density with an interval of 0.5 kg m^{-3} . The KET and STCC regions are bounded by red solid and dashed lines, respectively.

Dong et al. 2020), implying that the integral of spectral wb , equivalent to the BP here, will have a maximum at the wave-number where wb changes sign.

The forward energy cascade occurs at scales smaller than 30 km near the surface (Fig. 7c). With the increase of depth, the cross-over scale, at which the sign of Π changes, decreases quickly, indicating that the forward energy cascade mainly energizes submesoscale motions within a very shallow layer near the surface (Cao et al. 2021). This feature is common in other western boundary current regions, such as the Gulf Stream (Ajayi et al. 2021) and the Agulhas Current regions (Schubert et al. 2020). It should be noted that, however, the forward energy cascade at submesoscales is likely underestimated in the present analysis due to the limited horizontal resolution, as illustrated via the comparison of results from models with horizontal resolutions of 3 and 0.5 km (Capet et al. 2008; Srinivasan et al. 2023). At larger scales, the inverse energy cascade dominates,

which exhibits a similar feature as BP, indicating that a portion of the KE generated by baroclinic instability is cascaded to larger scales. This result is confirmed by Qiu et al. (2022), in which the cross-scale KE flux was estimated from ADCP measurements using the third-order structure function.

In the STCC region, the scale-dependent vertical structure differs from that in the KET region in the following aspects: First, due to the weak background current, ADV makes only a minor contribution to the fine-scale KE budget, which does not vary remarkably across different scales (Fig. 7e). Second, the vertical dipole pattern of PW is found for all scales, with the magnitude higher at larger scales (Fig. 7f). Third, both the inverse energy cascade and BP penetrate a shallower depth compared with those in the KET region (Figs. 7g,h), which could be attributed to the shallower mixed layer and stronger stratification in the STCC region (Fig. 1d). The results agree with the theoretical predictions by Feng et al. (2025a), who

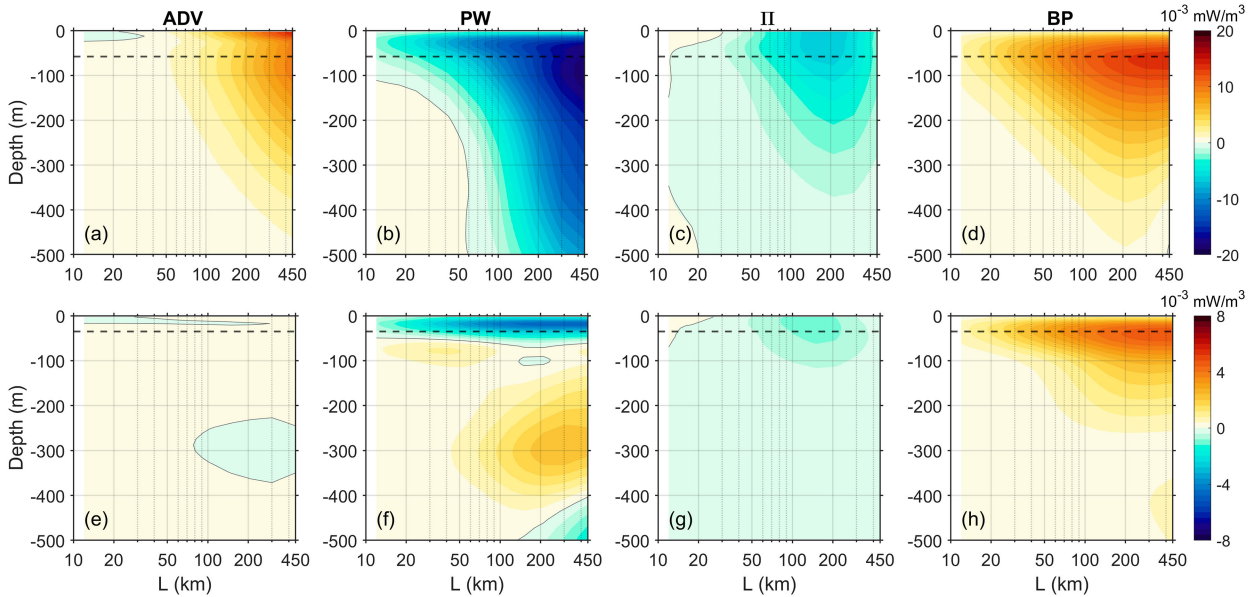


FIG. 7. Horizontally averaged fine-scale KE budget (shadings; mW m^{-3}) in (a)–(d) KET and (e)–(h) STCC regions. In each panel, the dashed line indicates the horizontally averaged MLD.

demonstrated a larger Charney depth in the KET region than in the STCC region. In other words, a shallower MLD results in weaker energy conversion via baroclinic instability.

