Decadal Shifts of the Kuroshio Extension Jet: Application of Thin-Jet Theory*

YOSHI N. SASAKI†
International Pacific Research Center, University of Hawaii at Manoa, Honolulu, Hawaii

NIKLAS SCHNEIDER
International Pacific Research Center, and Department of Oceanography, University of Hawaii at Manoa, Honolulu, Hawaii

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ABSTRACT

Meridional shifts of the Kuroshio Extension (KE) jet on decadal time scales are examined using a 1960–2004 hindcast simulation of an eddy-resolving ocean general circulation model for the Earth Simulator (OFES). The leading mode of the simulated KE represents the meridional shifts of the jet on decadal time scales with the largest southward shift in the early 1980s associated with the climate regime shift in 1976/77, a result confirmed with subsurface temperature observations. The meridional shifts originate east of the date line and propagate westward along the mean jet axis, a trajectory inconsistent with the traditionally used linear long Rossby waves linearized in Cartesian coordinates, although the phase speed is comparable to that in the traditional framework. The zonal scale of these westward propagation signals is about 4000 km and much larger than their meridional scale. To understand the mechanism for the westward propagation of the KE jet shifts, the authors consider the limit of a thin jet. This dynamic framework describes the temporal evolution of the location of a sharp potential vorticity front under the assumption that variations along the jet are small compared to variations normal to the jet in natural coordinates and is well suited to the strong jet and potential vorticity gradients of the KE. For scaling appropriate to the decadal adjustments in the KE, the thin-jet model successfully reproduces the westward propagations and decadal shifts of the jet latitude simulated in OFES. These results give a physical basis for the prediction of decadal variability in the KE.

1. Introduction

After leaving the coast of Japan around 35°N, the Kuroshio penetrates as a free jet into the North Pacific as the Kuroshio Extension (KE) and retains its sharp jet structure with associated fronts to the date line and beyond. The KE jet reaches its maximum speed on the order of 1 m s⁻¹ at the surface and is accompanied by meridional potential vorticity (PV) gradients at the jet axis that are much larger than the planetary vorticity gradient (Fig. 1).

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† Current affiliation: Graduate School of Science, Hokkaido University, Sapporo, Japan.

Corresponding author address: Yoshi N. Sasaki, Science 8th bldg, 8-3-20, Graduate School of Science, Hokkaido University, N10, W8, Sapporo 060-0810, Japan. E-mail: sasakiyo@mail.sci.hokudai.ac.jp

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The KE undergoes significant changes on decadal time scales and has attracted much attention because of its hypothesized impact of the North Pacific and North America climate (e.g., Latif and Barnett 1996; Peng and Whitaker 1999; Xie et al. 2000; Schneider et al. 2002; Liu and Wu 2004; Kwon and Deser 2007; Qiu et al. 2007; Bond and Cronin 2008; Kelly et al. 2010). One of the largest decadal changes in the KE region occurred in the early 1980s associated with the 1976/77 climate regime shift (Nitta and Yamada 1989; Trenberth 1990). In the KE region, ocean waters cooled from surface to 400-m depth (Deser et al. 1996, 1999) and subsurface water freshened (Joyce and Dunworth-Baker 2003), a change that will show to be confined to the KE jet axis in observations and an eddy-resolving ocean model hindcast. This change indicates the southward shift of the KE jet.

Observational (e.g., Miller et al. 1998; Deser et al. 1999; Lysne and Deser 2002; Qiu and Chen 2005) and numerical studies (e.g., Seager et al. 2001; Taguchi et al. 2007) suggest that decadal variability in the KE is primary
potential vorticity is defined by reference Coriolis parameter, circles) at 150°E August 2005. The upper-layer quasigeostrophic potential vorticity is defined by βy − f0(H − h)/H, where f0 is the reference Coriolis parameter, β is the meridional gradient of the Coriolis parameter, y is the meridional coordinate, h is the layer thickness, and H is the reference layer thickness. The bottom of the upper layer is defined as the 26.5σθ surface. The density field is calculated from gridded temperature and salinity fields based on Argo data by Roemmich and Gilson (2009).

forced by basin-wide wind stress curl changes with the KE lagging by several years. Lysne and Deser (2002) showed that decadal subsurface temperature variations in the KE region are related to the 4-yr leading wind stress curl fluctuations that have large amplitudes around east of the date line. These results suggest the importance of westward propagation signals to the KE region, but the mechanisms for the westward propagation and for the meridional shifts of the KE jet are not fully understood. A better understanding of these processes is required to interpret and predict decadal variability in the North Pacific.

There are a number of approaches to reconcile how propagation signals from east influence the KE jet latitude. A widely used mechanism to account for westward propagation signals, such as sea level anomalies (SLAs), is long Rossby waves derived from the PV equation for a 1.5-layer ocean linearized in Eulerian coordinates (e.g., Sturges and Hong 1995; Schneider and Miller 2001; Qiu 2003; Qiu and Chen 2005). For example, SLAs associated with the linear long Rossby waves in this framework along the latitude of the jet can change the jet latitude, because the waves propagate westward along a constant latitude. Nevertheless, the validity of this model, termed Eulerian in the following, is questionable in the KE region. Qiu and Chen (2010) noted that the Eulerian model does not reproduce satellite observed SLAs around the path of the upstream KE. In addition, the Eulerian framework fails to describe the meridional shifts of the KE jet of eddy-resolving model hindcasts (Taguchi et al. 2007). For the linearization in the Eulerian framework, the assumption is that perturbations of velocity and sea level are much smaller than the background velocity and sea level gradients, respectively. However, if the strong jet shifts meridionally, the associated perturbations are comparable in magnitude to the background flow and gradient in Cartesian coordinates (as will be shown in section 2). To overcome this problem, we employ a thin-jet framework.

