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Title	Impacts of Salinity Variation on the Mixed-Layer Processes and Sea Surface Temperature in the Kuroshio-Oyashio Confluence Region
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Citation	Journal of Geophysical Research Oceans, 126(8), e2020JC016914 https://doi.org/10.1029/2020JC016914
Issue Date	2021-08
Doc URL	http://hdl.handle.net/2115/83993
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Туре	article (author version)
File Information	Kido_et_al(2021)_KOCR_ori.pdf



1	Impacts of salinity variation on the mixed-layer processes
2	and sea surface temperature in the Kuroshio-Oyashio
3	confluence region
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10	(4 th revision)
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17	Key points:
18	1. Coherent temperature and salinity variations are identified in the Kuroshio-Oyashio
19	confluence region during boreal winter to spring
20	2. The dynamical stability of the Kuroshio Extension is the key factor responsible for these
21	temperature and salinity variations
22	3. Changes in density field associated with salinity anomalies significantly affect the strength of

23 vertical mixing and sea surface temperature

24 Abstract:

25 In this study, salinity variations in the Kuroshio–Oyashio confluence region (KOCR) are 26 examined through analyses of observational datasets and an ocean reanalysis product, and their 27 potential impacts on sea surface temperature are assessed by sensitivity experiments using a one-28 dimensional mixed layer model (1-D ML model). We have detected prominent covariations in 29 near surface temperature and salinity in the KOCR during the boreal winter to spring. Further 30 investigation revealed that such covariations are closely related to the dynamical stability of the 31 Kuroshio Extension (KE), and anomalous warming and saltening (cooling and freshening) are 32 observed in the KOCR when the upstream of the KE is in an unstable (a stable) state. It is found 33 that modulation heat and freshwater transport by mesoscale eddies and large-scale current 34 anomalies are closely related to such observed variation. Then, we have quantitatively estimated 35 the impacts of these salinity variations on local density by a detailed decomposition of total 36 anomaly fields. Although the total density anomalies are dominated by contributions from 37 temperature, the salinity contribution has sizable magnitude especially in the northern part of the 38 KOCR, where the background temperature is low and the dependence of density on temperature 39 variations is weak. To further quantify the impact of salinity anomalies, we conducted a series of 40 sensitivity experiments utilizing the 1-D ML model. The results from these experiments revealed 41 that salinity anomalies significantly alter the strength of vertical mixing and eventually lead to 42 differences in sea surface temperature of approximately 1.0 °C.

 $43 \qquad (245 words < 250 words)$

45 **Plain Language summary:**

46 The Kuroshio–Oyashio confluence region (KOCR) in the western North Pacific Ocean 47 undergoes well-defined low-frequency variation on interannual to decadal time scales, and play a 48 pivotal role in the climate variability of the North Pacific. Although significant progress has been 49 made in understanding the dynamical and thermodynamical characteristics of the KOCR 50 variations, less attention has been paid to salinity variability. With the variation in the density of 51 seawater, salinity can potentially exert significant effects on various physical processes in the 52 upper ocean. Therefore, it is important to properly describe its features and assess its possible 53 impacts. In pursuit of these objectives, herein we investigate salinity variations in the KOCR and 54 its possible impacts through analysis of observational datasets and sensitivity experiments using 55 a simplified model. We found that the upper ocean temperature and salinity in the KOCR exhibit 56 distinct covariation during boreal winter, and they are primarily caused by modulations of ocean 57 circulation in the same region. A detailed decomposition of density anomalies and numerical 58 experiments demonstrated that these salinity anomalies significantly affect the strength of 59 vertical mixing and sea surface temperature, suggesting that salinity has a potential to play an 60 active role in low-frequency variations of the KOCR.

61 (**199 words < 200 words**)

63 **1. Introduction**

64 The upper ocean circulation in the western North Pacific is characterized by two 65 remarkable western boundary currents, namely the Kuroshio and Oyashio, which constitute the 66 subtropical and subarctic gyres, respectively (Fig. 1a) (Qiu, 2002; Yasuda, 2003). Warm and 67 saline subtropical water is transported poleward by the Kuroshio, whereas cold and fresh water 68 from the subpolar region is advected equatorward by the Oyashio. After separating from the 69 coast of Japan, these western boundary currents and associated water masses meet together to 70 form the Kuroshio–Oyashio confluence region (KOCR). Due to the marked differences in the 71 physical properties of these water masses, the hydrographic structures of the KOCR are far from 72 uniform and many complicated but intriguing features, such as sharp sea surface temperature 73 (SST) and salinity (SSS) fronts (Figs. 1b, c), vigorous submesoscale/mesoscale eddies, as well as 74 abrupt changes in vertical stratification (Fig. 1d) are observed (Kida et al., 2015; Roden, 1972; I. 75 Yasuda, 2003; Yuan & Talley, 1996).