5. Energy transport by pressure work

The results in section 4 indicate an important role of PW in the fine-scale KE budget, particularly for submesoscales, which were not well acknowledged in previous studies. Hence, a detailed examination on PW is conducted for clarification. We decompose PW into horizontal and vertical components:

$$PW = \underbrace{-\nabla_H \cdot (\overline{\mathbf{u}_H \overline{p}} - \overline{\mathbf{u}_H} \overline{p})}_{PW_H} - \underbrace{\partial_z (\overline{w \overline{p}} - \overline{w} \overline{p})}_{PW_V}. \quad (21)$$

Figures 8a and 8c show the decomposition results for $L = 24$ km. The vertical component generally dominates the total PW (Fig. 8c), while the horizontal component is weak, especially within the mixed layer (Fig. 8a). This result demonstrates an important role of PW in the vertical redistribution of submesoscale KE. To clarify the direction of vertical energy transport, we show the vertical eddy pressure flux for the same scale:

$$F_V = \overline{w \overline{p}} - \overline{w} \overline{p}. \quad (22)$$

As presented in Fig. 8e, F_V is mostly negative, indicating a downward transport of fine-scale KE by pressure work. These results indicate that the fine-scale KE generated by BP in the mixed layer is transported downward by the PW, resulting in a growth of fine-scale KE beneath the mixed layer. The accumulation of subsurface KE associated with this redistribution is comparable to that linked to BP (Fig. 7), highlighting that

vertical PW plays a key role in the deep-reaching submesoscale processes in addition to the deep Charney-type baroclinic instability (Capet et al. 2016; Feng et al. 2025a; Zhang et al. 2025).

As the scale increases to 102 km, the horizontal component of PW becomes greater than the vertical component, particularly beneath the mixed layer (Figs. 8b,d). An alternating pattern of positive and negative PW_H is found, with a larger magnitude in the KET region (Fig. 8b). It indicates that the horizontal energy transport by PW is important for mesoscale eddies, which has also been identified in other high EKE regions (Yang et al. 2021a; Pang et al. 2022; Hao et al. 2022). Such a spatial structure of PW cannot be obtained with the spectral approach, which yields only a spatially averaged spectrum of PW for the given domain (Capet et al. 2008; Wang et al. 2019).

The vertical component PW_V for $L = 102$ km resembles that for $L = 24$ km, but the negative values appear at depths deeper than MLD in the KET region (Fig. 8d). This result corresponds to the deep-reaching vertical eddy pressure flux due to mesoscale eddies, which is enhanced with the increase of depth (Fig. 8f). At mesoscales, energy conversion by baroclinic instability occurs at greater depth than at submesoscales (Figs. 6h and 7d), resulting in strong and deep downward eddy pressure flux. The flux can even reach the sea bottom in eddy-rich regions, such as the KET (Yang et al. 2021a), the Gulf Stream (Zhai and Marshall 2013), the Antarctic Circumpolar Current (Matsuta and Masumoto 2023), and the Red Sea (Zhan et al. 2016).

The scale dependence of PW_H and PW_V is further examined in the KET and STCC regions (Fig. 9). In the KET region, PW_H is negative for all scales considered in this study and intensifies with increasing scale, indicating that fine-scale KE is transported out of this region, largely by mesoscale