The purpose of this study is to examine the process of the westward propagation and the mechanism for the shifts of the KE jet on decadal time scales using the framework of a thin jet that goes back to studies of Robinson et al. (1975) and Flierl and Robinson (1984). The thin-jet theory has been used to investigate meanders of the Gulf Stream (Cushman-Roisin et al. 1993, hereafter CPR93; Lee and Cornillon 1996). The thin-jet limit considers the temporal evolution of a PV front in the natural coordinates and is arguably a better dynamic framework to understand the shift of the sharp front and swift current in KE than a linearization in Cartesian coordinates employed hitherto. We seek to test this approach on decadal adjustment of the KE jet as simulated in a high-resolution ocean general circulation model (OGCM) hindcast. In section 2, we review the model simulation and observations used for validation and describe decadal variability of the KE in the OGCM for the Earth Simulator (OFES) and its comparison with observations. The propagation tendency of the KE jet is examined in section 3. In section 4 we adapt the thin-jet model to scaling appropriate to the decadal adjustments in the KE and then show in section 5 that the thin-jet model successfully captures the features of the KE jet variations. A summary and discussion are presented in section 6.

2. Model and data

a. Ocean model hindcast

The eddy-resolving OFES (Masumoto et al. 2004; Sasaki et al. 2004, 2008a) is based on the National Oceanic and Atmospheric Administration/Geophysical Fluid Dynamics Laboratory (NOAA/GFDL) Modular Ocean Model version 3 (MOM3) (Pacanowski and Griffies 1999). The model domain is global in longitude and extends from 75°S to 75°N, with resolutions of 0.1° in latitude and longitude and 54 levels in the vertical, and it includes realistic bottom topography. The model is forced with surface wind stress, heat and freshwater fluxes based on the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996). The surface heat flux is calculated with the bulk formula (Taguchi et al. 2007). For the linearization in the Eulerian framework, the assumption is that perturbations of velocity and sea level are much smaller than the background velocity and sea level gradients, respectively. However, if the strong jet shifts meridionally, the associated perturbations are comparable in magnitude to the background flow and gradient in Cartesian coordinates (as will be shown in section 2). To overcome this problem, we employ a thin-jet framework.

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of Rosati and Miyakoda (1988) from model surface temperatures and atmospheric variables of the reanalysis data. Sea surface salinity is forced by an equivalent freshwater flux and is also restored to its monthly climatology of *World Ocean Atlas 1998* (WOA98; Boyer et al. 1998a,b,c) with a time scale of 6 days. Within 3° from the near northern and southern artificial boundaries (72°–75°N, 72°–75°S), temperature and salinity fields at all depths are restored to the climatological monthly mean of WOA98.

From a state of rest and climatological temperature and salinity fields of WOA98, the model was spun up for 50 yr with monthly climatological atmospheric forcing. Following the spinup, OFES is forced with daily-mean atmospheric fields from 1950 to 2004; we exclude the adjustment of the upper ocean from the initial values of the climatological run during the first 10 yr and analyze the hindcast output from 1960 to 2004. Because we focus on interannual to decadal variability, a 9-month running mean filter is applied to all model output (unless noted otherwise).

This OFES hindcast integration has been used in a number of studies that include comparisons with available observations (e.g., Nonaka et al. 2006; Taguchi et al. 2007). The climatology of the simulated sea surface height shows the correct separation of the Kuroshio from the Japan coast around 35°N, the narrow frontal structure of the KE jet, and the recirculation gyre south of the KE jet (Fig. 2a). OFES also reproduces the steady meander of the KE jet in the upstream region between 142° and 155°E (Mizuno and White 1983). The KE jet spreads out east of 155°E in climatology, but the narrow jet structure with a width of about 100 km extends to the date line in monthly-mean fields (Fig. 2b).

### b. Upper-ocean temperature observations

To compare the OFES hindcast of the KE region with observations, we employ monthly-mean subsurface temperature data on a 1° × 1° grid over a depth range of the surface to 700-m depth from 1960 to 2004 (Ishii et al. 2003, 2006). This dataset is based on objective analysis of in situ observations using the latest observational databases. The horizontal decorrelation scale is 300 km at the surface and vertically increases at a rate of 30 km per 100 m (Ishii et al. 2006). Following past studies (e.g., Deser et al. 1999), we use temperature data at 400-m depth to represent upper-ocean variability. Similar to the model output, observations are smoothed with a 9-month running mean filter.

