Origin of Decadal-Scale, Eastward-Propagating Heat Content Anomalies in the North Pacific*

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ABSTRACT

Upper ocean heat content (OHC) is at the heart of natural climate variability on interannual-to-decadal time scales, providing climate memory and the source of decadal prediction skill. In the midlatitude North Pacific Ocean, OHC signals are often found to propagate eastward as opposed to the frequently observed westward propagation of sea surface height, another variable that represents the ocean subsurface state. This dichotomy is investigated using a 150-yr coupled GCM integration. Simulated OHC signals are distinguished in terms of two processes that can support eastward propagation: higher baroclinic Rossby wave (RW) modes that are associated with density perturbation, and spiciness anomalies due to density-compensated temperature and salinity anomalies. The analysis herein suggests a unique role of the Kuroshio–Oyashio Extension (KOE) region as an origin of the spiciness and higher mode RW signals. Wind-forced, westward-propagating equivalent barotropic RWs cause meridional shifts of the subarctic front in the KOE region. The associated anomalous circulation crosses mean temperature and salinity gradients and thereby generates spiciness anomalies. These anomalies are advected eastward by the mean currents, while the associated surface temperature anomalies are damped by air–sea heat exchange. The accompanying surface buoyancy flux generates higher baroclinic, eastward-propagating RWs. The results suggest that the large OHC variability in the western boundary currents and their extensions is associated with the spiciness gradients and axial variability of oceanic fronts.

1. Introduction

Pacific decadal variability (PDV) is a crucial low-frequency variability that regulates, together with a global warming trend due to anthropogenic forcing, near-term (10–30 yr) climate and weather in Pacific rim countries, as well as ecosystems in the Pacific Ocean (e.g., Mantua et al. 1997; Nakamura et al. 1997; Minobe 1997; Schneider and Cornuelle 2005; Di Lorenzo et al. 2008; Solomon et al. 2011; Liu 2012). Because of the societal impact of PDV (and the potential for predicting this variability), climate centers have started decadal climate predictions using climate models with initialized ocean states (Smith et al. 2007; Keenlyside et al. 2008; Pohlmann et al. 2009; Mochizuki et al. 2010; Kim et al. 2012, and references therein). The success of such predictions depends in part on the metrics used to assess their forecast skills. Metrics based on subsurface ocean temperature rather than those on sea surface temperature show better predictive skills (Mochizuki et al. 2010; Branstator and Teng 2010; Teng and Branstator 2011; Mochizuki et al. 2012), suggesting that better tracking of the subsurface ocean conditions is key for assessing and improving decadal prediction experiments.

There are two commonly used variables to monitor decadal variability of subsurface oceanic condition: sea surface height (SSH) and ocean heat content (OHC).
These two quantities have different zonal propagation characteristics on interannual-to-decadal time scales. Sea surface height anomalies, as observed with a satellite altimeter, exhibit westward propagation and are interpreted as equivalent barotropic Rossby waves (Seager et al. 2001; Schneider and Miller 2001; Qiu 2003; Qiu and Chen 2005, among others). In contrast, OHC anomalies are often found to propagate eastward as a part of signals that propagate clockwise along the North Pacific subtropical gyre in historical observations (Zhang and Levitus 1997) and also more prominently in atmosphere–ocean coupled general circulation models (CGCM; Latif and Barnett 1994; Kwon and Deser 2007; Teng and Branstator 2011). While advection by the mean flow is suggested as the underlying cause, OHC anomalies in general are active tracers that propagate as baroclinic Rossby waves along trajectories distinct from mean flow advection. Clarifying the dynamics of the propagating OHC anomalies is the subject of this contribution.

A number of processes can account for the dichotomy of the propagation of sea level and heat content anomalies. Surface Ekman pumping primarily excites equivalent barotropic mode Rossby waves that propagate westward largely independent of the wind-driven gyres (e.g., non-Doppler shift mode; Pedlosky 1996; Liu 1999). In contrast, surface buoyancy fluxes excite second and higher baroclinic mode (hereafter simply called higher baroclinic modes) Rossby waves that are strongly affected by the gyres and have eastward ray paths in the Kuroshio Extension (Liu 1999). In contrast to these Rossby waves that are associated with density perturbations (e.g., Liu and Alexander 2007), temperature and salinity anomalies that are density-compensated (i.e., spiciness anomalies; e.g., Veronis 1972; Schneider 2000) are passive tracers and are advected with the flow [see also the introduction in Kilpatrick et al. (2011)]. The frontal regions of the Kuroshio and Oyashio Extensions separate high-spiciness warm and salty waters of subtropical gyre from the low-spiciness cool and fresh subpolar gyre. Latitude shifts of the Kuroshio–Oyashio Extension jets (Seager et al. 2001) may lead to spiciness fluctuations that account for the eastward-propagating heat content anomalies. Accordingly, d’Orgeville and Peltier (2009) have reported that spiciness anomalies in a CGCM are advected eastward along the subpolar gyre from the western and central Pacific to the Bering Sea and set the decadal time scale of variability.