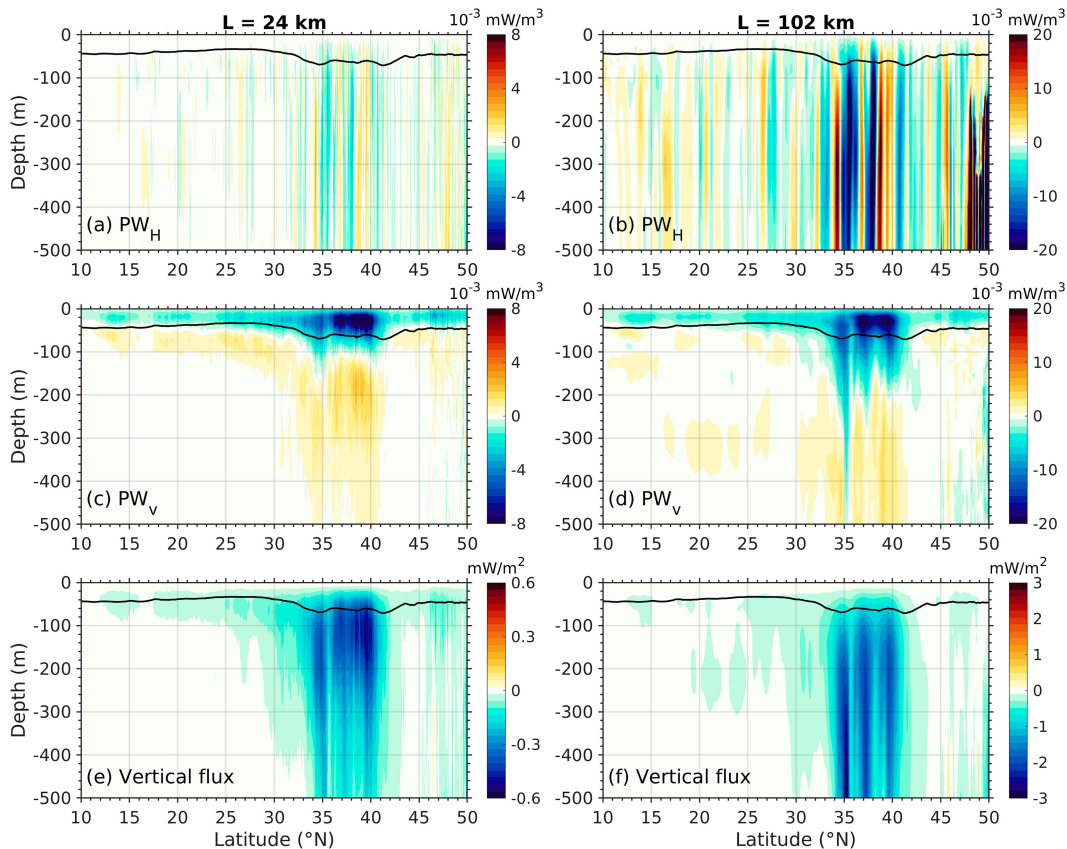


FIG. 8. Zonally averaged (150° – 160° E) (a),(b) horizontal and (c),(d) vertical components of PW (shadings; mW m^{-3}) as well as (e),(f) vertical eddy energy flux (shading; mW m^{-2}) at scales of 24 and 102 km. The black curve in each panel denotes zonally averaged MLD.

eddies (Fig. 9a). However, this term would vanish when integrating over the whole basin or globally (von Storch et al. 2012). For $L > 100$ km, PW_H maximizes at a depth of approximately 300 m. In contrast, PW_V exhibits a vertical dipole pattern that persists across scales (Fig. 9b). As L increases, the zero crossing depth of PW_V first increases and then decreases, reaching its maximum at $L = 200$ km, consistent with the structure of BP (Fig. 7d). Additionally, PW_V has a larger magnitude than PW_H for $L < 50$ km, indicating a more significant role of vertical KE transport by PW at submesoscales. For $L > 50$ km, PW_V is larger (smaller) than PW_H within (beneath) the mixed layer, and both are negative (Figs. 7a,b). These results reveal that the relative contribution between PW_H and PW_V is scale dependent, which has been overlooked by previous studies based on the Reynolds decomposition. This is because, in the Reynolds decomposition, eddies are defined as perturbations deviating from the mean value, and scale-dependent processes cannot be explicitly separated (Yang et al. 2013; Zhan et al. 2016; Hao et al. 2022).

The pressure work is much weaker in the STCC region than that in the KET region and exhibits a different vertical structure. The horizontal component, PW_H , is weaker compared with PW_V . The PW_H is mostly positive within 0–400 m for $L < 100$ km (Fig. 9c), and as the scale increases, the zero

crossing depth of PW_H decreases, which is approximately 200 m for $L = 450$ km. As for PW_V , it also exhibits a vertical dipole pattern analogous to that in the KET region (Fig. 9d). However, its negative values are largely restricted within the mixed layer, except for small values at a depth range of 50–150 m for $L > 100$ km. Below 150 m, PW_V is positive, with maximum values at a depth of 300 m for $L > 50$ km, suggesting that mesoscale eddies are intensified by the vertical energy flux as well in the subsurface STCC region. Qiu et al. (2014) demonstrated that in the STCC, mesoscale eddies with a dominant scale of $O(250)$ km are generated via the interior baroclinic instability (BP term), whereas our results illustrate that PW plays a comparable or even more important role than BP in enhancing mesoscale eddies in the subsurface layer.

The vertical dipole pattern of PW_V can be interpreted within the framework of the quasigeostrophic (QG) baroclinic instability theory. To illustrate this mechanism, we perform a linear instability analysis using the QG potential vorticity (PV) equation (see the appendix for details). Horizontally averaged geostrophic velocities and stratification in the KET and STCC regions are used as inputs. As shown in Fig. 10a, two pronounced peaks of growth rates are detected at 120 and 152 km for the KET and STCC regions, respectively. According to Srinivasan et al. (2023), the filtering scale L in the coarse-graining