76 The oceanic variables (e.g. temperature, salinity, sea surface height (SSH), and current 77 fields) in the KOCR undergo significant variations on various time scales, and interannual to 78 decadal variability in surface and subsurface temperature is especially strong near the 79 climatological fronts (Figs. 1e, f) (Nakamura et al., 1997; Nakamura & Kazmin, 2003; Nonaka et 80 al., 2006). As SST fronts in the KOCR affect the baroclinicity of the lower troposphere and 81 anchor the latitude of the extratropical storm tracks (Nakamura et al., 2004), changes in the 82 KOCR SST not only significantly modulate the local atmospheric conditions but also exert 83 substantial effects on large-scale atmospheric circulation (Frankignoul et al., 2011; Kwon et al., 84 2010; Ma et al., 2015; Smirnov et al., 2015; Taguchi et al., 2012). The KOCR is also known as a 85 key region for water mass formation. Reflecting the intricate hydrographic structures, strong 86 horizontal gradients of the mixed layer depth (MLD) can be seen in the KOCR (Fig. 1d), and

87 they also experience large low-frequency variations as well as other oceanic parameters (Fig. 1f) 88 (Oka et al., 2012; Suga et al., 2004). The inhomogeneous distribution of the MLD in the KOCR 89 is conducive to the formation of mode waters, which are characterized by vertically uniform 90 water properties and believed to play an important role in long-term climate variability and 91 biogeochemical processes. Indeed, various types of mode water are formed in the vicinity of the 92 KOCR, such as the subtropical mode water (Masuzawa, 1969), central mode water (Suga et al., 93 1997), and transition region mode water (Saito et al., 2007), and their variations are closely 94 linked to the variability of the KOCR (see review by Oka & Qiu, 2012). For these reasons, a 95 comprehensive description and understanding of oceanic variations in the KOCR are of 96 particular interests from various perspectives.

97 Thanks to the progress of observational platforms and numerical ocean models in recent 98 decades, significant advances have been made in understanding the driving mechanisms of 99 oceanic variations in the KOCR and other western boundary current regions (Kelly et al., 2010; 100 Kwon et al., 2010). In particular, many studies have attempted to clarify the processes that 101 contribute to the generation of SST anomalies in the KOCR, which is a key variable for air-sea 102 interactions (Pak et al., 2017; Qiu, 2000; Qiu & Kelly, 1993; Vivier et al., 2002). Unlike majority 103 of the extratropical ocean, SST variations in the KOCR are predominantly regulated by ocean 104 dynamical process, rather than being passively forced by the atmosphere (Sugimoto & Hanawa, 105 2011; Tanimoto et al., 2003). The dynamical and thermodynamical processes responsible for 106 these SST anomalies are governed by multiple factors, with both deterministic forcing and 107 intrinsic variability contributing to their variations. On the one hand, upwelling/downwelling 108 Rossby waves excited by the large-scale wind stress curl anomalies in the central to eastern part 109 of the North Pacific alter the intensity of the inertial jet, the latitude of the subtropical-subarctic

110 gyre, and the thermocline depth of the KOCR, giving rise to significant SST anomalies (Kwon & 111 Deser, 2007; Nonaka et al., 2006, 2008; Schneider et al., 2002; Seager et al., 2001). These 112 changes in large-scale ocean circulation also affect the strength of mesoscale eddy activity (Qiu 113 & Chen, 2005, 2010) and associated heat transport (Itoh & Yasuda, 2010; Sasaki & Minobe, 114 2015; Sugimoto et al., 2014). Furthermore, changes in the Ekman transport associated with 115 anomalous wind forcing may also contribute to the generation SST anomalies (Nakamura & 116 Kazmin, 2003; Yasuda & Hanawa, 1997). In addition to such deterministic forcing, internal 117 variability arising from nonlinearities in the western boundary current system (Pierini, 2006; 118 Pierini et al., 2009; Taguchi et al., 2007) also play an important role in the low-frequency 119 variability of the KOCR, particularly on the frontal scale (Nonaka et al., 2012, 2016, 2020; 120 Taguchi et al., 2007). Superimpositions of these two factors (i.e. external forcing and intrinsic 121 variability) and mutual interactions between them control the observed variability in the KOCR 122 (Qiu & Chen, 2010; Taguchi et al., 2007, 2010).

123 Although our knowledge of the upper ocean dynamics and thermodynamics of the KOCR 124 has been considerably enhanced by a large body of previous literature, less is known about 125 salinity, which also exhibits pronounced low-frequency variation (Fig. 1g). This is attributed to 126 the paucity of in-situ salinity observations and difficulty in accurately simulating salinity in 127 numerical ocean models; however, the deployment of Argo profiling floats in the 2000s have 128 rapidly changed this situation. Newly available datasets based on the Argo profiles have enabled 129 the identification of long-term trends and low-frequency variability in the surface and subsurface 130 salinity in the western North Pacific (Geng et al., 2018; Kitamura et al., 2016; Nan et al., 2015; 131 Yan et al., 2013). The governing mechanisms of such salinity variations have also been explored 132 by means of a salinity budget analysis (Geng et al., 2018; Kitamura et al., 2016; Nagano et al.,

