Divergent Eddy Heat Fluxes in the Kuroshio Extension at 144°–148°E.
Part II: Spatiotemporal Variability

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ABSTRACT
The Kuroshio Extension System Study (KESS) provided 16 months of observations to quantify divergent eddy heat flux (DEHF) from a mesoscale-resolving array of current- and pressure-equipped inverted echo sounders. KESS observations captured a regime shift from a stable to unstable state. There is a distinct difference in the spatial structure of DEHFs between the two regimes. The stable regime had weak downgradient DEHFs. The unstable regime exhibited asymmetry along the mean path with strong downgradient DEHFs upstream of a mean trough at ~147°E. The spatial structure of DEHFs resulted from episodic mesoscale processes. The first 6 months were during the stable regime in which fluxes were associated with eastward-propagating 10–15-day upper meanders. After 6 months, the Kuroshio Extension underwent a regime shift from a stable to unstable state. This regime shift corresponded with a red shift in mesoscale phenomena with the prevalence of ~40-day deep externally generated eddies. DEHF amplitudes more than quadrupled during the unstable regime. Cold-core ring (CCR) formation, CCR–jet interaction, and coupling between ~40-day deep eddies were responsible for asymmetry in downgradient fluxes in the mean maps not observed during the stable regime. The Kuroshio Extension has prominent deep energy associated with externally generated eddies that interact with the jet to drive some of the biggest DEHF events. These eddies play an important role in the variability of the jet through eddy–mean flow interactions. The DEHFs that result from vertical coupling act in accordance with baroclinic instability. The interaction is not growth from an infinitesimal perturbation, but from the start is a finite-amplitude interaction.

1. Introduction
The Kuroshio, the western boundary current (WBC) of the Northwest Pacific, plays a critical role in redistributing heat poleward from the tropics in order to balance the global energy budget. The Kuroshio leaves from the coast at ~35°N and flows east as a free jet, called the Kuroshio Extension. The Kuroshio Extension is characterized by vigorous meandering and variability on a range of time scales from the mesoscale (from days to months) to interannual (Mizuno and White 1983). The mesoscale is the most energetic, and satellite altimetry observations point to the Kuroshio Extension region as one of the regions of highest eddy kinetic energy (EKE) in the world’s oceans (Ducet and Le-Traon 2001). Mesoscale eddies drive heat fluxes that transport heat poleward across strong current systems, which otherwise primarily take a zonal path and act as barriers to cross-front flow.

In Part I [Bishop et al. (2013), hereafter Part I], using current- and pressure-equipped inverted echo sounder (CPIES) observations taken during the Kuroshio Extension System Study [KESS; see Part I and Donohue et al. (2010) for details on the KESS experiment], 16-month mean maps of divergent eddy heat fluxes (DEHFs) were quantified. There was asymmetry in cross-front DEHFs along the mean path with strong downgradient fluxes upstream of a mean trough at ~147°E. The mean vertical structure of the DEHFs had subsurface maxima within the main thermocline ~400 m. Downgradient parameterizations with constant eddy diffusivities ranging from 800 to 1400 m^2 s^-1 fit the vertical structure of the data upstream of the trough, but did not account for weak upgradient fluxes downstream of the mean trough.

The mesoscale variability in the first 1000 km to the east of Japan, observed from satellite altimetry, modulates on decadal time scales between stable and unstable regimes characterized by minimal and vigorous meandering, respectively (Qiu and Chen 2005b). Associated
with enhanced eddy interaction during unstable regimes is a weakening of the southern recirculation gyre and subsequent reduction in alongfront transport. With the exception of Hall (1991), who found that the Kuroshio Extension at 152°E had significant energy conversion from mean to eddy potential energy (EPE) on the anticyclonic side of the current at 350 and 625 dbar, the role of eddy heat fluxes in eddy–mean flow interactions throughout the water column and between meandering regimes remain unknown in the Kuroshio Extension from observations. The KESS experiment fortuitously observed a regime shift from a stable to unstable meandering regime.

In addition to decadal variability in meandering states, recent observations from KESS revealed that there are mechanisms for cyclogenesis present in the Kuroshio Extension (Greene et al. 2009) that differ from what was observed in the Gulf Stream during the Synoptic Ocean Prediction Experiment (SYNOP) in the 1980s. Greene et al. (2012) found that during KESS, the strongest variability in the abyssal ocean was due to 30–60-day topographically controlled eddies that had formed outside the observational array. The origin of these eddies was postulated to be the Shatsky Rise downstream at ~160°E, and model outputs from the Ocean General Circulation Model for the Earth Simulator (OFES) support this claim (Greene 2010). In other frequency bands, Tracey et al. (2011) found that vertical coupling between southwestward-propagating deep eddies and downstream-propagating upper-ocean meanders resulted in the growth of upper meanders, depending upon the phasing between the deep and upper ocean.

The goal of this paper is to determine among the varied mesoscale phenomena observed during KESS the following:

(i) The features responsible for driving the largest DEHFs that make up the 16-month mean spatial structure in Part I.
(ii) To characterize the differences in the DEHF spatial structure between stable and unstable meandering regimes.
(iii) Based on Greene et al. (2012) and Tracey et al. (2011), the role played by deep external eddies in developing the mean structure of the DEHFs.