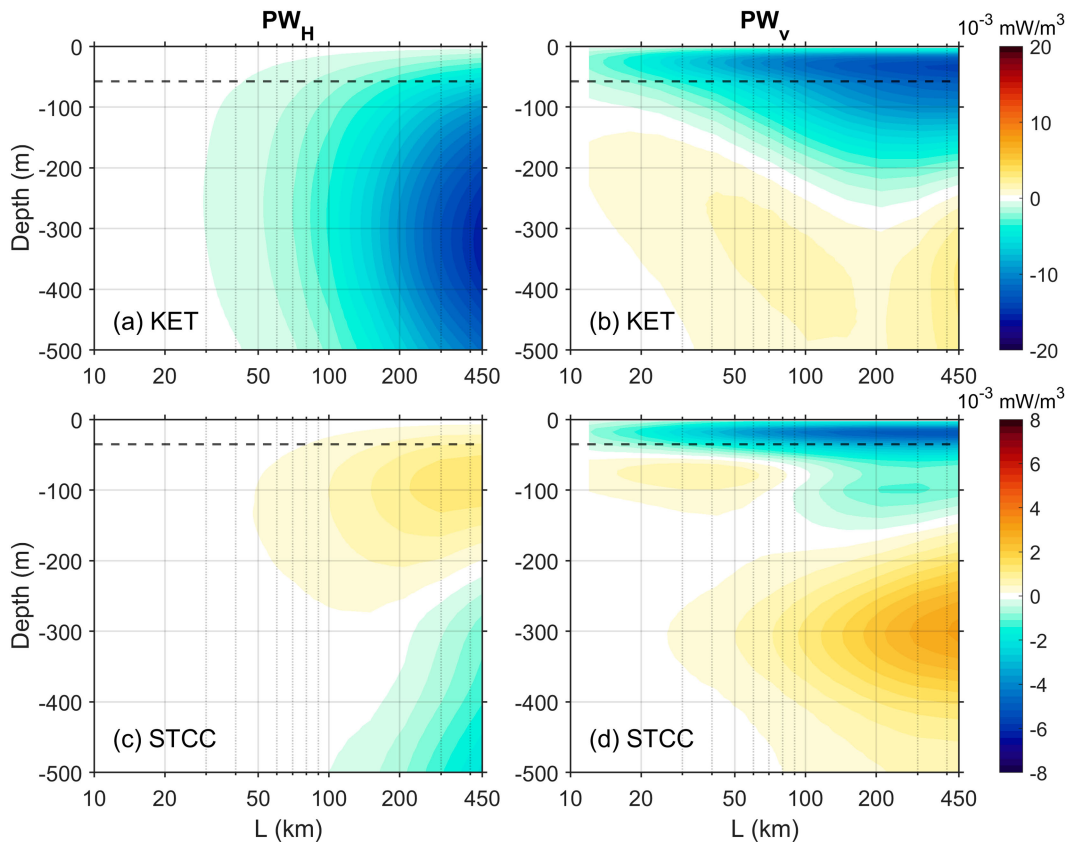


FIG. 9. Horizontally averaged horizontal and vertical components of PW in the (a),(b) KET and (c),(d) STCC regions. In each panel, the dashed line indicates the horizontally averaged MLD.

framework is related to the Fourier wavelength λ by $\lambda \approx 2.4L$, since the top-hat filter used in this study is not spectrally sharp. Therefore, the two selected modes correspond to $L \approx 50$ km for the KET region and $L \approx 63$ km for the STCC region, respectively. The unstable modes at wavelengths shorter than 20 km are not considered despite their high growth rate, as their associated L are smaller than the effective resolution of OFES2_NP30. The obtained streamfunctions exhibit a vertically decaying structure from the sea surface (Fig. 10b), a typical feature of the Charney-type baroclinic instability (Smith 2007; Tulloch et al. 2011; Feng et al. 2025a). Despite the vertical decay, the amplitude of the unstable modes remains remarkable beneath the mixed layer compared to that near the surface, indicating the penetration of unstable waves into the ocean interior.

We further calculate vertical pressure work and buoyancy production using the streamfunction of unstable waves (see the appendix for the equations). In the two regions, the vertical pressure work transports kinetic energy both upward and downward from intermediate depths (200–400 m in the KET region and 80–250 m in the STCC region), thereby offsetting the local energy input via buoyancy production (Figs. 10c,d). The downward energy transport is consistent with the model results shown in Figs. 7 and 9. In contrast, the upward energy

transport toward the surface is not evident in the model diagnostics, possibly due to nonlinear processes and surface forcing effects that are absent in the linear instability analysis. The results further indicate that vertical energy redistribution by PW_V could be related to linear wave dynamics. Physically, the vertical eddy pressure flux is related to the vertical propagation of Rossby waves generated from instability region. For Charney-type baroclinic instability with energy source near the surface, the resulting Rossby waves tend to propagate away from the source region, thereby leading to a net downward energy transport (Pedlosky 1987; Vallis 2017).

Beyond its role as a compensating term in the local KE budget, PW_V can be interpreted as the redistribution of energy associated with the wave–eddy adjustment in the QG turbulence. During oceanic baroclinic instability, buoyancy conversion injects eddy kinetic energy preferentially near the surface or within regions of strong vertical shear. This localized energy source excites Rossby wave–like disturbances that radiate away from the instability region, transporting energy both horizontally and vertically, which is evident in the QG model simulations (Roulet et al. 2012). The resulting divergence of eddy pressure flux, therefore, reflects the propagation of wave activity away from the source region rather than local energy production or dissipation.