133 2014; Sugimoto et al., 2013), but the relative importance of the freshwater flux and advective 134 processes has not been conclusively established in these studies. Such discrepancies could be due 135 to insufficient spatiotemporal resolutions of the observational and reanalysis products used in 136 these studies; therefore, further quantitative assessments based on more comprehensive 137 observations and/or high-resolution ocean models are required. In addition, the possible impacts 138 of these salinity variations on density structures and the evolution of SST are yet to be assessed, 139 although salinity has been shown to play active roles in the tropical climate variability, such as 140 the El Niño-Southern Oscillation (ENSO) (Hasson et al., 2013; Vialard & Delecluse, 1998; Zhu 141 et al., 2015) and the Indian Ocean Dipole (Kido et al., 2019a, 2019b; Kido & Tozuka, 2017; Li et 142 al., 2018; Zhang et al., 2016). Given that seawater density becomes more dependent on salinity 143 than temperature in lower temperature conditions (Gill, 1982), salinity variations in the KOCR 144 have the potential to affect upper ocean processes and related parameters.

145 To address these issues, we investigate the features and mechanisms of salinity variations 146 in the KOCR through an analysis of the observational datasets and an eddy-resolving ocean 147 reanalysis product. In addition, the potential impacts of salinity upon the mixed layer processes 148 are further examined by means of sensitivity experiments using a one-dimensional mixed layer 149 (1-D ML) model. The remainder of this paper is organized as follows. In Section 2, we outline 150 the observational datasets and ocean reanalysis product used in this study. We also briefly 151 describe the 1-D ML model adopted to assess the salinity impacts. The main features of salinity 152 variations in the KOCR and their underlying mechanisms are discussed in Section 3. Then, in 153 Section 4, we assess the impacts on the density structure and SST through a decomposition of the 154 density anomalies and sensitivity experiments using the 1-D ML model. A summary and 155 discussion are presented in Section 5.





Figure 1. Long-term mean climatology of annual mean: (a) sea surface height 157 (SSH: in m); (b) sea surface temperature (SST: in °C) and its meridional gradient 158 (color); (c) sea surface salinity (SSS: in psu) its meridional gradient (color); and (d) 159 mixed layer depth (MLD in m) derived from the Four-Dimensional Variational Ocean 160 Reanalysis for the Western North Pacific over 30 Years (FORA-WNP30). The 161 contour intervals in (a), (b), (c), and (d) are 0.1, 4×10⁻⁶, 6×10⁻⁷, and 10, respectively, 162 and the MLD is defined as the depth at which the potential density increases by 163 0.125 kg m⁻³ from the sea surface. (e)-(h): As in (a)-(d), but with the standard 164 deviation of interannual anomalies (with monthly climatology removed and a 3-165 166 month running mean is applied to anomaly fields). Contour intervals in (e), (f), (g), 167 and (h) are 0.1, 0.3, 0.05, and 10, respectively. 168

169 **2. Data and method**

170 2.1 Observational data

171 In the present study, we analyze the Argo-based gridded temperature and salinity field 172 provided by the Scripps Institution of Oceanography (Roemmich & Gilson, 2009; hereinafter 173 RG09), which has a horizontal resolution of 1° longitude $\times 1^{\circ}$ latitude and 58 vertical levels, 25 174 of which are in the upper 300 m. Monthly data from January 2004 to December 2019 is used in 175 this study. Prior to the analysis, a linear interpolation for 5-m intervals is applied onto the vertical 176 profiles in temperature and salinity. To check the robustness of results, we also adopted the Grid 177 Point Value of the Monthly Objective Analysis (MOAA-GPV) (Hosoda et al., 2008). 178 To complement the limited spatiotemporal coverage of the Argo data and extend the 179 analysis period to 1990s, we use the Four-Dimensional Variational Ocean Reanalysis for the 180 Western North Pacific over 30 Years (FORA-WNP30). The FORA-WNP30 is an eddy-resolving 181 ocean reanalysis product developed by the Meteorological Research Institute of the Japan 182 Meteorological Agency (MRI-JMA) and the Japan Agency for Marine-Earth Science and 183 Technology (JAMSTEC) (Usui et al., 2017). The core ocean model employed for the FORA-184 WNP 30 is the MRI Community Ocean model version 2.4 (Tsujino et al., 2006), which is 185 configured for the western North Pacific (117°E to 160°W, 15°N to 60°N). The model has a 186 spatially varying horizontal resolution, with $1/10^{\circ}$ from 117° E to 160° E (from 15° N to 50° N) 187 and 1/6° from 160°E to 160°W (from 50°N to 60°N) in the zonal (meridional) direction, and 54 188 vertical levels, with increasing grid spacing from 1 m at the surface to 600 m at the bottom (set to 189 6300 m depth). Atmospheric forcing of the model is derived from the JRA-55 atmospheric 190 reanalysis product (Kobayashi et al., 2015) with a daily resolution. The FORA-WNP30 191 assimilates various observational data, such as in situ temperature and salinity profiles (including