### c. Decadal variations in the Kuroshio Extension

Before considering the detailed behavior of the KE jet axis, dominant variability of SLAs of OFES in the upstream KE region is identified by an empirical orthogonal function (EOF) analysis on monthly output (gray dashed box in Fig. 3a) and is compared with subsurface observations. The leading EOF mode explains 21.3% of the SLA variance. The regression of SLAs (Fig. 3a) onto the corresponding principal component (PC1) shows narrow and zonally elongated positive SLAs concentrated along the KE jet from 140° to 175°E with maximum SLAs of greater than 25 cm and a spatial structure that mimics the steady meander pattern in the upstream region (Fig. 2a). These concentrated SLAs are comparable in magnitude to the climatological sea level gradients there, suggesting that this mode is related to the meridional shift of the jet. Further poleward, the loading decays rapidly but remains positive, whereas, south of 34°N, the loading changes sign and is weakly negative.

PC1 is dominated by decadal variations (solid line in Fig. 3b) and was generally positive from 1960 until the late 1970s, except in the mid-1960s, and rapidly shifted to negative values in the early 1980s. After this sudden change, PC1 generally stayed negative (the mid-1980s and late 1990s) or around 0 (the early 1990s and early 2000s) until the end of the data in 2004. This temporal evolution of PC1 is highly correlated ($r = 0.94$) to variations of the jet latitude (dashed line in Fig. 3b), defined

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**FIG. 2.** (a) Climatology and (b) monthly mean in August 1989 of sea surface height in OFES. The contour interval is 10 cm.
as maximum zonal velocity positions at 100-m depth averaged between 142°–155°E from monthly outputs. The latitude of the KE jet axis differs by about 1.5° between the positive and negative phases of PC1.

The corresponding vertical KE jet structures are similar in the positive and negative phases in spite of the latitude change (Fig. 4). In either phase, the zonal velocities reach maximum values of 70 cm s⁻¹ at surface and tilt southward with depth, but the difference between both phases around the jet axis is less than 5 cm s⁻¹. Hence, the leading mode of the simulated SLAs in the upstream KE region captures a meridional shift of the jet that conserves its meridional and vertical structures.

A similar spatial pattern and temporal structure are found in observations of upper-ocean temperature. The correlation of observed temperature at 400-m depth with PC1 (Fig. 5a) has highest positive values in the upstream KE region and has a narrow meridional and broad zonal structure, similar to the simulated SLA pattern (Fig. 3a). Because a northward shift of the KE jet is associated with subsurface warming, the positive correlations in the upstream KE region are consistent with the positive SLAs in OFES. Consistently, the correlations of simulated temperature at 400-m depth onto PC1 exhibit narrow positive values along the KE jet (not shown). The observed 400-m temperature time series averaged over the upstream KE region (solid line in Fig. 5b) are well correlated with PC1 ($r = 0.74$) and show rapid cooling in the early 1980s, consistent with the southward shift of the KE jet in OFES. Hence, we conclude that OFES successfully reproduces the observed decadal fluctuations of the KE jet.

In closing this subsection, we examine the skill of the Eulerian model, based on the linearized vorticity equation in a 1.5-layer reduced gravity ocean in the long-wave limit (e.g., Schneider and Miller 2001; Qiu 2003; Sasaki et al. 2008b). The equation of free waves is

$$\frac{\partial}{\partial t} \eta' - c_R \frac{\partial}{\partial x} \eta' = 0,$$

where $\eta'$ is SLA and $c_R$ is the phase speed of long baroclinic Rossby waves in the Eulerian framework. The free
waves propagate westward along a constant latitude that is line of constant background planetary vorticity. The SLAs in the KE region induced by free wave propagation from the east are determined by forcing the Eulerian model with the monthly SLAs in OFES at the date line (180°), in a procedure that we will repeat for the thin-jet model in section 5. The latitude-dependent phase speed $c_R$ is the same as used by Qiu (2003).

The regression of SLAs of the Eulerian model onto PC1 shows positive and negative values north and south of 34°N, respectively (Fig. 6a), roughly consistent with the SLAs in OFES (Fig. 3a). However, these positive SLAs have a meridional scale larger than the SLAs in OFES and observations of temperature (Fig. 5a), and the SLAs predicted by the Eulerian model around the KE jet are much smaller than those in OFES, consistent with results of Taguchi et al. (2007). These small SLAs north of the KE jet in the Eulerian model are much less effective at changing the KE jet latitude.

Along the latitude of the jet axis, westward propagation signals are weak in OFES (Fig. 7a). The lag correlations between PC1 and SLAs in OFES averaged over 34°–37°N are small east of 160°E, and the westward propagation tendency is not seen. This result confirms that westward propagation waves along a constant latitude cannot account for the shifts of the KE jet. In addition, the vertical structure of the jet changes in response to forcings in the Eulerian framework, because the SLAs are assumed to be smaller than the strong sea level gradients of the background jet. This is not the case in OFES (Fig. 4). Hence, another dynamic framework is necessary, and we will explore the thin-jet model in the next section.

3. Variations in the Kuroshio Extension jet axis

To apply the thin-jet theory of CPR93, we cast decadal fluctuations in the KE in terms of the jet location and determine at each longitude from 140°E to 170°W the KE jet latitude as the position of maximum monthly

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**Fig. 5.** (a) Correlation coefficients of observed monthly temperature at 400-m depth onto PC1. The gray dashed box indicates the region where the area-averaged temperature anomalies are calculated. (b) Observed temperature anomalies at 400-m depth averaged over 34°–37°N, 140°–155°E (solid, right axis) and PC1 (dashed, left axis). Before calculating the correlations and regional average, a 9-month running mean filter is applied to the temperature data.