The present study aims to understand the role of these processes in determining the propagation, particularly of the less studied eastward-propagating OHC signals. The first objective is to test the hypothesis on the propagation features of decadal-scale SSH and OHC signals as Rossby waves by examining the vertical structures of these signals. The second objective is to distinguish OHC signals in terms of two processes that can support eastward propagation: higher baroclinic mode Rossby waves and spiciness anomalies. For this purpose, we introduced a diagnostic to split temperature anomalies into parts associated with density and spiciness perturbations. The third objective is to examine the origin of the eastward-propagating Rossby waves and spiciness signals and their linkage to the eastward-propagating sea level anomalies. We use a 150-yr control integration of an ocean–atmosphere CGCM that realistically captures the propagating signals as an internal variability of the simulated coupled system and thus enables the examination of their origins. We will show that the commonly used climate indices of OHC in the upper North Pacific Ocean are dominated by spiciness signals that are generated by wind-driven, decadal-scale, latitudinal shifts of the subarctic front.

The rest of the paper is organized as follows. Section 2 describes the model used in this study, followed by sections 3–5 that present the results. Section 6 provides a summary and discussions.

2. Model

The model used in this study is the CGCM for the Earth Simulator (ES) (CFES; Komori et al. 2008a,b). Its atmospheric component, the atmospheric GCM (AGCM) for the ES (AFES; Ohfuchi et al. 2004; Enomoto et al. 2008; Kuwano-Yoshida et al. 2010), is based on the Center for Climate System Research (CCSR)–National Institute for Environmental Studies (NIES) AGCM, version 5.4.02 (Numaguti et al. 1997). The ocean component, Coupled Ocean–Sea Ice Model for the ES (OIFES; Komori et al. 2005), is based on the Modular Ocean Model, version 3 (MOM3; Pacanowski and Griffies 2000). Computational codes of AFES and OIFES have been substantially rewritten to attain their high computational efficiency on the particular architecture of ES and to improve physical parameterizations.

We use a 150-yr control simulation of CFES that has been integrated at a medium resolution where the atmospheric component has a horizontal resolution of T119 and the ocean component has 0.5° z-coordinate levels and the ocean component has 0.5° latitude–longitude grid intervals with 54 z-coordinate levels. The initial conditions for OIFES are temperature and salinity fields with no motion for the ocean component, taken from climatological January mean fields of World Ocean Atlas 1998 (Boyer et al. 1998a,b,c), whereas those for AFES are the climatological atmospheric fields for 0000 UTC 1 January based on the 40-yr European
Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). Greenhouse gas concentrations are set at present-day values. The coupling interval of AFES and OIFES is 1 h. This particular version of CFES and another version configured with a higher horizontal resolution have been used for climate studies on North Pacific decadal variability (Taguchi et al. 2012b; Miyasaka et al. 2014), tropical Atlantic variability (Richter et al. 2010), and extratropical air–sea interaction at oceanic mesoscales/fronts (Nonaka et al. 2009; Komori et al. 2008b; Taguchi et al. 2012a; H. Sasaki et al. 2013; Kuwano-Yoshida et al. 2013).

3. Propagating signals and vertical structures

This section describes the interannual-to-decadal variability of SSH and OHC in the North Pacific Ocean simulated in CFES. To differentiate dynamical processes associated with the signals as outlined in section 1, we focus on their propagation features and vertical structures.

a. Interannual-to-decadal variability of SSH and OHC

Despite the relatively coarse resolution of the CFES’s ocean component that does not resolve oceanic mesoscale eddies, the CFES integration represents salient features of the North Pacific Ocean’s circulation and variability. Most prominent in the mean field are the two extensions of western boundary currents in the so-called Kuroshio–Oyashio Extension (KOE) region1: the subarctic front (SAF) and the Kuroshio Extension (KE) (e.g., Mizuno and White 1983; Yuan and Talley 1996; Yasuda 2003; Nonaka et al. 2006). The former is characterized by sharp, largely density-compensating gradients in temperature and salinity in the upper ~200 m of the ocean, and is discernible in the simulated annual mean OHC field (Fig. 1a), defined here as temperature vertically averaged over the upper 400 m of depth. While this depth was chosen for consistency with earlier studies discussed in the introduction, a choice of 700 m as the lower boundary does not impact results significantly (see section 3c). The SAF manifests also as an OHC front that extends from ~41°N just off Japan toward the east-northeast. The KE has much deeper vertical structure with the main pycnocline reaching down to ~700 m and is better detected in the SSH field.

The simulated annual mean SSH field exhibits frontal structures, the KE of CFES, at ~38°N and from the east of Japan to 170°E, along the northern periphery of the subtropical gyre (Fig. 1b). Because of the limited resolution, the simulated Kuroshio overshoots beyond the latitude of ~35°N where the observed Kuroshio separates from the Japanese coast.

Interannual-to-decadal variability of SSH and OHC anomalies are narrowly confined along the KE and the SAF, respectively (shading in Fig. 1) and caused by latitudinal excursion of the fronts due to wind stress curl variation in the extratropical North Pacific basin. The pronounced decadal variability of SAF simulated in CFES is documented in Taguchi et al. (2012b) along with the associated SST, sea surface heat flux, and atmospheric circulation fields.

b. Zonal propagation

Simulated interannual-to-decadal anomalies of SSH and OHC show distinct zonal propagation across the North Pacific basin. Since the SSH and OHC fronts and their centers of variability are located at different latitudes, we show the propagation along two zonal bands marked by their variance maxima. The northern band is defined as a region where the standard deviation of OHC exceeds 0.9 K in a tilted 10°-wide semizonal band roughly following the SAF from 40°N, 142°E to 48°N, 130°W. Signals are averaged with latitude over the color shaded region in Fig. 1a, while the southern band is defined as a zonal band between 36° and 39°N (Fig. 1b).