The paper is organized as follows. The next section will outline the data and method for determining DEHFs. Next, the spatiotemporal variability results will be presented, followed by a ring census using satellite altimetry data. Next, the connection to deep external eddies will be shown with a case study of a CCR formation event. Last, the paper will explore eddy energetics and end with a section of discussion and summary.

2. Methods

a. Data

Here, 46 CPIES were deployed in a ~600 km × 600 km array spanning the Kuroshio Extension jet for 2 years during KESS (Fig. 1). The CPIES array was centered in the region of highest surface EKE from satellite altimetry (143°–149°E) and spanned the meander envelope from north to south, capturing almost one full wavelength of the quasi-stationary meander crest–trough–crest to the east of Japan (Mizuno and White 1983). The CPIES array had nominal horizontal spacing of 88 km, to resolve mesoscale variability. Here, 26 of the CPIES were collocated with Jason-I altimetry lines for comparative studies (Park et al. 2012; Bishop et al. 2012).

Using observations from the CPIES array, mapped fields of the geostrophic current \( \mathbf{u} \), temperature \( T \), and density \( \rho \) were determined throughout the water column, which is well documented in Donohue et al. (2010), and summarized in Part I. Geostrophic currents determined with the CPIES express the vertical structure as the sum of an equivalent-barotropic internal mode \( \mathbf{u}_I \) and a nearly depth-independent external mode \( \mathbf{u}_E \) measured in the deep ocean. The absolute geostrophic currents are the summation of the two modes

\[
\mathbf{u} = \mathbf{u}_I + \mathbf{u}_E.
\]

b. Divergent eddy heat flux

Eddy heat flux is defined as the temporal correlation between the geostrophic currents and temperature field and multiplied by the spatially and depth-averaged density \( \rho_0 \) and specific heat at constant pressure \( C_p \)

\[
\rho_0 C_p \mathbf{u} \overline{T'} = \rho_0 C_p \mathbf{u}_I \overline{T'} + \rho_0 C_p \mathbf{u}_E \overline{T'},
\]

where an overbar indicates a time mean and a prime indicates a deviation from the time mean (e.g., \( \mathbf{u}' = \mathbf{u} - \mathbf{u}_c \)); \( \rho_0 = 1027.5 \text{ kg m}^{-3} \) for the KESS observations and \( C_p \approx 4000 \text{ J kg}^{-1} \text{ C}^{-1} \) for seawater. Equation (2) has units of watts per square meter and is commonly reported in the literature with units of kilowatts per square meter. Eddy heat flux for the CPIES maps using Eq. (1) for the geostrophic currents is

\[
\mathbf{u} \overline{T'} = \mathbf{u}_I \overline{T'} + \mathbf{u}_E \overline{T'},
\]

where multiplication by \( \rho_0 \) and \( C_p \) is implied.
literature (Cronin and Watts 1996; Jayne and Marotzke 2002; Marshall and Shutts 1981). In Part I, it was determined that the eddy heat flux associated with the internal mode geostrophic currents $u_{IT}$ is completely rotational (nondivergent) and proportional to the lateral gradient of temperature variance $T_0^2$, as in Marshall and Shutts (1981). Eddy heat flux associated with the external mode currents $u_{ET}$ contains the divergence. However, $u_{ET}$ is not completely rotation free. Objective analysis (OA) was used to further decompose the eddy heat flux field into divergent and rotational components, which uses nondivergent correlation functions to map the best fit nondivergent vector field to $u_{ET}$. The divergent component of the eddy heat flux is determined by taking the difference between the full vector field and the best fit nondivergent field from the OA

$$\bar{u}_{ET}^\text{div} = \bar{u}_{ET} - \bar{u}_{ET}^{\text{OA}}. \quad (4)$$

The subscript $E$ has been dropped from the eddy velocity term for the divergent flux for convenience. Results using OA to remove rotation in Part I showed that the $u_{ET}$ field was mostly divergent, but that it was necessary to remove small rotational effects to correctly diagnose eddy diffusivity.

3. Spatiotemporal variability in divergent eddy heat flux

a. Spatial structure

DEHFs [Eq. (4)] in the Kuroshio Extension region exhibit a different spatial structure depending on whether it is in a stable or unstable meandering regime. The horizontal structure of the vertically integrated cross-front DEHF between the base of the sea surface mixed layer and deep reference depth

$$\rho_0 C_p \int_{-H_{\text{ref}}}^{H_{\text{Ek}}} n \cdot \bar{u}_{ET}^\text{div} dz, \quad (5)$$

where $H_{\text{Ek}} = 100$ m, $H_{\text{ref}} = 5300$ m, $n = |\nabla T|^{-1} \nabla T$ is the cross-front unit vector, and $\nabla = (\partial_x, \partial_y)$ is the horizontal gradient operator, is presented in Fig. 2 for the 16-month,
stable, and unstable regimes. The vertically integrated eddy heat flux field resembles the horizontal structure at all depths because the advecting velocity is independent of depth and the frontal temperature structure is vertically aligned with depth. Superimposed is the depth-averaged baroclinic conversion (BCdiv)

$$BC_{\text{div}} = -\frac{\alpha g}{\theta_z} W^{\text{div}} \cdot \nabla T,$$

where $\alpha$ is the effective thermal expansion coefficient estimated from historical hydrography as $1.74 \times 10^{-4} \degree C^{-1}$, and $g$ is the acceleration due to gravity. Positive BCdiv is an estimate of the energy conversion from mean potential energy (MPE) to EPE with units of square centimeters per second cubed.