**Fig. 6.** Regression coefficients of monthly SLAs calculated by (a) the Eulerian model [Eq. (1)] and (b) the thin-jet model [Eq. (27)] onto PC1. The SLA data are smoothed by a 9-month running mean filter. Contour intervals are (a) 1 and (b) 4 cm, and shading indicates the regions where the absolute correlations are >0.4. The gray dashed box in (b) denotes the domain of the reconstruction of the SLAs by the thin-jet model.
absolute velocity of those grid points between 30° and 40°N that have a positive zonal velocity from OFES output. The longitude range encompasses the simulated KE jet locations that extend from the coast of Japan to the date line (Fig. 2b). Because the vertical structure of the jet is equivalent barotropic, as assumed in CPR93, we determine the upper-layer interface from the correlation coefficients between SLAs and 7 isopycnal depths at 0.5σθ intervals from 24.5σθ to 27.5σθ at each grid point. The correlation between SLAs and the isopycnal depth of 26.5σθ is the highest in the KE region, and we use the velocity averaged between the surface and 26.5σθ. We note that our results are insensitive to the specific choice of the lower interface. Furthermore, jet positions are noisy because of eddies, meanders, and bifurcations. However, results below are robust if instead the KE jet position is defined by a fixed PV contour.

Meridional displacements of the KE jet in OFES (Fig. 8), smoothed to annual means to emphasize low-frequency signals, propagate westward, and influence the fluctuations of the upstream KE jet, especially the southward shift in the early 1980s. Meridional displacements first appear in the downstream KE region around 180°–170°W and then propagate westward and reach to the upstream KE region after about 5 yr. The prominent northward and southward shifts appear in 1960–75 and 1977–94, respectively, although the southward shifts of 1983–86 disappear around 165°E. Amplitudes of the jet shift become progressively smaller from the downstream (about 1.5°) to the upstream (about 0.8°), a point we will discuss in section 6.

The lag correlations between PC1 and the meridional jet displacements quantify the westward propagation (Fig. 7b). At 0-yr lag, the high correlations around 142°–155°E are consistent with PC1 representing the meridional shift of the upstream KE jet. The lag correlation is dominated by a tendency for westward propagation from the downstream region, particularly west of the date line, to the upstream region. This clear westward propagation suggests that the forcing east of the date line plays the dominant role in the shifts of the upstream KE jet. The typical transit time from 180° to 150°E is about 5 yr, corresponding to a phase speed of about 1.7 cm s⁻¹, comparable to that used in the Eulerian models (e.g., 2.8 cm s⁻¹ at 36°N; Qiu 2003).

The associated lag composites of SLAs show asymmetries consistent with a meridional displacement of the jet axis and determine the scales needed for the derivation of the thin-jet model. Composites are based of years when the magnitude of annual-mean PC1 exceeds one standard deviation with statistical significance assessed using a t test.

When the KE jet shifts northward, large and statistically significant positive SLAs are located at and north of the mean jet axis that has a slight but significant...
meridional tilt (Fig. 9). Five years prior, the positive SLAs are located in the upstream region east of 165°E and northward of the climatological jet position (Fig. 9a), consistent with northward shift of the KE jet there. These SLAs propagate westward along the meridionally tilted path, are located from 140°E to 170°E 3 yr later (Fig. 9d), and when reaching the upstream region are located over the mean jet axis (Fig. 9f). At this time, the SLAs in the upstream region intensify likely because of non-linear dynamics such as PV advection (Hogg et al. 2005; Taguchi et al. 2005). The zonal scale of these SLAs is about 4000 km (Fig. 9a), which is much larger than their meridional scale.

The southward shifts of the KE jet again propagate westward, but the SLA anomalies are negative and located to the south of the mean jet position (Fig. 10). Although the negative SLAs are not clear at 5- and 4-yr lags probably because the large southward shift around the late 1980s in the downstream region propagated to only 165°E (Fig. 8), at a 3-yr lag weak but statistically significant negative SLAs are located east of 160°E and southward of the climatological KE jet axis (Fig. 10c). At a 2-yr lag, these negative SLAs exhibit narrow-banded structures from 150° to 170°E and east of 175°E (Fig. 10d) and propagate westward until reaching the upstream region (Figs. 10e,f).

4. Derivation of the thin-jet equation

The results of the previous section indicate that the meridional shifts of the KE jet propagate westward along the mean jet axis with a phase speed of the same order as that of the linear long Rossby waves. To clarify the mechanism for this westward propagation, we modify the thin-jet equation of CPR93. Here, we only present the main points of the derivation, because it closely mirrors CPR93, except for the scaling estimates and the assumption that the jet is nearly zonal, suitable for describing temporal evolution of the KE jet on decadal time scale. The reader is referred to CPR93 for more complete presentation of the expansion. As in the Eulerian model, because we are interested in the propagation of free waves, the extension of the theory to consider effects of wind forcing is left for a future study.