SSH signals along the southern band are dominated by westward propagation (Fig. 2a), consistent with an equivalent barotropic mode Rossby wave forced by Ekman pumping associated with wind stress curl in the North Pacific basin, as shown earlier using satellite observations (Qiu 2003; Qiu and Chen 2005, 2010; Y. Sasaki et al. 2013) and ocean models (Miller et al. 1994, 1998; Seager et al. 2001; Schneider and Miller 2001; Taguchi et al. 2005, 2007; Sasaki and Schneider 2011; Nonaka et al. 2012). The phase speed of the propagation in CFES is estimated to be about ~7.1 cm s\(^{-1}\) (westward) in this latitudinal band, 2.5 times faster than observed (e.g., Qiu 2003).

Contrary to the SSH propagation, the propagation of OHC signals along the northern band displays a distinct eastward propagation with the speed of 4.6 cm s\(^{-1}\) (Fig. 2b), matching the zonal current velocity averaged over the upper layer of 0–400 m (Fig. 1c). The propagation of OHC signals along the southern band, on the other hand, exhibits mixed propagation across the basin (Fig. 2c). Note that to the west of 170°E OHC signals are highly correlated (>0.9 at a given longitude) with SSH anomalies, so both capture the same dynamics. In the

1 Although there is no such current as the “Oyashio Extension,” we use the term KOE for brevity throughout the manuscript to refer to the region that encompasses the subarctic front and the Kuroshio Extension.
FIG. 1. Climatological mean (contours) and standard deviation (color shading) of annual mean (a) OHC (°C and K, respectively) and (b) SSH (cm) anomalies. Red lines superimposed in (a) designate meridional boundaries of the northern band, between which quantities are meridionally averaged for analyses of propagation features and vertical structures discussed in section 3. Red lines superimposed in (b) designate the southern band. See text for the physical basis for the definition of these two analysis bands. (c) Climatological mean of annual zonal current velocity (color shading; cm s⁻¹) averaged over the surface to 400-m depth. Superposed with contours is the mean SSH as in (b). All quantities are based on the 150-yr CFES integration.
FIG. 2. Longitude–time sections of (a) SSH (cm) anomalies along the southern band, (b) OHC (K) anomalies along the northern band, (c) OHC (K) anomalies along the southern band, and (d) principal component of subsurface density anomalies $A_2(x,t)$ (kg m$^{-3}$) based on the 150-yr CFES integration. The anomalies of SSH and OHC are from their monthly climatological mean with trends removed and are averaged over the respective (semi) latitudinal bands. The southern band is defined as a latitudinal band between 36° and 39°N. The northern band is defined as a region where the standard deviation of OHC exceeds 0.9 K in a tilted 10°-wide semilatitudinal band roughly following the SAFZ (color shaded region between the two red lines in Fig. 1a). Thick green lines indicate constant phase speeds of $-7.1$ cm s$^{-1}$ (westward) in (a) and (c) and 4.6 cm s$^{-1}$ (eastward) in (b).
central and eastern half of the basin, however, OHC and SSH anomalies have independent components and the former shows a robust eastward signal absent from the latter. These contrasting propagation features between SSH and OHC anomalies are consistent with observational and modeling studies as discussed in section 1. The eastward-propagating OHC signals along the northern band originate near the western boundary, following the arrival of the westward-propagating SSH signals of the same sign along the southern band and suggest a coupling of these eastward-propagating OHC signals and the westward-propagating SSH signals (see section 5).

c. Vertical structures

To determine the dynamical processes that govern the propagation of SSH and OHC, we investigate the associated vertical structures. Mean potential density fields contoured in Fig. 3a show that the southern section cuts across quasi-stationary undulations of the Kuroshio jet west of 150°E and enters the northern edge of the subtropical gyre with its deep pycnocline around 150°–170°E. At the date line, this section samples a shallowing pycnocline and strong surface stratification of the eastern North Pacific (Fig. 3a). The pycnocline in the east is largely maintained by mean salinity stratification below 100-m depth (contours in Fig. 3d) where the mean temperature field shows the existence of an inversion around 100–300-m depth between 165° and 135°W (contours in Fig. 3c) as has been found in the climatology of the observed temperature and salinity profiles (de Boyer Montégut et al. 2007). The simulated mean temperature inversion is formed between the mesothermal layer (temperature minimum) around 100–200-m depth and the dicothermal layer (temperature maximum) around 300–400-m depth, which is associated with high salinity due to advection of warm and saline water (e.g., Ueno and Yasuda 2000). In the northern section, the pycnocline resides at a shallower depth to the west of the date line and stratification and zonal gradient are weaker than the southern section (Fig. 3b).

The standard deviation of potential density exhibits distinct vertical structures between the southern and the northern sections (shading in Figs. 3a,b). In the southern section, standard deviations exceeding 0.2 kg m⁻³ are confined to the upper 300 m above the mean pycnocline with two centers of action: one in the region of the western boundary current and recirculation gyre, and the other in the far eastern Pacific (Fig. 3a). Two representative sites in the western and eastern side of the basin at 150°E and 155°W, respectively, capture this zonal variation of density standard deviation and mean structure (red and black curves in Figs. 4a,b). In contrast, in the northern section the potential density variability is weak, and of the same order as that below the pycnocline in the southern section. This indicates that baroclinic Rossby wave dynamics are less prominent for the OHC along the northern band compared to the southern band. Thus, we focus on the southern section in the following discussion to further characterize the density anomalies associated with the propagating signals.