192 Argo profiles), gridded sea surface height (SSH), SST, and sea ice concentrations derived from 193 satellites using the four-dimensional variational scheme called the MOVE-4DVAR (Usui et al., 194 2015). In addition to temperature and salinity, we analyze the SSH and horizontal velocity to 195 explore the related physical processes. Surface and subsurface oceanic fields of the FORA-196 WNP30 have been validated against various types of in-situ and satellite observation (Usui et al. 197 2017) and realistically reproduce the observed seasonal features, such as the location of the 198 ocean fronts (Kida et al., 2015) and the wintertime ML distribution (Suga et al., 2004) in the 199 western North Pacific (for example, see their Fig. 12). As in the MOAA-GPV, all three-200 dimensional data were linearly interpolated into 5-m intervals in the vertical direction. The 201 FORA-WNP30 data are available for between January 1982 and December 2014 as a daily 202 average, but we focus on the period from 1991 to 2013, as the surface atmospheric flux product 203 employed in this study (J-OFURO3; for a description, see below) is not available for other 204 periods. We note that the results are qualitatively similar even if we use outputs from the entire 205 period.

In addition to the gridded Argo data and FORA-WNP30, the surface variables (net heat surface fluxes, including shortwave and longwave radiation, sensible and latent heat fluxes, and freshwater flux) from the Japanese Ocean Flux Datasets with Use of Remote Sensing Observation (J-OFURO3) (Tomita et al., 2019) at a horizontal resolution of 0.25° and available from 1991 to 2013 are employed to examine the possible contribution of atmospheric forcing.

211

212 2.2 1-D model experiment

To quantitatively assess the potential impact of salinity anomalies on the mixed layer formation and evolution of SST, we need to explicitly deal with the vertical mixing operating

within the upper ocean. Given the strong horizontal currents and associated large heat and salt transports in this region (Qiu & Kelly, 1993; Vivier et al., 2002), here we adopt a 1-D ML model that can implicitly incorporate advective effects through prescribed forcing (Kido & Tozuka, 2017) and conducted a series of sensitivity experiments. The 1-D ML model employed in this study is a level-2.5 turbulence closure model that was originally formulated by Furuichi et al. (2012). The governing equations for temperature (*T*), salinity (*S*), and horizontal velocity (*u* and *v* represent zonal and meridional velocity, respectively) in the 1-D ML model are as follows:

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(\kappa_T \frac{\partial T}{\partial z} \right) + \frac{1}{\rho_0 C_p} \frac{\partial I}{\partial z} + Res_T(t, z) \tag{1}$$

$$\frac{\partial S}{\partial t} = \frac{\partial}{\partial z} \left(\kappa_S \frac{\partial S}{\partial z} \right) + Res_S(t, z) \tag{2}$$

$$\frac{\partial u}{\partial t} = f(v - v_{geo}) + \frac{\partial}{\partial z} \left(\kappa_V \frac{\partial u}{\partial z} \right) + Res_u(t, z)$$
⁽³⁾

$$\frac{\partial v}{\partial t} = -f(u - u_{geo}) + \frac{\partial}{\partial z}(\kappa_V \frac{\partial v}{\partial z}) + Res_v(t, z)$$
⁽⁴⁾

where ρ_0 (=1023 kg m⁻³) is the reference density, C_p (=3940 J kg⁻¹ K⁻¹) the specific heat 222 223 of the seawater, I the penetrating shortwave radiation, and f the Coriolis parameter. Here, the 224 shortwave penetration was computed by assuming the Jerlov water type IA (Paulson & Simpson, 225 1977). κ_T, κ_s , and κ_V denote the vertical diffusion coefficients of heat, salt, and momentum, 226 respectively, and are internally computed in the model by the turbulence closure scheme, which 227 is primarily based on the local density stratification (calculated from T and S) and vertical shear 228 of horizontal currents at each vertical level (see Furuichi et al. (2012) for details). The major update from Kido & Tozuka (2017) is the implementation of geostrophic velocity (u_{geo} , v_{geo}), 229 which is externally added to the model as the boundary condition. This modification is essential 230 231 for the better simulation of velocity fields in the midlatitude ocean, particularly over regions with swift currents. The last term in each equation, namely $Res_T(t, z)$, $Res_S(t, z)$, $Res_u(t, z)$, and 232

233 $Res_{v}(t, z)$, are referred to as the dynamical correction term, and it represents a contribution from 234 the three-dimensional processes (e.g., horizontal and vertical advection), which cannot be 235 explicitly treated in the 1-D framework. These terms are estimated on the basis of the specified 236 temperature, salinity, and horizontal velocity as described below using technique proposed in 237 Kido & Tozuka (2017). To summarize, the external forcing components necessary for 238 conducting the model experiment are surface atmospheric forcing (heat, freshwater, and 239 momentum fluxes), geostrophic current fields, and three-dimensional oceanic variables used for 240 the initializations and computations of the dynamical corrections. Such implicit incorporations of 241 advective effects into the 1-D ML model have also been adopted in other studies (Vivier et al., 242 2002), and shown to realistically serve as a substitute for the 3-D dynamical processes. For more 243 comprehensive descriptions of our 1-D ML model, interested readers are referred to Furuichi et 244 al. (2012) and Kido & Tozuka (2017).