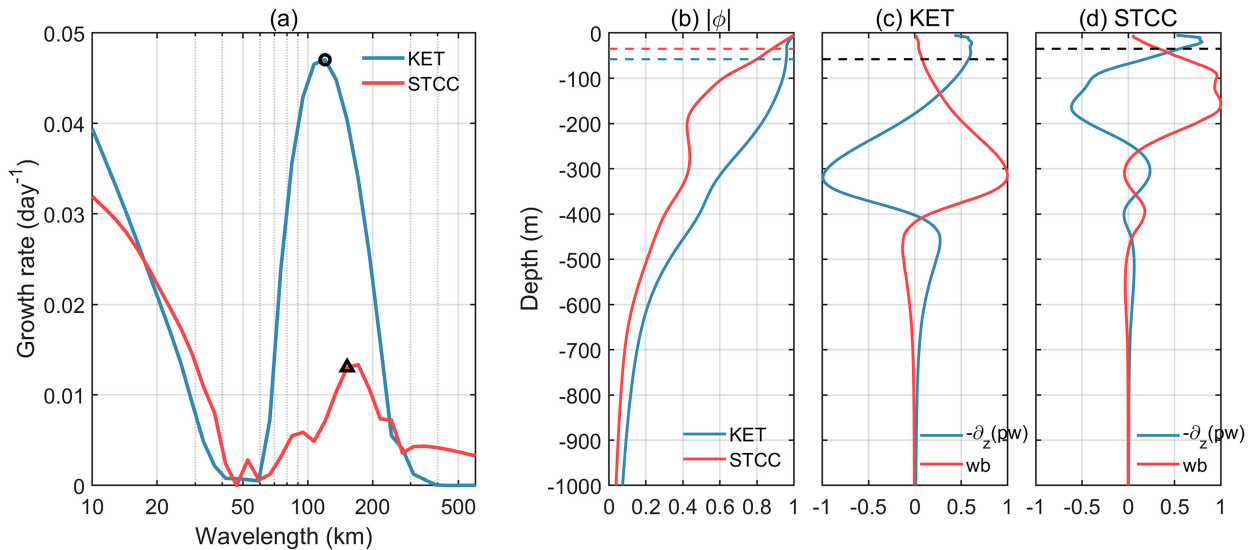


FIG. 10. (a) Growth rates for unstable waves as a function of wavelength for the KET and STCC regions. (b) Amplitudes of unstable modes indicated by black circle and triangle in (a) for the KET and STCC regions, respectively. Vertical PW and BP for (c) the KET and (d) STCC regions calculated from the streamfunction of unstable modes. In (b)–(d), dashed lines denote the horizontally averaged MLD in the two regions, and all the vertical profiles are normalized by their maximum magnitude.

6. Seasonal variation

The seasonal variation of the fine-scale KE budget is further examined. In this section, we focus only on the results for $L = 24$ km (submesoscale), which are presented in Fig. 11. Note that the tendency terms $\partial_t KE_f$ are shown in Fig. S4, as they are much smaller than other budget terms. In the KET region, fine-scale KE exhibits a vertically coherent peak in April, lagging behind the MLD peak by 2 months. Although high KE_f is notable within the mixed layer in winter and spring, the magnitude of KE_f below the mixed layer reaches nearly one-third of that near the sea surface, indicating the deep penetration of submesoscale processes. In contrast, the dominant energy source, BP, varies synchronously with MLD, both of which peak in February (Fig. 11g). A similar seasonal cycle is also found for the downscale KE flux (positive Π), despite its smaller magnitude and shallower depth range (Fig. 11c).

PW balances BP within the mixed layer, which peaks in the same month as BP (Fig. 11i), thereby considerably removing KE_f generated by BP from the mixed layer. The vertical structure of PW exhibits distinct seasonal variation. In winter, PW exhibits a vertical dipole pattern. Notably, the positive PW has an even greater magnitude than BP beneath the mixed layer, highlighting the significant role of PW in redistributing submesoscale KE_f into the ocean interior. From winter to autumn, the positive PW gradually decreases to zero, while the negative PW appears to propagate downward from the surface to deeper layers, reaching a maximum depth of approximately 200 m in summer. The residual (Res) term is positive within the mixed layer and negative below that (Fig. 11k). The positive residuals within the mixed layer are likely due to the energy input by external forcing such as wind (Yang et al. 2021b), whereas the negative ones may indicate the dissipation of KE due to viscosity effect.

The submesoscale KE_f and its budget terms in the STCC region exhibit seasonal cycles similar to those in the KET region (Fig. 11b). Due to the shallower mixed layer and weaker meridional density gradient in the STCC region (Fig. 1), BP cannot penetrate as deeply as in the KET region, even in winter (Fig. 11h). Correspondingly, PW is also confined in the upper 150 m (Fig. 11j). The small magnitude of Π again indicates a minor role of the energy cascade in affecting the seasonal cycle of submesoscale KE, which is also the case for ADV (Figs. 11d,f).