To examine the density signatures associated with the propagating signals along the southern section, depth–longitude sections are constructed for the regression coefficients of potential density anomalies onto local anomalies of SSH (Fig. 3c) and OHC (Fig. 3d). Westward-propagating signals, correlated with the SSH anomalies, are associated with a single-signed vertical structure with maximum loading overall at the depth of the pycnocline consistent with the equivalent barotropic Rossby waves across the entire basin. Potential density anomalies in the pycnostad layer over the depth of 100–400 m between 150°E and 180° result from meridional displacements of pycnoclines that shoal northward (Fig. 3e). Regressions with the OHC anomalies also bear the equivalent barotropic mode structure west of the date line, synonymous with the SSH anomalies (red solid and dotted lines in Fig. 4c). From the central to the eastern basin to the east of the date line, however, the density signature of the SSH and OHC anomalies are very different (black solid and dotted lines in Fig. 4c), the latter exhibiting sign reversal in the vertical around the depth of the lower pycnocline (~200 m at 155°W as shown with black dotted line) and suggesting higher baroclinic modes Rossby waves. Within this latitudinal band along the northern periphery of the subtropical gyre, mean ocean currents are eastward (Fig. 1c) and are expected to advecr higher baroclinic mode waves toward the east, consistent with the dominant eastward propagation of OHC anomalies in the central to eastern part of the North Pacific basin (Fig. 2c). The equivalent barotropic structure to the west is expected to be unaffected by the mean current, because of the cancellation of the wave propagation tendencies due to advection and change of background meridional vorticity gradient (Pedlosky 1996; Liu 1999). Similar eastward-propagating higher baroclinic mode signals have been discussed with ocean models forced with prescribed surface buoyancy forcing (Liu 1999; Nonaka and Xie 2000; Thompson and Ladd 2004; Osafune and Yasuda 2012). Here, these signals emerge as an internal variability of the coupled ocean–atmosphere system and allow us to examine their origins (see section 5).

The density signature of higher baroclinic modes associated with the eastward-propagating OHC anomalies is insensitive to the depth range of OHC. The regression of temperature and salinity anomaly profiles at the
eastern site ($155^\circ$W) onto the local OHC anomalies (Fig. 4d) demonstrates that the temperature signal has a deep and single-signed vertical structure (red curve) thus cannot produce the vertical structure with a sign reversal in the associated density anomalies (black curve). The density anomalies of higher baroclinic modes result from salinity contributions confined in the upper 200-m depth and partially offset by the temperature contribution. Because of the single-signed structure in temperature contribution, repeating the analysis of the propagation features and the vertical structures of OHC anomalies defined as vertical averaged temperature over
the upper 700 m, instead of 400 m, hardly alters the results (not shown).

d. Empirical mode decomposition

Section 3c identified the westward (eastward)-propagating equivalent barotropic (higher baroclinic) mode Rossby waves from the density anomalies in the southern section by regression analysis. These signals can also be separated by an empirical mode decomposition. Meridionally averaged between 36° and 39°N, the density anomalies \( \rho'(x, z, t) \) in the vertical section between 142°E and 120°W and over the entire ocean depth are decomposed into empirical orthogonal functions (EOFs) as follows:

![Diagram of vertical profiles](image-url)
where \( i \) denotes the EOF mode; \( A_i \) is the principal component (PC); \( \Phi_i \) is the eigenvector; and \( x \) and \( t \) designate longitude and time, respectively. The vertical structures of the two leading EOF eigenvectors \( \Phi_{1,2} \) are shown in Fig. 4c. The first mode (red curve), which explains 50.6\% of the total variance, shows a single-signed vertical structure with the maximum loading around 100-m depth and resembles the equivalent barotropic mode structure associated with the westward-propagating SSH and OHC anomalies near the western boundary (red curves in Fig. 4c). Consistently, the spatiotemporal structure of the first principal component \( A_1(x,t) \) is characterized by the westward propagation and the large variance in the west (not shown). The second mode (black curve in Fig. 4e), which explains 25.1\% of the total variance, displays a sign reversal around 150-m depth and thus resembles the higher baroclinic mode structure associated with the eastward-propagating OHC anomalies in the east (black dotted curve in Fig. 4c), although the zero crossing depth in the former is slightly shallower than the latter. The Hovmöller diagram of the second principal component \( A_2(x,t) \) shown in Fig. 2d shows eastward propagation that enhances its amplitude in the eastern part of the basin, which well captures the propagation feature of the OHC anomalies in the central to the eastern parts of the basin (Fig. 2c). Based on the high fidelity of the second EOF mode in capturing the higher baroclinic mode signals in this particular zonal section, we use the projection of the density anomalies onto the second EOF eigenfunction \( A_2(x,y,t) \) as a proxy of the extended amplitude of the higher baroclinic mode signals in the entire North Pacific basin for the discussion on their origin (section 5):

\[
A_2(x,y,t) = \int_0^\infty \rho'(x,y,z,t) \Phi_2(z) \, dz, \tag{2}
\]

where \( y \) denotes latitude, and the integral is over the entire ocean depth from the bottom \( -H \) to the surface.