245 In this study, the 1-D ML is configured at each grid point over a domain covering the 246 western North Pacific region ($135^{\circ}-170^{\circ}E$, $30^{\circ}N-50^{\circ}N$) with 0.5° longitude $\times 0.5^{\circ}$ latitude. The 247 model has a variable vertical resolution, from 2 m near the surface to 10 m at the bottom (the 248 maximum depth is set to 1000 m). Daily shortwave/longwave radiation, 10 m winds, air 249 temperature, specific humidity, and monthly precipitation from J-OFURO3 (Tomita et al., 2019) 250 are used to force the model. Turbulent heat flux, evaporation, and wind stress are calculated 251 using the bulk formulae of Kara et al. (2005). Geostrophic currents are calculated using the 252 density field of the FORA-WNP30, assuming a level of no motion at 2000 m. Initial and 253 restoring conditions (three-dimensional profiles of temperature, salinity, and horizontal current) 254 are then taken from the FORA-WNP30.

255	Using atmospheric and oceanic forcing as outlined above, we first conducted a
256	preliminary experiment to obtain the dynamical correction terms necessary for the realistic
257	simulation of oceanic variability in the KOCR. For this experiment, the model was initialized
258	from October 15 of each year from 1991 to 2012 and then integrated forward for 12 months with
259	atmospheric forcing and geostrophic currents, while restoring the modeled temperature, salinity,
260	and horizontal velocity toward the values derived from the FORA-WNP30 with a nudging time
261	scale of 5 days. During the integration of this experiment, the dynamical correction terms (e.g.,
262	$\frac{T_{FORA}-T}{5 [days]}$ in the case of temperature, where T_{FORA} denotes the temperature value of the FORA-
263	WNP30) were computed at each time step and stored as 1-day averaged values for use in the
264	subsequent experiments. These dynamical corrections were essential for realistically constraining
265	the time evolution of the temperature and horizontal velocity in the subsequent sensitivity
266	experiments. Next, we performed the control (CTL) experiments, for which the model was
267	initialized and integrated with the same atmospheric and oceanic conditions with those in the
268	preliminary experiment, but the nudging of the temperature and horizontal velocity was turned
269	off and the dynamical correction terms obtained from the preliminary integrations were used
270	instead (note that salinity was still relaxed to the FORA-WNP30's value with the nudging time
271	scale of 5 days). As in the preliminary experiment, the CTL experiment was also conducted for
272	all years from 1991 to 2013 in order to simulate the observed variations in the KOCR. As the
273	dynamical correction terms were archived with a high temporal resolution (1-day averaged
274	values were used), the temperature, salinity, and current fields from the CTL experiment were
275	very similar to, or nearly identical, to those from the preliminary experiment (figures not shown).
276	This experiment was used as a reference for comparison to the sensitivity experiment described
277	in Section 4.2.

278 **3. Features of salinity variation in the KOCR**

279 **3.1 Features**

280 First, we examine the time series of temperature and salinity anomalies averaged over the 281 KOCR (142°–153°E, 36°–40°N; see the inset black box in Fig. 1g) obtained from the Argo 282 product and FORA-WNP30 (Figs. 2a, b). Note that this box is chosen to adequately cover the 283 transition region between the Kuroshio Extension (KE) and the Subarctic Front (Kida et al., 284 2015). Over the KOCR, SST and SSS anomalies exhibit coherent interannual to decadal 285 fluctuations, and their time evolution in the Argo product agree well with those in the FORA-286 WNP30. Indeed, the temporal correlation coefficient between the Argo product and FORA-287 WNP30 was 0.67 for SST and 0.62 for SSS, both of which were statistically significant at a 90% 288 confidence level based on the bootstrap method (here, we have generated 10,000 randomly 289 ordered data and estimated the confidence intervals of the correlation coefficients). The spatial 290 patterns of climatological surface and subsurface temperature and salinity over the KOCR also 291 agreed well between the Argo and FORA-WNP30 (figures not shown). For these reasons, we 292 conclude that the FORA-WNP30 adequately reproduced the observed oceanic variability in this 293 region. More thorough validations of the FORA-WNP30 against various observational data are 294 presented by Usui et al. (2017).

To delineate the seasonality of salinity variation, the standard deviation of interannual variability of area-averaged SSS anomalies over the KOCR and their correlation coefficient with SST anomalies over the same region were calculated for each calendar month (Figs. 2c, d). In both the Argo and reanalysis products, the peak of the SSS variation was found to be around the boreal spring, although the FORA-WNP30 slightly underestimated the observed amplitude (Fig. 2c). The coherence between the temperature and salinity was also strong during this season (Fig.

2d), while it was relatively weak during the summer and fall. Considering this seasonality of SSS
variability and annual cycle of MLD, in the analysis that follows, we will primarily focus on the
salinity variation during the late winter to boreal spring (February-April averaged values).