The model is equivalent barotropic with primitive equations on the beta plane, for prognostic variables of layer thickness and layer-averaged zonal and meridional
velocity. Hence, the model excludes baroclinic instability, which may render perturbations unstable (Flierl 1999). Following CPR93, the equations are nondimensionalized by a reference depth $H$, a deformation radius $(g'H)^{1/2}/f_0$, a gravity wave speed $(g'H)^{1/2}$, and a time scale $1/f_0$, where $g'$ is the reduced gravity and $f_0$ is the reference Coriolis parameter. Then, the equations are converted from the Cartesian coordinates to the natural coordinates $(s, n)$ and velocity components $(U, V)$ downstream and perpendicular to the left of the time-dependent jet center line $[X(s, t), Y(s, t)]$ (Fig. 11). The resultant primitive equations are

$$
\partial_t (JU - A + nJ\partial_y C)\partial_s + (V - C)\partial_n U \\
- [1 + J\partial_y C + \beta(Y + n\partial_s X) + JKU]V = -J\partial_s h,
$$

(2)

$$
\partial_t (JU - A + nJ\partial_y C)\partial_s + (V - C)\partial_n V \\
+ [1 + J\partial_y C + \beta(Y + n\partial_s X) + JKU]U = -\partial_n h,
$$

(3)

and

$$
\partial_t (JU - A + nJ\partial_y C)\partial_s + (V - C)\partial_n h \\
+ (J\partial_s U + \partial_n V - JKV)h = 0,
$$

(4)

where the quantities $K(s, t)$, $A(s, t)$, $C(s, t)$, and $J(s, n, t)$ are defined as

$$
K = \partial_s X \cdot \partial_{ss} Y - \partial_{ss} X \cdot \partial_s Y,
$$

(5)

$$
A = \partial_s X \cdot \partial_s X + \partial_s Y \cdot \partial_s Y,
$$

(6)

$$
C = \partial_s Y \cdot \partial_s Y - \partial_s X \cdot \partial_s X,
$$

(7)

and

$$
J = 1/(1 - nK).
$$

(8)
The term $K$ is the curvature of the jet, and $A$ and $C$ are the along-jet and across-jet velocity of the jet center line $(X, Y)$, respectively. The terms $X$ and $Y$ obey the relation

$$\left( \frac{\partial}{\partial x} X \right)^2 + \left( \frac{\partial}{\partial y} Y \right)^2 = 1. \tag{9}$$

The nondimensional constant $\beta$,

$$\beta = \frac{\left( g' H \right)^{1/2} df}{\int_0^s dy},$$

compares the Rossby radius to the meridional scale of an order one change of planetary vorticity. For the KE, the Rossby deformation radius is on the order of 40 km (for $H$ and a typical $g'$ used in the Eulerian model), and $\beta$ is on the order of $10^{-2}$ and small.

Following CPR93, the thin-jet approximation is introduced by scaling these nondimensional equations and expanding the prognostic variables in a small parameter $\varepsilon$ that describes the sharpness of the front, where $\varepsilon$ is a small number of order $\beta^{1/2}$. CPR93 applied the thin-jet framework to investigate meanders of the jet whose wavelength and period are less than about 1000 km and 1 yr, respectively. We modify this scaling to be suitable for long-wavelength (about 8000 km) fluctuation on decadal time scales (Figs. 2, 9, 10). Furthermore, unlike CPR93, we use different scaling for the time-mean and fluctuation parts of the jet axis locations, $X = \bar{X}(s) + X'(s, t)$ and $Y = \bar{Y}(s) + Y'(s, t)$, denoted by an overbar and prime, respectively. Our scaling is as follows:

- along-jet distance and time-mean zonal meander scale: $s - \bar{X} - \varepsilon^{-2}$;
- time-mean meridional meander scale: $\bar{Y} \sim \varepsilon^{-1}$;
- zonal and meridional meander fluctuations: $X' \sim Y' \sim \varepsilon^{-1}$;
- across-jet distance: $n \sim 1$;
- upper-layer depth variation: $h \sim 1$;
- along-jet velocity: $U \sim 1$;
- across-jet velocity: $V \sim \varepsilon^2$; and
- time scale: $t \sim \varepsilon^{-4}$,

so that the jet width $n$ is of the same order of the Rossby radius and the along-jet scale $s$ is two orders of magnitude larger than the Rossby radius. Because the jet is nearly zonal, the zonal scale of the time-mean jet path is much larger than its meridional scale.

Using these scalings, Eq. (9) becomes

$$\frac{\partial}{\partial s} \bar{X} \approx 1. \tag{10}$$

In addition, all prognostic variables in Eqs. (2)–(4) are expanded in terms of $\varepsilon$,

$$U = U_0 + \varepsilon U_1 + O(\varepsilon^2).$$

Then, the resultant leading order equations are

$$U_0 \frac{\partial}{\partial s} U_0 + V_0 \frac{\partial}{\partial n} U_0 - V_0 = -\frac{\partial}{\partial n} h_0, \tag{11}$$

$$U_0 = -\frac{\partial}{\partial n} h_0, \quad \text{and} \tag{12}$$

$$\frac{\partial}{\partial s} (h_0 U_0) + \frac{\partial}{\partial n} (h_0 V_0) = 0. \tag{13}$$

Together with boundary conditions far from the jet axis that across-jet velocity and variations of upper-layer thickness in the $s$ direction are zero, Eqs. (11)–(13) imply that $h_0$ is independent of $s$ and $V_0 = 0$ (CPR93). Thus, the zeroth order describes a geostrophic, uniform jet. To obtain the location of the jet axis and its temporal evolution, the next order equations have to be considered. This mirrors the derivation of the Eulerian model, where the leading order specifies the geostrophic balance of the flow, and the next order describes its evolution.