The spatial structure of the 16-month mean eddy heat flux is dominated by the unstable regime where Figs. 2a and 2c have similar features. DEHF is strong and downgradient upstream of the mean trough at $147^\circ$E along the mean path.

The stable regime (Fig. 2b), however, is characterized by weaker fluxes that are mostly downgradient along the mean path. The region of strongest downgradient fluxes in the stable regime shifted upstream compared with the unstable regime. The upstream shift was due to the presence of low-amplitude frontal meander growth, which will be discussed in an upcoming section. Another difference between regimes is that the stable regime did not have any CCRs present. CCR formation events and their relation to eddy heat flux will too be discussed further in an upcoming section. South of the jet there are strong equatorward heat fluxes in Figs. 2a and 2c from the passage of a CCR, which cooled the surrounding waters. It was shown in Part I that this feature was associated with convergence of eddy heat fluxes that induced upwelling.

Mean–eddy energy conversion is different between meandering regimes. The latitudinal dependence of the zonally averaged BCdiv across the KESS array in Figs. 2a–2c is shown in Fig. 2d. On average, BCdiv is positive along the mean path between $34^\circ$ and $36^\circ$N indicating that the eddies release available potential energy of the mean flow. During the stable regime, BCdiv peaks at $0.5 \times 10^{-3} \mathrm{cm^2 s^{-3}}$ at $35^\circ$N. In contrast to the stable regime, BCdiv is more than two times larger during the unstable regime reaching a maximum $1.25 \times 10^{-3} \mathrm{cm^2 s^{-3}}$ at $35^\circ$N. The range of BCdiv between stable and unstable meandering regimes along the mean path ($0.5–1.25 \times 10^{-3} \mathrm{cm^2 s^{-3}}$) when equated with EKE is sufficient to produce a depth-averaged eddy velocity of 9–15 cm s$^{-1}$ after one day. Negative conversion is observed to the south of the jet near $33^\circ$N, but only during the unstable regime. This is consistent with upwelling and cooling in the region owing to the passage of a CCR as described above.

b. Meridional eddy heat transport

Differences between the full and divergent 16-month mean meridional eddy heat transport was examined in Part I in context of past studies (Qiu and Chen 2005a; Volkov et al. 2008; Jayne and Marotzke 2002). In this section, the divergent meridional eddy heat transport variability between stable and unstable regimes is...
explored. This is a natural question considering that there is enhanced eddy variability during the unstable regime. It is expected that this would result in larger eddy heat transport across the Kuroshio Extension front.

The total divergent meridional eddy transport was quantified by vertically and zonally integrating the meridional DEHF:

$$Q_{\text{div}}^{\text{tot}} = \rho_0 C_p \int_{x_i}^{x_e} \int_{z=H_{\text{ref}}}^{H_{\text{ref}}} u' T'^{\text{div}} \, dz \, dx,$$

(7)

where $x_i$ and $x_e$ are the beginning and end longitudes, respectively, that are a function of latitude based on the KESS array grid. Here, $Q_{\text{div}}^{\text{tot}}$ is shown as a function of latitude across the mean jet path from 34.25° to 36.25°N in Fig. 2d for the 16-month, stable, and unstable averaging time periods. The stable and unstable regimes had $Q_{\text{div}}^{\text{tot}}$ at ~32.25°N that reached a maximum of 0.02 and 0.06 PW, respectively. There is substantial variability between meandering regimes with a threefold increase in $Q_{\text{div}}^{\text{tot}}$ from the stable to unstable meandering regime.

c. Time series

Time series of cross-front DEHF in the Kuroshio Extension reveal stark differences in the episodic nature between the stable and unstable regimes. Time series of the cross-front eddy heat flux at 400-m depth are shown in Fig. 3a for four locations where the mean fluxes were strongest along the mean path plus one location to the south of the jet. The spatial locations are numbered in Fig. 3b. Here, 400 m was chosen because this is approximately the middepth of the thermocline and the mean fluxes were observed to be strongest at this depth in Part I.

1) STABLE REGIME

The first six months of observations, from June to November 2004, are characterized by the stable regime (Fig. 1b) in which the cross-front eddy heat flux at 400-m depth reached a maximum of 0.02 and 0.06 PW, respectively. There is substantial variability between meandering regimes with a threefold increase in $Q_{\text{div}}^{\text{tot}}$ from the stable to unstable meandering regime.
heat flux structure in Fig. 2b with weak heat fluxes that are downgradient. Tracey et al. (2011) showed that this regime was dominated by small-amplitude, ~50-km peak–peak lateral displacement, upper-baroclinic frontal waves with periods of 10–15 days. These waves traversed the KESS array from west to east with phase speeds of 20–25 km day\(^{-1}\) and wavelengths of 200–300 km that exhibited meander growth at times when coupling to deep eddies. Signatures of the growth of these waves are particularly noticeable at sites 1 and 2 in the cross-front DEHFs between June and mid-August 2004 (yeardays 150–225).