7. Summary and discussion

This study examines the spatial structure and scale dependence of the KE budget in the northwest Pacific using a submesoscale-permitting OGCM. The findings underscore the importance of the vertical energy transport by pressure work in the energy redistribution across both space and scales. Using a coarse-graining approach, we reveal that at submesoscales (~ 10 km), the vertical component of pressure work facilitates a downward transport of KE from the mixed layer to the ocean interior, which is as significant as baroclinic instability in intensifying submesoscale motions beneath the mixed layer (Fig. 12a). The energy cascade is weaker than the pressure work and largely upscale, with downscale energy transfer confined only along the KET. As a result, the seasonal cycle of fine-scale KE budget within the mixed layer is determined by a balance between BP and PW, both of which peak in winter but minimize in summer.

As the horizontal scale increases to mesoscale (~ 100 km), a considerable amount of KE generated via baroclinic instability is transferred upscale via inverse energy cascade (Fig. 12b). Concurrently, the horizontal component of pressure work becomes increasingly important, particularly in the Kuroshio Extension region, signifying an enhanced horizontal energy

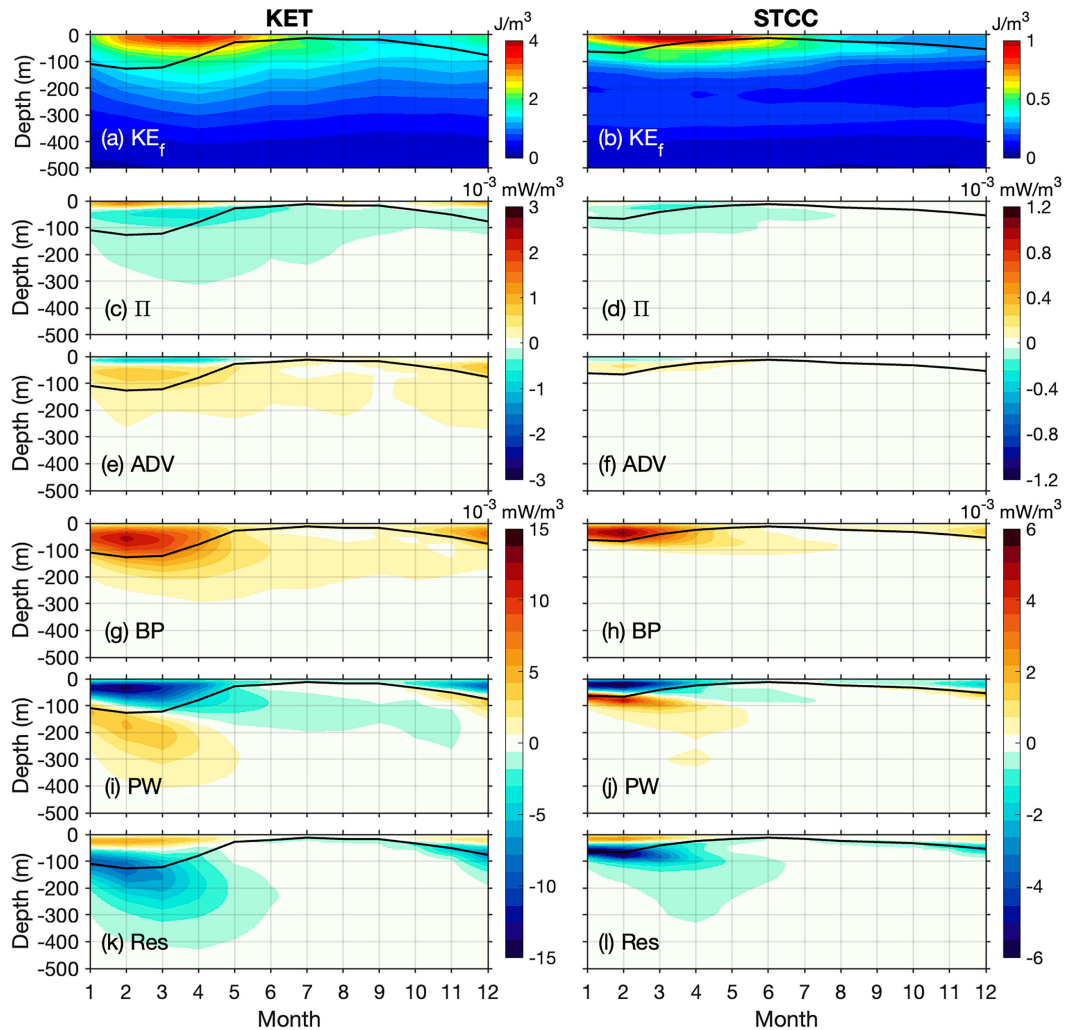


FIG. 11. Depth–time maps of horizontally averaged monthly mean (a),(b) fine-scale KE, (c),(d) II, (e),(f) ADV, (g),(h) BP, (i),(j) PW, and (k),(l) residual terms for $L = 24$ km in the KET and STCC regions. The black curve in each panel indicates the monthly time series of horizontally averaged MLD.

transport by mesoscale eddies. At these larger scales, the vertical pressure work also contributes to a downward energy transport from the upper ocean to the interior. The energy cascade for mesoscales is mainly upscale and comparable in magnitude to the pressure work, reflecting a barotropization tendency of geostrophic eddies (Smith and Vallis 2001).