In summary, both the empirical vertical mode decomposition using EOF and the vertical projection using regression analysis discussed in this section consistently separate two dynamical signals in the southern zonal section (36°–39°N) out of the CFES simulation: the westward-propagating, equivalent barotropic mode signals detected in SSH and OHC anomalies in the western half of the Pacific basin, and the higher baroclinic mode signals detected in the OHC anomalies from the central to eastern part of the basin.

4. Density and spiciness components of OHC

As mentioned in section 1, the OHC anomalies may evolve as an active tracer because temperature affects seawater density in a monotonic fashion, or may be in part density compensated by salinity anomalies, leading to so-called spiciness anomalies that evolve as a passive tracer. In the previous section, the propagating SSH and OHC signals in the latitudes 36°–39°N are linked with the equivalent barotropic and higher baroclinic mode Rossby wave signals. However, the eastward-propagating OHC signals in the southern band may partly be a manifestation of the advection of spiciness anomalies by background eastward currents. Furthermore, at the northern analysis band along the SAF, eastward-propagating OHC anomalies are associated with weak density variations, and are expected to be dominated by spiciness variations. To characterize and explore the fate of OHC anomalies, it is therefore desirable to split heat content anomalies into parts associated with density and spiciness perturbations, a transformation discussed in this section.

a. Temperature anomaly decomposition

The temperature anomalies described above and associated changes of salinity imply displacements of isopycnals and changes of temperature along isopycnals, called spiciness anomalies, with distinct propagation characteristics of Rossby waves or passive tracers, respectively. To explore the role of spiciness in the propagation of heat content, we generalize to three dimensions the linear decomposition of temperature and salinity fluctuations to temperature and spiciness anomalies, with distinct propagation characteristics of Rossby waves or passive tracers, respectively. To explore the role of spiciness in the propagation of heat content, we generalize to three dimensions the linear decomposition of temperature and salinity fluctuations to temperature and salinity anomalies associated with isopycnal heave and with change of spiciness along isopycnals (Bindoff and McDougall 1994; Schneider et al. 2005).

Instantaneous fields of temperature \( T \) and of potential density\(^2 \) \( \rho \) at location \( x \) are interpreted as spatial displacements of their temporal averages:

\[
\phi(x) = \overline{\delta T}(x - \delta x), \tag{3}
\]

where variable \( \phi \) with average denoted by an overbar \( \overline{\ } \) stands for \( T \) or \( \rho \), and the displacement vector \( \delta x \) includes vertical and lateral shifts resulting from adiabatic advection and diabatic processes.

For values of \( \delta x \) that are small compared to the ratio of gradient and curvature of density and temperature, anomalies \( \delta \phi = \phi - \overline{\phi} \) are linearized:

\[\text{footnote 2 }\text{Alternatively, neutral density (Jackett and McDougall 1997) can be used here. You (2010) reported some differences (≈0.2 kg m}^{-3}\text{) between potential and neutral density surfaces in the North Pacific SAFZ.}
\[ \delta \phi = -\nabla \phi \cdot \delta \mathbf{x}. \]  

(4)

Writing \( \delta \mathbf{x} = \delta \mathbf{x}_d + \delta \mathbf{x}_s \) in components aligned and perpendicular to the time mean density gradient

\[ \delta \mathbf{x} = \left( \delta \mathbf{x}_d \cdot \frac{\nabla \rho}{|\nabla \rho|} \right) \frac{\nabla \rho}{|\nabla \rho|} + \delta \mathbf{x}_s \]  

(5)

yields with (4) applied to \( \rho \):

\[ \delta \mathbf{x} = -\delta \rho \frac{\nabla \rho}{|\nabla \rho|^2} + \delta \mathbf{x}_s. \]  

(6)

The application of (4) to \( T \)

\[ \delta T = \frac{\nabla T \cdot \nabla \rho}{|\nabla \rho|^2} \delta \rho + \delta T_s. \]  

(7)

then yields as the first term on the right the component \( \delta T_s \) associated with the deflection of the density field. The term \( \delta T_s = -\nabla T \cdot \delta \mathbf{x}_s \) is the spiciness component with density-compensating contributions due to temperature and salinity, and results from displacements \( \delta \mathbf{x} \) perpendicular to the density gradient.

Vertical integration of the components \( \delta T_p \) and \( \delta T_s \) yields the active and passive components of heat content. The limits of the integration can be fixed levels such as the surface and 400 m, or isopycnal levels; time mean (or winter) isopycnals suffice based on the linearization in (4). Care has to be taken when including the surface mixed layer and its large seasonal cycle, but results may be insensitive if the depth range of the integration reaches much deeper than the mixed layer.

b. Relative contribution of density and spiciness components

We now apply the temperature anomaly decomposition (7) to the CFES simulation to quantify the relative contribution to the density and spiciness components from the surface to 400-m depth heat content: OHC\(_p\) and OHC\(_s\). The standard deviation of annually averaged OHC\(_p\) and OHC\(_s\) (Fig. 5) show that the majority of the total OHC variability (Fig. 1a) is associated with the spiciness component OHC\(_s\). The density component OHC\(_p\) emerges in regions off Japan and northeast of Hawaii where the variability of total OHC and SSH are both large. The former is due to the variability of the simulated Kuroshio Extension, while the latter represents that of the subtropical countercurrent (Xie et al. 2011; Kobashi and Kubokawa 2012). Consistent with the map of the standard deviation, Hovmöller diagrams for the northern analysis band along the SAF (Figs. 6a,b) display dynamical component signals of OHC\(_p\) that are nearly absent while the propagation feature of spiciness component OHC\(_s\) replicates the signal of the total OHC (Fig. 2b). The same diagrams for the southern band of 36°–39°N (Figs. 6c,d) also demonstrate that the westward-propagating OHC\(_p\) is only distinct in the far west and that OHC\(_s\) dominates in the central-to-eastern basin. These analyses suggest that the spiciness component \( \delta T_s \) dominates the eastward-propagating upper ocean OHC anomalies.