305

306 Figure 2. (a) Time series of 3-month averaged SST and (b) SSS anomalies averaged over the KOCR (142°E-153°E, 36°N-40°N; see black box in Fig. 1g) from the Argo 307 product (black) and FORA-WNP30 (red). Note that anomalies of the Argo product 308 (FORA-WNP30) are relative to their seasonal mean values from 2004 to 2019 309 310 (1991 to 2013). The correlation coefficients between the Argo product and FORA-WNP30 are shown in the lower left. (c) Standard deviation of the SSS anomalies 311 312 averaged over the KOCR as a function of calendar months from the Argo product 313 (orange) and FORA-WNP30 (blue). (d) As in (c), but for correlation coefficients

between SST and SSS anomalies over the KOCR. The unit for SST (SSS) is °C
(psu).

316

317 As can be seen in the time series of SST and SSS anomalies (Figs. 2a, b) and their 318 correlation (Fig. 2d), anomalous saltening (freshening) in the KOCR tends to co-occur with 319 warming (cooling) therein. To objectively detect such anomalous events and extract features common to all cases, we will define positive and negative years as follows: a positive (negative) 320 321 year is defined as one with both February to April averaged SST and SSS anomalies over the 322 KOCR that are larger (smaller) than their 0.5 standard deviation. According to this criterion, 6 323 (8) positive years and 4 (6) negative years can be identified in the Argo data (FORA-WNP30) 324 (Table 1).

325

	Positive years	Negative years
Argo (RG09)	2007, 2008, 2009,	2011, 2012, 2014,
	2013, 2016, 2019 (6	2015 (4 events)
	events)	
Reanalysis (FORA-	1997,1999, 2000,	1996, 2003, 2004,
WNP30)	2002, 2007, 2008, 2009,	2005, 2010, 2011 (6
	2013	events)
	(8 events)	

326 Table 1. Positive and negative years identified in the Argo data (RG09) and

327 reanalysis product (FORA-WNP30).

329 Composites of the temperature and salinity field during the mature phase of positive and 330 negative years (averaged for February-April) obtained from the Argo product are shown in Fig. 3. 331 We note here that the features in the composites with January–March mean are qualitatively 332 similar to the following composites. During positive years, warm and saline water from the 333 subtropical region extends farther to the north compared to negative ones (Figs. 3a, b, d, e). As 334 expected from the definition of events, regions with significant differences in SST tend to be 335 collocated with those in SSS (Figs. 3c, f). Similar patterns are also found in composites from the 336 FORA-WNP30 (Fig. 4), although small-scale features are more evident and the amplitude of 337 SSS anomalies is slightly smaller than that in the Argo data. As the gross features of composite 338 fields constructed from the Argo data of the overlapping period (2004-2013) were very similar to 339 the original ones (figures not shown), quantitative differences between the Argo and FORA-340 WNP30 may be caused by those in their horizontal resolution and analysis method, rather than 341 the analysis period. Such anomalous warming events in the KOCR have also been noted in 342 several previous studies (Masunaga et al., 2016; Qiu et al., 2017; Sugimoto et al., 2014); 343 however, these studies have not specifically focused on associated salinity variations. These 344 studies have pointed out that the low-frequency fluctuations of SST over the KOCR are related 345 to the modulation of dynamic states of the KE. To determine whether such arguments also apply 346 to our selected positive/negative years, we will explore the origin of these temperature and 347 salinity variations in the next subsection.

348



Figure 3. (a)-(c): Composite of SST fields (in °C) during February-April of (a) positive and (b) negative years from the Argo data. Differences between positive and negative years (i.e. (a) minus (b)) are shown in (c). The contour intervals in (a) and (b) are 2, whereas those in (c) are 0.4. Differences that are significant at the 80% confidence levels based on a two-tailed t-test are green dotted in (c). (d)-(f): As in (a)(c), but for SSS fields (in psu). The contour intervals in (d) and (e) are 0.2, whereas those in (f) are 0.05.



- 359 Figure 4. As in Fig. 3, but from the FORA-WNP30.
- 360

361 To examine the driving mechanisms of temperature and salinity variations over the 362 KOCR, it is helpful to emphasize their vertical structure. For this purpose, latitude-depth sections 363 of zonally averaged $(142^{\circ}-153^{\circ}E)$ composited temperature, salinity, and potential density from 364 the Argo data are depicted in Fig. 5. For both positive and negative years, prominent density-365 compensating temperature and salinity fronts are seen around 36°-39°N and they extend to the 366 upper 200 m (Figs. 5a, b, d, e). Comparison of the temperature and salinity fields between the 367 positive and negative years reveals that strong warming and saltening is observed in the KOCR 368 during the positive years, while the opposite is observed in the negative years (Figs. 5c, f). Large 369 differences in temperature and salinity are found near the surface to the north of 38°N (i.e., the 370 subarctic region), whereas they are found near the thermocline depth (from 200 to 400 m depth)

to the south (Figs. 5c, f). Such latitudinal differences in the vertical structures of low-frequency
thermohaline anomalies have also been noted by Nonaka et al. (2006).

373 Interestingly, the differences in potential density are characterized by meridional dipole 374 structures (Fig. 5i) with negative anomalies (i.e., a decrease in density) to the south and positive 375 anomalies to the north. The causes of these density anomalies will be discussed in the next 376 section by decomposing these anomalies into respective contributions from temperature and 377 salinity. Similar patterns of temperature, salinity, and potential density anomalies are also 378 observed in composited fields obtained from the FORA-WNP30, although anomalies over the 379 northern part are slightly underestimated (Fig. 6). Hence, we believe that the differences in the 380 upper ocean fields between the positive and negative years are robust features across the datasets 381 and analysis periods. In the next subsection, we will explore the governing mechanisms of these 382 events and possible links to large-scale variability by inspecting the features of other variables.