2) UNSTABLE REGIME

After yearday 300, the Kuroshio Extension transitioned from a stable to unstable meandering regime in which it remained for the subsequent 10 months of observations (Fig. 1c). During the unstable regime, longer period meanders in the 30–60-day band began to propagate into the KESS array traveling from west to east with average periods of 43.5 ± 3.7 days, propagation speeds of 9.6 km day\(^{-1}\) (7.8–12.1 km day\(^{-1}\)), and wavelengths of 418 ± 60 km (Tracey et al. 2011). The origin of these longer period meanders is speculated to have resulted from two possible mechanisms. One mechanism is that the jet interacted with a warm-core ring (WCR) at the first quasi-stationary meander crest to the east of Japan nearly 50 days prior to the regime shift. The second mechanism is the jet inflow at the Izu–Ogasawara Ridge. Qiu and Chen (2005b) argue that the jet shifts to the south during unstable regimes. When the jet shifts south, the inflow over the Izu–Ogasawara Ridge flows through a shallow segment, which leads to large-amplitude downstream meanders. When the longer period upper meanders appeared, simultaneously trains of deep, externally generated eddies in the 30–60-day band, with a nominal period of ~40 days, began propagating into the KESS array from the east-northeast and turned, approximately following bathymetry contours, to travel down the central line from the northeast to the southwest. These eddies interacted with the jet. The origin of the deep externally generated eddies is thought to be the Shatsky Rise downstream near 160°E and is discussed in Greene (2010).

During the unstable regime, there were three major events and processes that drove significant cross-front DEHFs along the mean path of the jet: one cold-core ring (CCR) formation, CCR–jet interaction, and deep external eddies, which were present throughout the unstable regime. From late November 2004 to February 2005 (yeardays 325–400), the eastward propagation of an upper-baroclinic wave was stunted, and the meander trough steepened until a CCR formed (CCRa). During this interval, there was an abrupt change in the eddy heat flux time series with eddy heat flux amplitudes more than quadrupling (Fig. 3a). The mean spatial structure of the eddy heat fluxes during CCRa formation shows strong poleward fluxes along the mean path just upstream of diffuence in the mean streamlines where the ring formed (Fig. 3c). Downstream of this region, there are weak upgradient fluxes. To the south of the mean path at the southern extremity of the CCRa formation, there are strong equatorward fluxes, but the mean–eddy energy conversion (i.e., BCdiv) is only weakly negative because the mean lateral temperature gradient is weak there (Fig. 2c). These fluxes resulted from the coupling of deep eddies with the newly formed ring, which produced the only large event during the site’s five time series (Fig. 3a).

After CCRa completely detached, another CCR (CCRb) that had formed to the east near the Shatsky Rise (~160°E) in the previous year propagated into the southeastern region of the KESS array and interacted with the jet (illustrated in Figs. 1b and 1c). CCRb was briefly absorbed by the jet and repinched off to reside to the south of the jet for 6 weeks until it was permanently reabsorbed into the jet. This ring–jet interaction occurred from March to July 2005 (yeardays 425–550) and the time series of eddy heat flux events is largest during the latter part of the CCR–jet interaction (yeardays 500–550). Amplitudes are similar to that during the CCRa formation. The mean spatial structure of the eddy heat flux during the ring–jet interaction is also very similar to the CCRa formation event with strong poleward fluxes along the mean path upstream of the trough (Fig. 3d). Near the end of the 16-month observations, there is a train of deep eddies in the 30–60-day band that propagated into the array from the northeast and traveled to the southwest. These deep eddies interacted with the jet causing strong poleward heat fluxes near 35°N, 145°E (sites 1 and 2) and 34°N, 147°E (site 3 and 4) along the mean path (Fig. 3e).

The results from this section suggest that there are varied mesoscale phenomena that produce cross-front DEHF events, particularly associated with CCRs. The region upstream of the mean trough at 147°E is favored for downgradient fluxes. The extrema in the DEHF time series may be astride the preferred path that deep external eddies take, following \(f/h\) contours where \(h\) is the fluid depth, down the central line of the KESS array (Greene et al. 2012), and will be discussed in more detail in an upcoming section.

d) Cross spectra

The variance-preserving cross spectra between the meridional external mode currents \(v_E\) and the temperature
field (i.e., $T$) at 400 m for the same locations in Figs. 3a and 3b provide a useful measure of the spectral content of the meridional eddy heat fluxes (Bryden 1979; Phillips and Rintoul 2000). Figure 4 displays a shift to lower frequencies in spectral content between the stable and unstable regimes, which is in agreement with the time series discussion of the previous two subsections. The eddy heat flux estimate at each location is retrieved by integrating the cross-spectrum

$$
\rho_0 C_p \overline{\nu E T} = \rho_0 C_p \int_0^\infty P_{\nu E T} \, d\nu,
$$

where $\nu$ is the frequency, and $P_{\nu E T}$ is the cross-spectral density between $\nu E$ and $T$.