The next order equations are

$$U_0 \frac{\partial}{\partial s} U_1 + (V_1 - C_0) \frac{\partial}{\partial n} U_0 - V_1 = -\frac{\partial}{\partial s} h_1, \tag{14}$$

$$U_1 + YU_0 = -\frac{\partial}{\partial n} h_1, \quad \text{and} \tag{15}$$

$$U_0 \frac{\partial}{\partial s} h_1 + (V_1 - C_0) \frac{\partial}{\partial n} h_0 + h_0 (\frac{\partial}{\partial s} U_1 + \frac{\partial}{\partial n} V_1) = 0. \tag{16}$$

The PV balance of these equations is shown in appendix. Here, $C_0$, the leading-order across-jet velocity of the jet center line, is given from Eqs. (7) and (10) by

$$C_0 = \frac{\partial}{\partial s} \bar{X} \cdot \frac{\partial}{\partial n} Y' \approx \frac{\partial}{\partial n} Y'. \tag{17}$$

Hence, the velocity of the jet normal to itself is dominated by its meridional component. Equations (14)–(16) represent the linearized equations in the natural coordinates and are similar to those derived by CPR93, but for the absence of the curvature vorticity term in Eq. (15) that is small for the long-wavelength scale of the KE. Note that the across-jet movement of the zeroth-order thickness gradient ($-C_0 \frac{\partial}{\partial n} h_0$) is more efficient in inducing thickness changes than the local stretching ($\frac{\partial}{\partial n} h_0$) in the mass conservation equation [Eq. (16)], because the jet is so thin that the across-jet distance is narrower than the scale of the shift causing large changes of $h$. Therefore, the zeroth-order thickness structure is frozen in time and only depends on $n$, as found in Fig. 4.

It is convenient to separate $(U_1, V_1)$ into a geostrophic part $(U_g, V_g)$ and an ageostrophic part $(U_a, V_a)$, where the geostrophic part is defined as
Thus, the geostrophic flow is nondivergent. Subtracting Eqs. (18) and (19) from Eqs. (14) and (15), respectively, we obtain the ageostrophic part as

\[ V_a = U_0 \partial_n U_1 + \left( V_1 - C_0 \right) \partial_n U_0 \quad \text{and} \quad U_a = -Y U_0. \]  

(20)

Equation (20) indicates that the across-jet ageostrophic flow is caused by nonlinear advection. The along-jet ageostrophic flow adjusts the geostrophic speed when the jet latitude changes, but nonlinear advection effect does not play a role in Eq. (21). This is because the along-jet velocity is much larger than the across-jet velocity, and thereby the advection terms that are related to the across-jet velocity can be ignored. Substituting Eqs. (18)–(21) into Eq. (15) yields the mass balance between the across-jet thickness advection and the divergence of ageostrophic flow,

\[ (V_a - C_0) \partial_n h_0 + h_0 (\partial_n U_a + \partial_n V_a) = 0. \]  

(22)

This equation is the thin-jet counterpart to Eq. (1).

Equation (22) together with Eqs. (18)–(21) determines the first-order velocity and thickness fields, but our interest is the temporal evolution of the jet position \( Y \). Following CPR93, the position of the jet is obtained by integrating Eq. (22) across the jet from large negative \( n \) to large positive \( n \),

\[ -C_0 \int \partial_n h_0 \, dn + \int h_0 \partial_n U_a \, dn = 0, \]  

(23)

where we ignore the term of the across-jet ageostrophic mass flux far from the jet (CPR93). This means that the across-jet ageostrophic mass flux [the sum of the first and fourth terms in Eq. (22)] does not contribute the shift of the jet axis and implies that the shifts of the jet axis are independent of the changes of across-jet thickness structure that accelerate the jet. Equation (23) indicates that the divergence of the beta-induced along-jet ageostrophic flow induces the across-jet thickness advection by the temporal shift of jet latitude.

Eliminating \( U_a \) from Eq. (23) via Eq. (21) and by the use of Eqs. (12) and (17) and \( \partial_n h_0 = 0 \) yields

\[ \partial_n Y' = b \partial_n Y = b \partial_n Y + b \partial_n Y', \]  

(24)

where the parameter \( b \) is given by

\[ b = \int h_0 \frac{dh_0}{dn} \, dn / \int \frac{dh_0}{dn} \, dn, \]  

(25)

or in dimensional form

\[ b = \beta g' (h_S + h_N) / 2 f_0^2, \]  

(26)

where \( h_S \) and \( h_N \) are the upper-layer thickness in the southern and northern sides of the jet, respectively. Equation (24) in steady state becomes

\[ 0 = b \partial_s Y', \]  

(27)

and indicate that meridional displacements of the jet propagate toward the west along the jet axis. The phase speed \( b \) is determined by the upper-layer thicknesses in the southern and northern sides of the jet and equals the average of the phase speeds of linear long Rossby waves to the north and south of the jet [Eq. (26)]. This situation is similar to interfacial waves along a velocity discontinuity that propagates with a phase speed of equal to the average of the velocity on the two sides of the discontinuity (Kundu 1990). The westward propagation of the jet shift in the thin-jet equation is consistent with the aforementioned westward propagation of the KE jet shifts along the mean jet axis (Figs. 7–10). This will be further confirmed in the next section.