While in the northern band spiciness gives rise to most of the OHC variability because the density signature there is very weak (Fig. 3b), in the southern band the dominance of spiciness over the density component of OHC results from the averaging out of the higher mode vertical structure of \( \delta T_p \) in the calculation of OHC. The vertical structures of \( \delta T_p \) and \( \delta T_s \) in latitudes of 36°–39°N (Fig. 7) shows single-signed \( \delta T_p \) west of 170°E, suggestive of downward displacements \( \delta T_s \) associated with the equivalent barotropic mode Rossby waves. This single-signed \( \delta T_p \) integrates into the OHC anomalies that propagate westward (see Fig. 2c). East of the date line, \( \delta T_p \) exhibits a tripolar structure in the vertical, negative in and above the pycnocline, and positive underneath, consistent with higher baroclinic modes. The negative \( \delta T_p \) underneath the pycnocline around 170°E–150°W\(^3\) is due to climatological mean temperature inversions (contours in Fig. 3c) that yield positive \( \nabla T \cdot \nabla \rho \) (not shown) in (7). Their contributions to the integration of OHC\(_p\) from the surface to 400-m depth largely cancel out, resulting in the insignificant contribution of OHC\(_p\) (Fig. 7a). On the other hand, \( \delta T_s \) is mostly single signed across the entire basin (Fig. 7d), and its vertical integral OHC\(_s\) therefore dominates the total OHC signal (Fig. 7b).

5. Origin of eastward-propagating signals

As the spiciness anomalies are the main constituent for the upper ocean heat content anomalies (section 4), we document in this section how the westward-propagating equivalent barotropic Rossby waves generate spiciness signals near the western boundary, and how the eastward-propagating spiciness anomalies lead to the higher baroclinic mode signals.

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\(^3\)Looking at \( \delta T_s \), which has a similar vertical structure and the opposite sign to \( \delta T_p \) here, one might be concerned that the temperature decomposition artificially generates the \( \delta T_s \) and \( \delta T_p \) anomalies out of nothing. Such a situation, however, would not occur because if \( \delta T = 0 \) but \( \delta S \neq 0 \), then \( \delta T_p \) and \( \delta T_s \) both should vanish as follows. Suppose that the equation of state is linearized consistent with the linear approximation in (7), so that

\[ \delta \rho = a_0 \delta T + \beta \delta S, \] where \( a_0 \) and \( \beta \) are the thermal and haline expansion coefficients, and \( a_0 \delta T + \beta \delta S = 0 \). Since \( \delta \rho = \beta \delta S \), \( \delta S = 0 \) therefore \( \delta T_s = 0 \) and thus with (7) \( \delta T_p = 0 \).
a. Generation of spiciness anomalies

The eastward-propagating spiciness anomalies (Figs. 6b,d) are generated in the western part of the North Pacific Ocean (i.e., the KOE region) by anomalous advection across mean spiciness gradients that has been reported in such regions as subtropical central Pacific (Schneider 2000) and subtropical eastern Pacific (Sasaki et al. 2010; Kilpatrick et al. 2011). To test this hypothesis, we examine the upper ocean mean spiciness gradient measured as along-isopycnal mean temperature gradients averaged over the depth range of 0–400 m,

\[
\frac{1}{400m} \int_{0}^{400m} (\nabla T)_x \, dz,
\]

and where \((\nabla T)_x\) is the temperature gradient across isopycnals. The mean spiciness gradients are particularly large along the subarctic frontal zone (SAFZ; contours in Fig. 5b), where the northward decrease of temperature partially offset by the northward decrease of salinity corresponds to large OHC$_x$ variability. This suggests the KOE region, particularly along the subarctic front, as the site of spiciness anomaly generation.

Furthermore, the North Pacific subarctic front exhibits a distinct interannual-to-decadal variability in its axial
position (e.g., Nakamura and Kazumin 2003; Frankignoul et al. 2011), due to oceanic adjustments to basin-scale wind forcing via barotropic and baroclinic Rossby waves (Seager et al. 2001; Nonaka et al. 2008). The CFES simulation successfully represents this axial variability of the subarctic front (Fig. 8a; see also Taguchi et al. 2012b). The northward component of anomalous currents associated with the axial variability of the subarctic front (vectors in Fig. 8b) intersects the mean spiciness gradients off northern Japan around 40°N (purple contours in Fig. 8b), thereby generating the anomalous advection of the spiciness anomalies (color shading in Fig. 8b).