During the stable regime, the meridional eddy heat flux has peaks at 30 days and shorter (Fig. 4b). During the unstable regime, peaks in the meridional eddy heat flux were more energetic and shifted to periods greater than 30 days with energy concentrated around 40 days and longer (Fig. 4c). The 16-month meridional eddy heat flux spectral content was dominated by variability during the unstable regime (Fig. 4a).

4. Ring census

In the last section, it was shown that large DEHFs are associated with CCR formation. A CCR formation census was done for a 15-yr period between 1993 and 2007 at the longitudes of 140°–165°E to see the representativeness of the 16 months of KESS observations of stable and unstable regime characteristics in the spatial structure of DEHF. The AVISO satellite altimetry product merged with RIO05 mean dynamic topography provided weekly snapshots of the Kuroshio Extension SSH path to do this study. The criterion to count a ring formation was to manually follow the 2.1-m SSH contour, which is representative of the jet axis and approximately equal to where the 12°C isotherm crosses the 300-m isobath (Mizuno and White 1983). When a trough developed a closed contour to the south of the jet axis and persisted for more than two weeks as a closed contour, this was considered a CCR formation. When a CCR formed, the location of the minimum SSH within the closed contour was considered to be the formation location. Stable and unstable regimes were distinguished after the work of Qiu and Chen (2010): relatively stable periods were 1993–95 and 2002–05, and unstable regimes were 1996–2001 and 2005–07.

A total of 40 CCRs formed between 1993 and 2007. Figure 5a shows the geographic distribution of CCR formations superimposed on topography. There is a gap near 150°E where no CCRs formed, which is the mean location of the second quasi-stationary meander crest to the east of Japan. Within the KESS region (upstream of 150°E), 16 CCRs formed of which only 1–3 formed during stable regimes. Figure 5b is a time series from the end of 1992 to mid-2007 of EKE from the AVISO altimetry product with RIO05 mean dynamic topography and spatially averaged over the KESS array dimensions, 32°–38°N and 143°–149°E. The stable periods were marked as defined in Qiu and Chen (2010), and the times when CCRs formed are marked. One CCR that formed during a stable regime was the ring during KESS and marks the transition from stable to unstable regimes at the end of 2004. Another CCR formed in 1993. It can be argued that this CCR marked the transition from an unstable to
stable regime because EKE levels were well above the mean during its formation, but dropped off dramatically, staying below the mean until mid-1995. From a closer visual examination of the biweekly paths of the Kuroshio Extension in Qiu and Chen (2010), 1993 was a year in transition from an unstable to stable regime and still had relatively high levels of EKE (Fig. 5b), which were above the average. It was not until the end of 1993 that the path was more stable and EKE values dropped, remaining below the mean. With that said, only one CCR formed within a stable regime in July 2002 (Fig. 5). After that ring formed, a CCR did not form in the KESS region for nearly 3 years from July 2002 to January 2005. The KESS observations during the stable regime were only six months, but this analysis suggests that it may be representative of the previous two years during the stable regime.

Downstream of 150°E, 24 CCRs formed between 1993 and 2007. From Fig. 5, more than half of the rings (16 of 24 CCRs) that formed downstream are congregated around the Shatsky Rise (155°–161°E). This suggests that the mechanism of flow–topography interaction may promote CCR formation downstream. In the region downstream of 150°E, 10 CCRs formed during stable regimes versus 14 that formed during unstable regimes.

5. Connection to deep external eddies

A mechanism that drives large DEHF is the interaction between external eddies and the Kuroshio Extension jet. Tracey et al. (2011) found that the deep ocean during KESS was populated with westward-translating eddies that were generated external to the KESS array in frequency bands of 3–60 days. They found that upper meanders and deep eddies jointly intensified when encountering each other if the phasing was right between the deep and upper ocean with the deep eddy offset $\frac{1}{4}$ of a wavelength ahead of the upper meander.

The strongest deep eddies, in the most energetic band (30–60 days), propagated down the central line of the KESS array from the northeast to the southwest at propagation speeds of 10–20 km day$^{-1}$ and interacted with the upper jet (Greene et al. 2012). These deep eddies were absent during the first six months when it was the stable meandering regime and weak cross-front DEHFs were observed. Deep eddies began to propagate down the central line in late October 2004 and were present thereafter. A train of deep eddies in the 30–60-day band may be responsible for the formation of the CCR near the end of 2004, but this merits further research.

The passage of 30–60-day deep eddies across the jet path is correlated with events in the cross-front DEHF time series (Fig. 6). After yearday 300, a train of deep eddies propagated down the central line and coupled with the jet. This coupling caused troughs and crests to grow. The cross-front DEHFs more than quadrupled in amplitude along the mean path during the passage of these deep eddies. Two significant events occurred that involved these deep eddies: a CCR formation and a deep
eddy–jet coupling event near the end of the 16-month observations. The CCR case will now be looked at in more detail.

**Cold-core ring formation case study**

Figure 7 shows the DEHF vectors at 400 m from mid-November 2004 to late January 2005 with the 30–60-day bandpass-filtered bottom pressure anomaly and upper geopotential anomaly. A wave train of highs and lows in the deep propagated from the northeast to the southwest down the central line of the array during this time. During the passage of these deep eddies, there was joint development between the deep and upper anomalies. Consequently, the jet axis steepened into a trough. Associated with the trough development were strong downgradient DEHFs within the trough and upstream from yeardays 325–345. The ring briefly detached and reattached between yeardays 350 and 365 before it finally detached completely. Interestingly though, cross-front DEHFs were weak during this reattachment and detachment process.