Equation (27) is a remarkable result, because the temporal evolution of the jet axis only depends on latitude changes of the path. The jet structure, however, influences the solution of the first-order thickness and velocity fields from Eqs. (18)–(21) and (22). Although the thin-jet model has a similar form to the Eulerian model [Eq. (1)] in that variations in the Coriolis parameter play an essential role in the propagation in both models, an important difference between the thin-jet and the Eulerian models is the propagation direction of signals. SLAs in the Eulerian model propagate along a constant latitude, but in the thin-jet model the shift of the jet axis propagates along the mean jet axis as Rossby waves. In both cases, the direction is aligned with the isopleths of mean PV fields. In the Eulerian case, the mean PV contours are determined by \( f \) and are zonal, but those in the thin-jet case are oriented along the mean jet axis.
5. Application of the thin-jet equation

Let us examine if free waves of the thin-jet model reproduce the westward propagation of the KE jet shift. The thin-jet equation [Eq. (27)] is solved numerically with horizontal resolution 0.5° to calculate the signal propagation from 142°E to 180° where the eastern boundary is determined by the drop of the lag correlations (Fig. 7b). Starting from zero initial condition, the model is forced at the eastern boundary by monthly anomalies of jet latitude \( Y \) in OFES. This model setting resembles the setting of the Eulerian model (section 2) and idealized regional OGCM experiments to examine KE jet changes associated with the 1976/77 climate regime shift that were forced at eastern boundary of the domain (170°E; Taguchi et al. 2005). The value of \( b \) is calculated from OFES output with upper-layer thicknesses \( h_N \) and \( h_S \) that are chosen as the depth of the 26.5σθ isopycnal 1.0° north and south of the jet, respectively. Values of \( b \) are insensitive to changes of the integration interval from 0.5° to 2.0° from the jet. The reduced gravity is calculated by separately averaging the density of the water above and below the 26.5σθ isopycnal. The phase speeds vary from 1.9 to 2.4 cm s\(^{-1}\) and are comparable but somewhat faster than the phase speed in OFES estimated from Fig. 7b.

The propagation pattern and amplitude of the jet shift predicted by the thin-jet model (Fig. 12) are overall similar to that of OFES (Fig. 8) with a correlation of spatial patterns of OFES and the thin-jet model of 0.62. The meridional shifts in the downstream KE region propagate westward and reach to the upstream KE region after about 5 yr, consistent with the signals in OFES.

In the upstream region, the thin-jet model successfully reproduces the sudden southward shift in the early 1980s, whereas the timing of the shift is somewhat earlier than that in OFES (Fig. 13). The thin-jet model also reproduces the northward position before the early 1980s and southward position after the early 1980s, although the amplitude of the thin-jet model is about 1.5 times larger than that of OFES. The correlation of the latitude anomalies between OFES and the thin-jet model is 0.61 and increases to 0.68 if the correlation is calculated from the period 1965–2004 to account for the zero initial condition that limits the effect of the eastern boundary forcing in upstream region for the first few years.

Sea level variations in the KE region are reconstructed using the numerical results of the thin-jet model. Based on the thin-jet theory, the meridional structure of the KE jet is fixed and only its latitude varies temporally. The reconstructed monthly sea surface height \( \eta \) is given by

\[
\eta(x, y, t) = \eta_L[x, y_0(x) + dy(x, t)],
\]

where \( \eta_L \) is the climatology of the sea surface height in OFES in Lagrangian coordinates, \( y_0 \) is the climatological latitude of the KE jet axis, and \( dy \) is its monthly anomaly. The \( \eta_L \) is calculated at each longitude as the composite of monthly sea surface height of OFES, relative to the axis of the KE jet, which is determined by the maximum absolute velocity of the upper layer; \( dy \) is estimated using the numerical results of the thin-jet equation (see Fig. 12).

SLAs reconstructed by the thin-jet model are in good agreement with those of OFES (Figs. 3a, 6b). The large positive SLAs are concentrated along the KE jet with their narrow meridional extent from 34° to 37°N with maximum values of about 25 cm, and the weak negative SLAs are located south of the 34°N (Fig. 6b), consistent with the SLAs in OFES (Fig. 3a). Hence, in contrast to
the Eulerian model, the thin-jet model for the displacement of a PV front successfully reproduces the westward propagation of the KE jet shifts and the leading mode of sea level variability in the KE region in OFES.

6. Summary and discussion

Decadal variability of the KE from 1960 to 2004 is examined using output from a hindcast with the OFES model and explored in the limit of a thin jet. OFES successfully captures the meridional shift of the KE jet on decadal time scales. The first EOF of the simulated sea level variability in the KE region shows large positive SLAs along the jet (Fig. 3a) that are consistent with observed subsurface warming along the KE jet (Fig. 5a). The corresponding principal component exhibits prominent decadal variability, including a rapid change in the early 1980s (Fig. 3b), in accordance with the observed subsurface temperature fluctuations (Fig. 5b).

The structure of SLAs in OFES is consistent with a displacement of the near zonal KE jet. Meridional shifts of the jet in the downstream region have a zonal scale of about 4000 km, much larger than the meridional scale (Figs. 9, 10), and propagate from downstream region along the mean jet axis (Figs. 7–10) with a phase speed of about 1.7 cm s$^{-1}$. In addition, the locations of the SLAs are different between the northward and southward shift cases. The phase speed is comparable to that of the Eulerian model; however, the propagation direction is different from that model.