The generation of spiciness anomalies in the KOE region can be confirmed by tracing back in time the spiciness component of the OHC (i.e., OHC_x) at the date line. Figure 9a shows lag-correlation/regression of

![Fig. 6](image_url)

*Fig. 6. (a) Longitude–time diagram along the northern band of OHC_p (K) computed from the density component of temperature anomalies (i.e., δT_p) based on the 150-yr CFES integration. (b) As in (a), but for the spiciness component (i.e., OHC_x) computed from δT_x. The anomalies are from their monthly climatological mean with trends removed and are averaged over the latitudinal band. (c),(d) As in (a),(b), but for the southern band.*

![Fig. 7](image_url)

*Fig. 7. (a) δT_p averaged over the upper 400-m depth. (b) As in (a), but for δT_x. (c),(d) Regression coefficients (color shading; K) of temperature anomalies onto the standardized local OHC anomalies in the longitude–depth section along the southern band: (c) density (i.e., δT_p) and (d) spiciness components (i.e., δT_x). Superimposed with contours is the annual mean potential density to depict the main pycnocline. All the quantities are based on the 150-yr CFES integration.*
OHC anomalies onto a reference time series of OHC at 43°N, 180°. The eastward-propagating OHC anomalies (Fig. 9a) originate from the KOE region at a −6-yr lag (i.e., 6 years ahead of the OHC signal passing across the date line). This generation of OHC anomalies in the KOE region is confirmed by a high correlation in 150-yr simulated time series (0.56 for raw monthly series and 0.68 for 3-yr running mean series) between OHC anomalies averaged over the upstream SAFZ (39°–44°N, 145°–160°E) and the latitudinal fluctuation of the SAFZ (Fig. 8a). This is also consistent with the good correspondence between the occurrence of large SSH anomalies in Fig. 2a and the generation of spiciness anomalies in Fig. 6b near the western boundary.

b. Generation of higher baroclinic modes

While the spiciness anomalies dominate the total OHC variability in the northern analysis band along the SAF with little density signature, the spiciness and the higher baroclinic mode signals exist in the southern analysis band in latitudes of 36°–39°N. We show in this section how the spiciness anomalies generate the higher baroclinic mode Rossby waves. The eastward propagation of OHC is associated with SST anomalies (blue
contours in Fig. 9a) that are damped by air–sea heat flux anomalies (upward heat flux anomalies for positive SST anomalies and vice versa; blue contours in Fig. 9b) at a rate of 40 W m$^{-2}$ K$^{-1}$ along the SAF in the KOE region. The eastward copropagation of SST and air–sea heat flux anomalies has also been found in the Community Climate System Model, version 2 (CCSM2; Kwon and Deser 2007) and an observational study by Frankignoul and Kestenare (2002) with similar rates. The air–sea heat flux disrupts the density compensation in the spiciness anomalies and gives rise to the higher baroclinic mode Rossby waves (Liu 1999). Hence, the growth and the eastward propagation of the higher baroclinic mode Rossby waves as measured with the extended second principal component $A'_2(x, t)$ of the subsurface density anomalies in the southern section (see section 3d for the

![FIG. 9. Lagged correlation (shading; contour intervals shown in color bar at the bottom) and regression coefficients (black contours) of the following quantities onto OHC$_x$ at 43$^\circ$N, 180$^\circ$: (a) OHC$_x$ (every 0.2 K K$^{-1}$) and (b) amplitude of higher baroclinic modes as measured with the extended second principal component $A'_2(x, t)$ of the density anomalies along the southern band over the entire ocean depth (see Fig. 4e for the vertical structure of its EOF eigenfunction; every 0.15 kg m$^{-2}$ K$^{-1}$). Superimposed with blue contours are regression coefficients onto OHC$_x$ at 43$^\circ$N, 180$^\circ$ of SST (0.2, 0.4, and 0.6 K K$^{-1}$) in (a) and sea surface turbulent heat flux (5, 10, 15, 20, and 25 W m$^{-2}$ K$^{-1}$) in (b).]
definition) can be detected downstream of the heat flux anomalies (Fig. 9b). This linkage of the generation of the higher baroclinic modes with the damping of SST anomalies is consistent with the earlier studies using ocean models where surface buoyancy flux forcing was prescribed (Liu and Shin 1999; Nonaka and Xie 2000; Osafune and Yasuda 2012; Thompson and Ladd 2004). In particular, Thompson and Ladd (2004) identified the forcing of the first and the second baroclinic mode variability separately in their OGCM simulation through a modal decomposition analysis, where the second mode variability is linked to surface buoyancy forcing. Our CGCM study furthermore establishes that the surface buoyancy forcing due to damping of the SST anomalies is induced by the preceding OHCx anomalies.

There is a subtle difference in the path along which the spiciness and higher baroclinic mode Rossby waves propagate. The OHCx anomalies propagate east-northeastward, deflecting slightly northward of the latitudes where SST and air–sea heat flux anomalies are large. That part of the OHCx anomalies that roughly follows our northern analysis band does not carry much of the SST anomalies after moving to the east-northeast (at 0- and +2-yr lag in Fig. 9a). Therefore, the generation and propagation of the higher baroclinic modes is confined to latitudes south of the SAF, a part of which is detected in the southern analysis band (Fig. 2d), while the OHC variability along the SAF is dominated by the spiciness component that accompanies neither much of the density anomalies nor the higher mode baroclinic Rossby waves.

6. Summary and discussion

We investigate the dynamics of the decadal-scale propagating signals of sea surface height (SSH) and upper ocean heat content (OHC) anomalies in the North Pacific simulated in a 150-yr coupled atmosphere–ocean general circulation model (CGCM) integration. Our analysis suggests that the following sequence of processes are involved in the generation and dynamical causes of the SSH and OHC signals. First, the arrival of wind-forced, westward-propagating equivalent barotropic Rossby waves at the western boundary causes a meridional shift of the gyre boundary in the Kuroshio–Oyashio Extension (KOE) region as most evident in the SSH field. Because the KOE region, particularly the subarctic frontal zone (SAFZ), is a region of large meridional spiciness gradients, anomalous advection associated with the decadal-scale fluctuation of the SAFZ position generates spiciness anomalies, which are then advected eastward by the mean currents. The associated surface temperature anomalies are damped by air–sea heat exchange and thereby generate density anomalies, which propagate eastward as higher mode Rossby waves.