Our findings reveal a tight coupling between horizontal and vertical pathways of energy redistribution. When submesoscale eddies cascade energy to larger horizontal scales, they simultaneously extend deeper into the water column through the vertical energy transport by pressure work. These horizontal and vertical energy transfers represent two facets of the same physical process: the quasi-isotropic inverse energy cascade in the three-dimensional geostrophic turbulence originally envisioned by Charney (1971). The quantitative relation between the horizontal and vertical scales is explicit in the surface QG model, which has proven effective in interpreting the vertical structure of ocean eddies (e.g., Lapeyre and Klein 2006;

Qiu et al. 2016; Zhang et al. 2024). Based on this relation, Zhang et al. (2024) revealed that the energy generated by baroclinic instability transfers simultaneously to both larger horizontal scales and deeper vertical modes, ultimately feeding energy into mesoscale eddies and sustaining the observed scale-dependent vertical structure of eddy energy.

The vertical redistribution of energy by pressure work also serves as a self-organization mechanism of submesoscale eddies: It channels KE into the ocean interior—where geostrophic balance dominates—thereby favoring an upscale cascade rather than dissipation through forward cascade or air–sea interactions near the surface. This mechanism is potentially important for sustaining submesoscale energy and enabling the inverse energy cascade. For oceanic baroclinic instability, the strongest Rossby waves thereby tend to propagate downward away from the source region, resulting in the downward energy transport via the vertical pressure work.

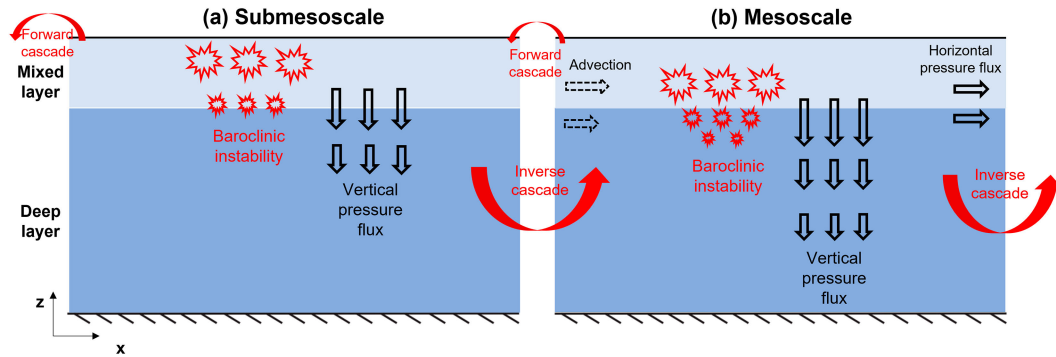


FIG. 12. Schematics of the KE budget for (a) submesoscale and (b) mesoscale motions. The horizontal (vertical) axis represents horizontal distance (depth). Red star-shaped patches indicate KE generation by baroclinic instability, which is primarily confined to the mixed layer at submesoscales but extends deeper in the water column at mesoscales. Red curved arrows denote cross-scale energy cascades, including forward cascades from submesoscale to smaller scales, inverse cascades from submesoscale to mesoscale, and forward cascades from mesoscale to submesoscale. Black solid arrows indicate energy transport by PW. Horizontal PW is significant only at the mesoscale, whereas vertical PW transports energy downward: It is largely confined to the mixed layer at submesoscales but penetrates to the deeper ocean at mesoscales. Black dashed arrows represent horizontal advective energy transport, which is considerable only at the mesoscale.

Previous studies have identified multiple pathways through which the energy is cascaded to smaller scales, including frontogenesis, wind-eddy interactions, and catalysis by inertial gravity waves (Renault et al. 2016; Barkan et al. 2017, 2021; Srinivasan et al. 2023), all of which are strongest near the surface. However, by transporting energy beneath the mixed layer, vertical pressure work reduces exposure to these surface dissipative processes and thereby strengthens the inverse energy cascade. Understanding this mechanism is essential for quantifying how energy is retained and redistributed within the ocean, as well as for predicting the evolution and persistence of submesoscale processes.

This study focuses primarily on the subtropical northwest Pacific, a region with high EKE, but the role of pressure work in the KE budget may differ across oceanic regimes. Using a global 0.1° OGCM, Wang et al. (2019) demonstrated that PW plays distinct roles in the spectral EKE budget in the equatorial regions and midlatitudes. Similarly, Zhai and Marshall (2013) found that the vertical pressure flux associated with mesoscale eddies is downward (upward) in the subtropical (subpolar) gyre of the North Atlantic. These findings suggest the regional dependence of the role of PW in energy redistribution. A more comprehensive understanding of the PW and its regionality, therefore, requires further examination of the scale-dependent KE budget across diverse oceanic regimes, which will be pursued in future work.