These characteristics of the westward propagation of the meridional shift of the KE jet are captured by the thin-jet framework, linearized in natural coordinates. The thin-jet equation in long-wavelength and low-frequency limit of the nearly zonal jet (section 4) shows that meridional displacements of the jet propagate toward the west along the mean jet axis with a phase speed given by the average of Rossby wave speeds to the north and south of the front [Eq. (27)]. The thin-jet model reproduces the westward propagation of the meridional displacement of the KE jet in OFES (Fig. 12) and the decadal fluctuations of the shift in the upstream KE jet, especially in the early 1980s (Fig. 13), albeit with an overestimated amplitude of the shifts of the upstream region. The reconstruction of sea level variations based on the thin-jet model reproduces the dominant decadal variability of OFES in the KE region and is far superior to the Eulerian model (Fig. 6). This is because the nonlinear profile of the sharp PV front of the jet is considered in the thin-jet model, whereas the Eulerian model only can describe linearized perturbations of a background intense jet. Thus, the thin-jet model is physically more appropriate than the Eulerian model. Our results indicate that (i) local wind forcing, which is not considered in the thin jet, is secondary for changes in the Kuroshio jet in the western North Pacific and (ii) the dynamics in the thin-jet framework accounts for the bulk of the KE jet fluctuations in OFES.

In this study, we described free waves of a jet that has a constant across-jet structure in along-jet direction. The weakening of the KE jet in OFES to the downstream region likely causes the increasing of amplitudes of the jet shift from the upstream to the downstream (Fig. 8). To include the progressive weakening of the jet into the thin-jet model will be the subject of a future study.

Our results give a physical basis for predictions of decadal variability of the KE jet and have substantial implications for climate and ocean ecosystem variability. Recent in situ and satellite observational studies show clear evidence of oceanic frontal scale variations in the western boundary current regions impacting the atmospheric boundary layer (Nonaka and Xie 2003; Tokinaga et al. 2006) and troposphere (Minobe et al. 2008). Thus, the adjustments of the KE jet investigated here may hold the key to longer-term predictions of climate anomalies in the Pacific/North American region.

In addition, meridional shifts of the KE jet accompany large sea temperature changes. Particularly in winter, sea surface temperature fluctuations in the KE and its recirculation region are important for the recruitment of Japanese sardine and chub mackerel (Noto and Yasuda 2003; Yatsu et al. 2005). In recent years, oceanic jets have been discovered in many regions and have attracted considerable attention (e.g., Nakano and Hasumi 2005; Maximenko et al. 2005). Recent progress in satellite observations and numerical modeling allows us to explore the mechanism of oceanic jet fluctuations. If these jets are aligned with strong PV fronts, such as in the western boundary current regions, our thin-jet equations may be a useful tool to the investigation of low-frequency variability of these oceanic jets and expand the dynamical toolbox from the traditionally used Eulerian model.

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Thus, PV changes associated with the thickness advec-
tions by along-jet velocity \( U_0 \) and the across-jet shift of the jet center \( C_0 \). The thickness advec-
tions by along-jet ageostrophic flow \( V_a \) and by across-jet geostrophic flow \( V_c \) cancel out each other because of the non-Doppler effect (e.g., Liu 1999). Next, combining the second term in Eq. (A1) and Eq. (21) gives

\[
U_0 \partial_s U_a = - \partial_s U_a. \tag{A3}
\]

Finally, we consider the last term in Eq. (A1) with Eqs. (20) and (21),

\[
- \left[ U_0 \partial_s U_1 + (V_1 - C_0) \partial_s U_0 \right] = - \partial_s \left[ U_0 \partial_s U_1 + (V_1 - C_0) \partial_s U_0 \right] + \partial_s U_0 \cdot \partial_s U_1 + \partial_s V_1 \cdot \partial_s U_0
\]

\[
= - \partial_s V_a + \partial_s U_0 (\partial_s U_1 + \partial_s V_1) = - \partial_s V_a + \partial_s U_0 (\partial_s U_a + \partial_s V_a)
\]

\[
= \partial_s U_0 \cdot \partial_s U_a - (1 - \partial_s U_0) \partial_s V_a. \tag{A4}
\]

The sum of Eqs. (A3) and (A4) shows the conservation of circulation as follows:

\[
U_0 \partial_s V_0 = - \left[ U_0 \partial_s U_1 + (V_1 - C_0) \partial_s U_0 \right] = - (1 - \partial_s U_0) (\partial_s U_a + \partial_s V_a).
\]

That is, the absolute vorticity advection changes the divergence of ageostrophic flow. The divergence of along-
jet ageostrophic flow is caused by not only the planetary vorticity advection [the left-hand side of Eq. (A3)] but also the relative vorticity advection [the left-hand side of Eq. (A4)].

Using Eqs. (A2)–(A4), the PV equation [Eq. (A1)] becomes

\[
(1 - \partial_s U_0) \left[ (V_a - C_0) \partial_s h_0 + h_0 (\partial_s U_a + \partial_s V_a) \right] = 0.
\]

Because the absolute vorticity \( (1 - \partial_s U_0) \) is not always zero, the square bracket term vanishes and we obtain Eq. (22),

\[
(V_a - C_0) \partial_s h_0 + h_0 (\partial_s U_a + \partial_s V_a) = 0.
\]

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