A decomposition of temperature anomalies into parts associated with dynamical (density) and with spiciness anomalies reveals that the OHC anomalies are dominated by the spiciness contribution. Because of the vertical structures of dynamical and spiciness components of temperature anomalies associated with OHC anomalies, the former tends to largely cancel out in the vertical integration, while the latter remains. The dominance of the spiciness component of OHC anomalies highlights an important role of the KOE region as an origin of the eastward-propagating OHC signals. In earlier forced OGCM experiments, such eastward-propagating, higher mode Rossby waves or spiciness anomalies are initiated by the surface buoyancy forcing (e.g., Liu and Shin 1999; Nonaka and Xie 2000; Osafune and Yasuda 2012; Thompson and Ladd 2004). Our CGCM simulation points to a new mechanism for the initiation of the eastward-propagating signals, namely the anomalous spiciness generation associated with the decadal-scale latitudinal excursion of the SAFZ in KOE region and the subsequent transformation of the spiciness signals into higher baroclinic modes Rossby waves.

Spiciness anomalies due to anomalous advection have been reported in the subtropical central Pacific (Schneider 2000) and subtropical eastern Pacific (Kilpatrick et al. 2011; Sasaki et al. 2010) downstream of the subtropical subduction regions. Our study extends the spiciness generation process due to anomalous advection into the KOE region with its large mean spiciness gradients between salty and warm subtropical and fresh and cold subpolar waters, and links the spiciness anomalies to the eastward-propagating OHC anomalies that occur upstream of subtropical subduction regions.

While this study focuses on decadal-scale latitudinal displacement of the SAFZ to transform the wind-forced westward-propagating SSH signals into the eastward-propagating OHC signals, altimeter observations sustained for two decades have revealed another prominent decadal variability involving the Kuroshio Extension (KE), namely the decadal swing between the stable and unstable dynamical states of the KE (Qiu and Chen 2005, 2010, 2011). When the KE is in a stable state, the KE jet is intensified and zonally elongated, its mean path shifts northward with reduced path variability, the eddy activity in the upstream (downstream) KE region is weakened (strengthened), and the formation of the subtropical mode water (a lighter variety of the central mode water) in winter is active (inactive) (Oka et al. 2012). The situation is reversed when the KE takes the unstable state with its jet path zonally contracted.
Analyzing available temperature and salinity data from profiling float and hydrographic measurements during the past decade, Qiu and Chen (2011) demonstrated that the salinity in the North Pacific Intermediate Water (e.g., Yasuda et al. 1996) core layer of 26.7 to 26.8 \( \sigma_t \) (potential density) underwent a distinct decadal change in the KE region, which is induced by the modulation of the eddy activity associated with the transition of the KE’s dynamical state. Such a process is not represented in the CFES simulation analyzed in this study due to the coarse resolution (0.5°) of its ocean component as in most of the current-generation climate models. The non-eddy-resolving ocean component lacks lateral mixing due to eddies in this region, and may overestimate meridional gradients in temperature and salinity fields, particularly across the SAFZ (Taguchi et al. 2012b). Therefore, the dynamics elucidated in this study may overemphasize the role of the SAF variability in generating the spiciness anomalies. Nevertheless, our CGCM simulation and analysis portray important dynamics of the decadal-scale SSH and OHC variability, reconciling the dichotomy of their contrasting propagation features often detected in both observations and climate simulations. The companion analysis of ocean reanalysis products and an eddy-resolving OGCM is underway.

To further extend the present study, it is of interest to revisit mode water variability (e.g., Hanawa and Talley 2001; Suga et al. 2004; Oka and Qiu 2012) in relation to higher baroclinic mode Rossby waves and spiciness. For instance, Thompson and Ladd (2004) suggested based on their OGCM simulation a link between anomalous subduction that is identified as the forcing of the second baroclinic mode and the formation of the central mode water (CMW). Our study suggests that the eastward-propagating higher baroclinic mode Rossby waves associated with positive OHC anomalies induce heave of the isopycnals in the shallow pycnocline and the deepening of the isopycnals underneath (Fig. 3d), that is, thickening of the simulated mode water. Furthermore, Suga et al. (2012) argued the possibility that the northward shift and the acceleration of the Kuroshio Extension jet caused a distinct increase in SST and sea surface salinity of the central mode water in the late 1980s regime shift, analogous to the generation of spiciness anomalies along the SAF in our study. The linkage between the variability of the higher baroclinic mode Rossby waves and spiciness simulated in this study and mode water variability in reality should be established in future studies. The eastward-propagating OHC signals could recirculate either in the subpolar gyre or the subarctic gyre. In downstream regions, they could affect SST and the overlying atmosphere and/or could be entrained in the subtropical thermocline. The fate of these signals is important in light of climate memory, as earlier CGCM studies identified the eastward-propagating subsurface anomalies as a necessary source for setting the decadal time scale (Kwon and Deser 2007; d’Orgeville and Peltier 2009), another outstanding topic for future studies.

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