The interaction of the upper jet and deep eddies in the 30–60-day band has characteristics of baroclinic instability, but is a finite-amplitude interaction from the start. Time series of 30–60-day bandpass-filtered upper geopotential anomaly referenced to 5300 m $\Phi'_I$ and the bottom reference geopotential anomaly $\Phi'_E = p_b/p_h$, where $p_b$ is the pressure anomaly at 5300 m and $p_h$ is the density at 5300 m at location 3 (Fig. 7), are plotted in Fig. 8. There is growth of both $\Phi'_I$ and $\Phi'_E$ during the CCR formation (yeardays 300–370). In Fig. 8b, $\Phi'_I$ and $\Phi'_E$ have a 40-day periodicity with the deeper leading the upper by 8 days. If the interaction is thought of as baroclinic instability of the two-layer Phillips model, then the deep is leading the upper by a phase shift of $\phi = 2\pi/5$. The phase needs to be $\pi > \phi > 0$ for growth, for which $\phi = 2\pi/5$ is close to optimal growth ($\phi = \pi/5$).

The jet–deep eddy interaction is also a heat advection process, which agrees with baroclinic instability (Cronin and Watts 1996). Associated with the jet–deep eddy interaction is cold and warm advection during growth, which can be seen from maps of the vertical velocity $w$ during the CCR formation (Fig. 9). The vertical velocity within the thermocline was estimated from variations in the depth of the $12^\circ C$ isotherm $Z_{12}$ by

$$w = \frac{\partial Z_{12}}{\partial t} + \mathbf{u} \cdot \nabla Z_{12}.$$  

which has been shown to be consistent with meanders and float observations in the Gulf Stream (Lindstrom and Watts 1994; Howden 2000). Note that $Z_{12}$ increases negatively from the surface downward so that $Z_{12} < 0$. Between yeardays 320 and 345, there is downwelling along the Kuroshio Extension path leading into the trough and upwelling leaving the trough. The vertical velocities within the trough are on the order of $\pm 100$–150 m day$^{-1}$ (1–2 mm s$^{-1}$), which are weaker than the observed vertical velocities during trough events in the
Gulf Stream (2–3 mm s$^{-1}$; Howden 2000). The downwelling within the trough is in precisely the same location as strong poleward eddy heat fluxes (yeardays 330–340 in Fig. 7). It is clear from the 16-month power spectrum of $w$ at the four locations in Fig. 9 that the tendency term in Eq. (9) is dominated by variability at frequencies higher than 1/10 day$^{-1}$ (Fig. 10a). Most of the variance associated with the advective term in Eq. (9) is at frequencies lower than 1/30 day$^{-1}$ (Fig. 10b). The advection of $Z_{12}$ is due to deep eddies because the equivalent-barotropic internal mode currents (i.e., $u_I$) are perfectly aligned with $Z_{12}$ contours

$$u \cdot V_{12} = u_E \cdot V_{Z_{12}}. \quad (10)$$

The cold advection within the trough during the CCR formation is due to the passage of a deep cyclone in the 30–60 day band that had a $\sim 2\pi/5$ phase offset leading the upper field. The cyclone advected the Kuroshio Extension front southward, causing the subthermocline layer to be stretched and further intensified the deep cyclone [see Greene et al. (2012) for details of this vertical coupling mechanism].

This CCR formation event is analogous to a cutoff low-blocking pattern in the jet stream. As the jet axis steepened and the trough aspect ratio approached $O(1)$, upper geopotential anomalies in the 30–60 day band were unable to propagate east past the region of large
diffluence. The geopotential anomalies were essentially blocked and turned to the south after yearday 305; following the deep 30–60-day eddies south into the subtropical gyre (Figs. 7 and 9). It is not until the CCR had completely pinched off that upper geopotential anomalies were able to propagate east again. The DEHF during this blocking event were strong and down-gradient within and just upstream of the region of

![Graph showing time series of geopotential anomalies](image)

**Fig. 8.** (a) The 16-month time series of 30–60-day bandpass-filtered geopotential anomaly referenced to 5300 m, $\Phi_E$, and bottom-referenced geopotential anomaly $\Phi_E$ at location 3 in Fig. 7. (b) Zoomed-in time series in (a) during the CCR formation.

![Images of vertical velocities during CCR formation](image)

**Fig. 9.** Vertical velocities during the CCR formation with 30–60-day bandpass-filtered upper geopotential anomaly (bold light and dark gray contours are positive and negative anomalies respectively; CI = 1 m$^2$/s$^2$), and geopotential referenced to 5300 m (thin gray contours; CI = 1 m$^2$/s$^2$) with the bold black contour marking the jet axis (38.52 m$^2$/s$^2$ geopotential contour).
diffluence. The spatial patterns of DEHF are consistent with other studies of blocking patterns and eddy fluxes in the atmosphere. DEHFs are strong and downgradient upstream of diffluence and may be reinforcing the block by slowing down the mean flow approaching the diffluence (Pierrehumbert 1986; Shutts 1986; Illari and Marshall 1983).