While the model's horizontal resolution is sufficient to capture submesoscale features related to mixed-layer instability (Sasaki et al. 2017), it remains coarse to fully resolve the finest submesoscale instabilities, such as symmetric instability (Dong et al. 2021). This potentially leads to an underestimation of their contribution to the energy budget, especially the forward energy cascade. In addition, the use of daily mean outputs filters out most of the near-inertial internal waves, although some of their effects may remain, as the period of

near-inertial waves varies with latitude. Hence, wave–eddy interactions, which contribute significantly to the forward energy cascade, may not be adequately assessed in this study (Barkan et al. 2021, 2024; Delpech et al. 2024). Moreover, considering the long lifetime of mesoscale eddies in the northwest Pacific (Chelton et al. 2011), the 1-yr-long data used in this study may not be sufficient to obtain a statistical equilibrium for the time-averaged results. Future work will, therefore, employ higher-resolution models, leverage outputs with finer temporal sampling to better represent high-frequency dynamics, and incorporate high-resolution observational datasets, such as the Surface Water and Ocean Topography (SWOT) products. These advances will provide a more complete and robust characterization of the scale-dependent energetics of oceanic flows.

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Data availability statement. The OFES2 model outputs can be requested from JAMSTEC (<https://www.jamstec.go.jp/ofes/ofes>).

html). Multiobservation global ocean 3D temperature, salinity, height, geostrophic current, and mixed layer depth can be downloaded from CMEMS (<https://doi.org/10.48670/moi-00052>).

APPENDIX

Linear Instability Analysis

Linear instability analysis is performed to investigate the role of vertical pressure work. The linearized quasigeostrophic potential vorticity (QGPV) equation and associated boundary condition writes (Smith 2007; Tulloch et al. 2011)

$$\begin{cases} q_t + \mathbf{U} \cdot \nabla q + \mathbf{u}' \cdot \nabla Q = 0, \\ \psi_{zt} + \mathbf{U} \cdot \nabla \psi_z + \mathbf{u}' \cdot \nabla \Psi_z = 0, \quad z = -H, \quad 0, \end{cases} \quad (\text{A1})$$

where $q = \nabla^2 \psi + \Gamma \psi$ is the eddy QGPV [$\Gamma = \partial_z[(f^2/N^2)\partial_z]$ is the vortex stretching operator], $Q = \beta y - (g/\rho_0)\partial_z(f\rho/N^2)$ is the mean PV (N is the buoyancy frequency), $\mathbf{u}' = (u', v') = (-\psi_y, \psi_x)$ is the eddy-induced velocity with the eddy streamfunction ψ , and $\mathbf{U} = (U, V)$ is the background velocity related to the mean streamfunction Ψ . Substituting a wave solution in the form of $\psi(x, y, z, t) = \Phi(z)e^{i(kx + ly - \omega t)}$ [$\mathbf{K} = (k, l)$ and ω are the wavenumber and frequency, respectively] to Eq. (A1) gives

$$\begin{cases} (\mathbf{K} \cdot \mathbf{U} - \omega)(\Gamma - K^2)\Phi = -(kQ_y - lQ_x)\Phi, \\ (\mathbf{K} \cdot \mathbf{U} - \omega)\Phi_z = (kU_z + lV_z)\Phi, \quad z = -H, \quad 0, \end{cases} \quad (\text{A2})$$

which is an eigenvalue problem for the normal mode $\Phi(z)$. The solution having a nonzero imaginary part of frequency denotes the unstable wave that grows exponentially, with a growth rate of $\text{Im}(\omega)$.

We solve the eigenvalue problem for Eq. (A2) numerically using finite-difference discretization for each given wavenumber, the details of which can be found in Smith (2007) and Tulloch et al. (2011). The velocity and density profiles from OFES2_NP30 are interpolated onto a uniform grid with a vertical resolution of 5 m. Subsequently, the discrete version of Eq. (A2) is solved using the eigen function in MATLAB.

With the obtained eddy streamfunction, the eddy vertical velocity w' , pressure p' , and buoyancy perturbation b' can be computed as

$$\begin{cases} w' = -\frac{if}{N^2}[(\mathbf{K} \cdot \mathbf{U} - \omega)\Phi_z - (kU_z + lV_z)\Phi], \\ p' = f\Phi, \\ b' = f\Phi_z. \end{cases} \quad (\text{A3})$$

Thereafter, the buoyancy production and eddy pressure flux can be calculated as $(1/2)\text{Re}(w'b'^*)$ and $(1/2)\text{Re}(w'p'^*)$, respectively, where $*$ represents the complex conjugate.

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