Composite of temperature, salinity, and density from Argo



- 392 fields (in kg m⁻³). The contour intervals are 0.2 in (g) and (h), whereas those in (i)
- 393 are 0.05.



Figure 6. (a)-(i): As in Fig. 5, but from the FORA-WNP30. (j)-(l): As in (a)-(c), but for the zonal current fields (in m s⁻¹). The contour intervals in (j) and (k) are 0.1 and 0.05 in (l).

399 3.2 Mechanisms

400 There are several candidates that induce co-variations in temperature and salinity 401 variations over the KOCR. First, changes in the local atmospheric conditions, such as anomalous 402 heat and freshwater exchanges at the sea surface can directly generate in-phase and/or out-of-403 phase variations in SST and SSS. Second, an anomalous strengthening or weakening of large-404 scale ocean circulation (both the Ekman and geostrophic current) may modulate temperature and 405 salinity advection, thereby creating significant anomalies. Third, a modulation in the strength of 406 the mesoscale eddy activities affects the magnitude of the eddy-induced transport of heat and 407 freshwater transport, thereby leading to significant temperature and salinity anomalies. To 408 identify relative contribution from these factors and underlying physical processes, composites of 409 various physical parameters are presented in Figure 7. Because the growing season of 410 temperature and salinity anomalies in the KOCR is a few months prior to the mature phase (the 411 lead–lag relationship between the SSS anomalies and SSS tendency anomalies is shown in Fig. 412 8), we focus on differences averaged from December to February. 413 Due to the outbreak of cold and dry air masses from the continent by the westerly wind, 414 the net surface heat flux is mostly upward (i.e., the ocean releases heat into the atmosphere) over 415 the western North Pacific (Figs. 7a, b). Differences between the positive and negative years (Fig. 416 7c) show that the ocean is more strongly cooled by the atmosphere during the positive years in 417 the KOCR. A detailed decomposition of the heat flux anomalies into individual components (i.e., 418 shortwave and longwave radiation as well as sensible and latent heat fluxes) reveals that warmer

419 SST during the positive years (Fig. 3c) and associated increases in the turbulent heat fluxes are

420 responsible for the total differences (figures not shown). Therefore, the heat flux anomalies serve

421 to dampen the SST anomalies and do not contribute to their generation and growth. These results

422 are in line with those from previous observational studies, which underlined the importance of
423 SST anomalies over the Kuroshio and Oyashio extension regions in driving heat flux variability
424 there (Masunaga et al., 2016; Sugimoto & Hanawa, 2011; Tanimoto et al., 2003).

425 Consistent with the heat flux fields and their interpretation described above, more (less) 426 freshwater is lost to the atmosphere over the KOCR during the positive (negative) years (Figs. 427 7d-f). This is conducive to surface saltening (freshening) in the positive (negative) years and 428 could contribute to the generation of the observed SSS anomalies. Unsurprisingly, these 429 differences in freshwater fluxes are primarily due to changes in evaporation, while no significant 430 differences were found in the precipitation fields. The maximum amplitude of freshwater flux differences is approximately 2×10^{-8} m s⁻¹, which leads to changes in the mixed layer salinity of 431 432 0.02 psu per month, assuming the mixed layer depth of 100 m. This value is rather smaller than 433 the observed total SSS differences (~0.2 psu) and could explain only one-third of the anomaly in 434 three months, suggesting that other processes, such as anomalous salinity advection, may also be 435 important for the generation of salinity variations.

To highlight the roles played by oceanic processes, we next compare low-passed sea surface height, surface current, and eddy kinetic energy (EKE) fields between positive and negative years (Figs. 7g–1). Here, the low-frequency (high-frequency) signals were obtained by applying a 300-day (Qiu & Chen, 2005) Lanczos low-pass (high-pass) filter (Duchon, 1979) to the total fields.

Differences in the SSH fields between the positive and negative years are characterized by a meridional dipole over the KE, with the higher (lower) SSH anomalies to the north (south) of the climatological eastward jet (Fig. 7i). More specifically, the signatures of the southern and northern recirculation gyres (Qiu et al., 2008) are markedly discernable and the KE jet is zonally