6. Eddy energetics

In the previous section, it was shown that the interaction of deep external eddies with the Kuroshio Extension jet have many characteristics of baroclinic instability. Downgradient DEHFs are only one piece of the puzzle in baroclinic instability. Downgradient DEHFs are a measure of the eddies drawing eddy potential energy from the mean potential energy of the Kuroshio Extension jet sloping isopycnals. Vertical eddy heat fluxes are what lead to an increase in EKE, which are a measure of the energy conversion from EPE to EKE in accordance with baroclinic instability. In this section, the balance between cross-gradient DEHFs and vertical eddy heat fluxes (Marshall and Shutts 1981)

\[
\begin{align*}
-\frac{\alpha g}{\Theta_z}u^T \cdot \nabla T &= \alpha g w/T^2
\end{align*}
\]

will be examined. The latitudinal structure of the zonally averaged BCdiv and PKC in Eq. (11) will be compared between regimes and their temporal correlation will also be quantified.

The zonally averaged EKE, BCdiv, and PKC are shown in Figs. 11a–11c. The EKE is peaked near the mean path of the jet at \( \sim 34.5^\circ \)N and more than doubles between the stable and unstable regimes (Fig. 11a). The BCdiv tends to be larger than the PKC, but they have a similar latitudinal structure with positive zones near the mean jet at \( \sim 35^\circ \)N. Latitudinal correlation coefficients between the zonally averaged BCdiv and PKC are 0.85, 0.96, and 0.83 for the 16-month, stable, and unstable regimes, respectively.

To explore the temporal variability of BCdiv and PKC, time series of BCdiv' and PKC', where a prime on BCdiv and PKC indicates that a time average has not been taken, are plotted in Fig. 11d. BCdiv' and PKC' were spatially averaged between 32° and 37.5°N and between 143° and 149°E. In estimating PKC', it must be stressed that only the advective portion of the vertical velocity field \( u \cdot V_{12} \) was considered. The tendency term vanishes for sufficiently long time averaging. This is displayed by separating Eq. (9) into mean and eddy terms, multiplying by the eddy temperature \( T' \) and taking a time average

\[
T' \frac{\partial Z_{12}'}{\partial t} = 0,
\]

This relation holds because there is a nearly linear relationship between \( T' \) and \( Z_{12} \) (Fig. 12)

\[
T' = mZ_{12} + b,
\]

where \( m \) is the slope, and \( b \) is the y intercept of the \( T' \) versus \( Z_{12} \) scatterplot. Figure 12a is a plot of the mean...
temperature at 400 m versus the mean depth of the 12°C isotherm and Fig. 12b is a similar plot, but of the eddy temperature at 400 m versus the eddy depth of the 12°C isotherm for a given day, yearday 452. On any given day, there is a 1:1 correspondence between the slopes of the $T$ at 400 m versus $Z_{12}$ and $T'$ at 400 m versus $Z'_{12}$, within some small uncertainty to the line fit, because the system has nearly parallel isotherms within the main thermocline (Hogg 1986). The intercepts (i.e., $b$) differ between the eddy and mean plots, but are constants for all time. Using the relation between $T'$ and $Z'_{12}$ in Eq. (13), it can be shown that Eq. (12) holds

$$
T' \frac{\partial Z'_{12}}{\partial t} = \sigma \left[ Z'_{12} \left( \frac{m}{2} Z'_{12} + b \right) \right] = 0.
$$

(14)

Figure 11d shows that time series of BCdiv' and PKC' are modestly correlated with a correlation coefficient of 0.43. However, the major events such as the CCR formation (yeardays 325–375), CCR–jet interaction (yeardays 480–550), and external eddy–jet interactions (yeardays 550–600) during the unstable regime are highly correlated. The correlation coefficient for the unstable regime (yeardays greater than 300) is 0.6. The time means of BCdiv' and PKC' are also not statistically different from each other with time means of $1.4 \pm 0.32 \times 10^{-3}$ and $1.2 \pm 0.33 \times 10^{-3}$ cm$^2$ s$^{-3}$, respectively. During the stable regime, BCdiv' and PKC' are correlated during the first 100 days (yeardays 150–250), but there are some PKC' events that are not related to BCdiv' from yeardays 250–325.
A useful tool is to plot the time-mean midthermocline PKC versus BCdiv as a scatterplot. For wave growth and decay, the unsteady EPE equation is written as

$$\frac{\partial \text{EPE}}{\partial t} = \text{BCdiv} - \text{PKC},$$

(15)

where EPE = $a_g \bar{T}^2/2$. Here, Eq. (15) has neglected triple correlations and the mean advection of EPE balances the rotational BC (Marshall and Shutts 1981). For wave growth ($\partial \text{EPT}/\partial t > 0$), BCdiv must exceed PKC and vice versa for wave decay. The stable (Fig. 13a) and unstable (Fig. 13b) regimes show that PKC and BCdiv are linearly related with slopes greater than one and less than one for the stable and unstable regimes, respectively. This is consistent with the concept that the cross-front divergent eddy heat fluxes are within the wedge of instability during the unstable regime, but not during the stable regime. This may explain why some events are uncorrelated during the stable regime (Fig. 11d yeardays 250–325).

The temporal changes in EKE' (Fig. 11e), where EKE' = $0.5(u'^2 + v'^2)$ is the EKE before time averaging, are also correlated with events when BCdiv' and PKC' are correlated; such as the three events during the unstable regime in the previous paragraph (CCR formation, CCR–jet interaction, and external eddy–jet interaction). The major events during the KESS observations are

![Fig. 12. Scatterplots of (a) $T_{400}$ vs $Z_{12}$ and (b) $T'_{400}$ vs $Z'_{12}$ for yearday 452.](image)

![Fig. 13. Scatterplot of PKC vs BCdiv at 400-m depth between 34° and 37°N for (a) stable and (b) unstable regimes with a linear regression fit (solid black line). The dashed line is the 1:1 line.](image)
associated with a mean–eddy energy conversion, from MPE to EPE, and further conversion to EKE in accordance with baroclinic instability.

7. Discussion and summary

DEHFs in the Kuroshio Extension arise from varied mesoscale processes. The dominant features responsible for the largest observed variability during KESS were deep externally generated topographically controlled eddies and CCRs. Growth of upper meanders associated with deep externally generated upstream-propagating eddies in various frequency bands (5–60 days) (Greene et al. 2012; Tracey et al. 2011) arises in accordance with baroclinic instability; downgradient DEHFs were associated with the growth of meanders, which release MPE of the mean jet to the eddies. This process is different from the canonical view of baroclinic instability. The SYNOP experiment revealed observationally for the first time the importance of vertical coupling between deep eddies and the Gulf Stream variability (Watts et al. 1995; Howden 2000; Savidge and Bane 1999). Cronin and Watts (1996) observed cross-front eddy heat fluxes from trough events that were associated with vertical coupling between deep eddies and the Gulf Stream. These deep eddies spun up in place owing to lateral excursions of the Gulf Stream, which can easily generate deep vorticity between 0.1 and 0.2f.

As was discussed in Greene et al. (2009), due to a more shallow thermocline and thicker subthermocline layer than in the Gulf Stream, cyclogenesis by baroclinic stretching from lateral excursions of the Kuroshio Extension jet is weak and insufficient to explain variability observed during KESS. At the extrema of lateral excursions, baroclinic stretching can generate vorticity of 0.1–0.12f. Deep external eddies with relative vorticity in excess of 0.1f were observed to interact with the jet; driving strong cross-front DEHFs. When the phasing is right, this vertical coupling between deep externally generated eddies and the upper-ocean jet, with the deep leading the upper as in Phillips-like baroclinic instability, is associated with the growth of upper meanders. A case study of these 30–60-day deep eddies showed that they can shift the jet south; forming steep meander troughs that sometimes form in place and pinch off CCRs. Some of the strongest DEHFs were associated with this process. It begs the question whether the Kuroshio Extension between 143° and 149°E is capable of locally generating its own deep vorticity through internal processes or whether external features needed to form extreme trough events such as a ring formation. Otherwise, during the stable regime, when there was an absence of 30–60-day deep eddies traveling down the KESS array central line, cross-front DEHFs were weaker than during the unstable regime and no CCR formed. In fact, prior to this, a CCR did not form from July 2002 to December 2004 during the stable regime as observed from the ring census. However, whether there were deep external eddies during this time period is unknown and outside of the time interval of the available data.

The results from this study suggest there are distinct differences in the mean structure of DEHFs between stable and unstable meandering regimes, and between the features that drive that structure. During the stable regime, DEHFs were weak and mostly downgradient. The dominant cross-stream DEHFs arose from 10–15-day upper meanders, which grew upstream near 145°N. The suppression of CCR formation during stable regimes suggests that processes observed during the first six months of the KESS observations during the stable regime are characteristic of the previous two years.

The unstable regime was characterized by a red shift in the spectral content, the presence of CCRs, and asymmetry in downgradient DEHFs along the mean path. CCRs are the features responsible for the asymmetry in downgradient DEHFs. Cross-front DEHFs were strong and downgradient upstream of a mean trough in the along streamflow. Vertical coupling between deep eddies and the upper jet, combined with two CCR formations produced the majority of the variance in the unstable regime.

Eddy energetics analysis points to the DEHFs as playing a role in eddy–mean flow interactions through energy conversion. Time series of BCdiv (proportional to cross-front DEHFs) were correlated with vertical heat fluxes (PKC) during strong events, suggesting on average that eddies are releasing available potential energy from the mean baroclinic jet to drive the kinetic energy of the eddies. This gives a firm ground on which to interpret the DEHFs in terms of dynamics and variability in EKE.

As with any observational study in the ocean, many more years of data are needed to accurately access long-term trends and convergence of statistics. The cross spectra of the meridional eddy heat fluxes for the 16 months show that there is still a lot of unresolved energy at long periods. For convergence of statistics, a time series long enough such that the energy at the longest periods vanishes would be needed. However, this study showed that there are varied mesoscale processes that drive DEHF and that there are distinct differences between stable and unstable regimes.

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