445 oriented during negative years (Fig. 7h), whereas the KE jet is weaker and convoluted during 446 positive ones (Fig. 7g). These features are suggestive of their link with the bimodal states of the 447 KE (Qiu & Chen, 2005, 2010; Taguchi et al., 2010); a positive (negative) year with weaker 448 recirculation gyres corresponding to an unstable (stable) state of the KE, as speculated in several 449 previous studies (Masunaga et al., 2016; Qiu et al., 2017; Sugimoto et al., 2014). The differences 450 in velocity fields (figures not shown) are nearly in geostrophic balance with those in SSH, 451 suggesting that current anomalies mostly come from the geostrophic components rather than the 452 Ekman current. During positive years, northeastward currents over the KOCR are more 453 prominent than negative years, especially in the southwestern area $(36^{\circ}N-38^{\circ}N)$; see also Figs. 454 6g-l). The stronger (weaker) northeastward current during positive (negative) years leads to an 455 increase (a decrease) in the advection of warm and salty water from the south, contributing to an 456 anomalous warming and saltening (cooling and freshening) over the KOCR. Such a modulation 457 of advective processes efficiently operates in the climatological frontal regions with strong 458 temperature and salinity gradients and may partly explain the differences in the peak depth of 459 anomalies between the northern and southern areas of the KOCR (Nonaka et al., 2006). The 460 region with significant current anomalies collocates with upstream portions of the quasi-461 stationary jet (QSJ, also referred to as the J-1) (Isoguchi et al., 2006; Wagawa et al., 2014), 462 which is a conveyor of warm and saline water from the KE to subarctic regions. Thus, these 463 current anomalies may be viewed as a modulation of the QSJ associated with changes in the 464 dynamical state of the KE, as suggested in an observational study by Wagawa et al. (2014). 465 In relation to changes in large-scale ocean circulation, the strength of mesoscale eddy 466 activity undergoes significant variations due to changes in barotropic/baroclinic instability as 467 well as those in the interaction with bottom topography (Qiu & Chen, 2005, 2010; Yang et al.,

2017). With respect to the KE, the EKE in the upstream regions substantially decreases when it
is in a stable state; conversely, the EKE increases when the KE is in an unstable state. (Itoh &
Yasuda, 2010; Qiu & Chen, 2005, 2010; Sasaki & Minobe, 2015; Sugimoto et al., 2014; Taguchi
et al., 2010). To confirm consistency with these previous findings, we calculated the EKE as
follows:

$$EKE = \frac{1}{2} (u'^2 + v'^2),$$
5)

473 where (u', v') denotes high-passed horizontal velocity.

474 During positive years, an elevated EKE level is seen during positive years compared to 475 negative years (Figs. 7j–l), supporting our argument that the positive (negative) years correspond 476 to an unstable (stable) and high eddy-activity state of the KE. As individual mesoscale eddies 477 serve to relax the meridional gradients of temperature and salinity through Lagrangian transports 478 of heat and salt (Dong et al., 2017; Itoh & Yasuda, 2010), the increase (decrease) in numbers and 479 strength of mesoscale eddies during the positive (negative) years is conducive for warming and 480 saltening (cooling and freshening) over the KOCR (Qiu et al., 2017; Sasaki & Minobe, 2015; 481 Sugimoto et al., 2014), as well as anomalous large-scale ocean circulation.



Figure 7: (a)-(c): Composite of net surface heat flux (in W m⁻²) during December-February of (a) positive and (b) negative years, and (c) their differences from the J-OFURO3. The contour intervals are 50. Here, positive values indicate heating of the ocean by the atmosphere. Differences that are significant at the 80% confidence levels based on a two-tailed t-test are green dotted in (c). (d)-(f): As in (a)-(c), but for net surface freshwater flux (evaporation minus precipitation) (in m

490 s⁻¹). The contour intervals are 1 x 10⁻⁸. (g)-(h): As in (a)-(c), but low-passed sea 491 surface height (in m). The contour intervals are 0.1. (j)-(l): As in (a)-(c), but for the 492 surface eddy kinetic energy (EKE: in m² s⁻²). The contour intervals are 0.03. 493

494 To confirm the importance of the various processes described above, lead-lag correlation 495 coefficients between February-April mean SSS anomalies over the KOCR and other variables 496 are presented in Fig. 8. As mentioned in Section 3.1 (Fig. 2d), SSS anomalies are highly 497 correlated with SST anomalies over the same region, with its maximum value around lag 0 (Fig. 498 8, black curve). The SSS tendency (i.e., the time derivative of SSS) has a significant positive 499 correlation with SSS anomalies when the former lead the latter by around 2 to 6 months, 500 suggesting that the maximum growth of SSS anomalies occurs a few months before their peak 501 season (grey curve). The freshwater flux anomalies (note that they are defined as evaporation 502 minus precipitation, so that the positive values correspond to increases in SSS) also exhibit 503 significant correlations, but their coefficients are relatively low (~ 0.4), when they lead the SSS 504 anomalies (blue curve). Similar features with a reversed sign were also found for the net heat 505 flux anomalies (figure not shown). Therefore, the local atmospheric anomalies are not the main 506 driver of the observed temperature and salinity variations, although they may play a secondary 507 role in determining their amplitude.

508 For ocean dynamical variables, both zonal and meridional low-pass filtered velocity 509 (orange and green curves) and EKE anomalies are significantly correlated with SSS with leading 510 SSS anomalies, supporting the idea that the temperature and salinity variations over the KOCR 511 are closely linked to the dynamical state of the KE. As the ocean background circulations and 512 mesoscale eddy fields mutually affect each other via eddy-mean flow interaction (Qiu & Chen,

513 2010; Taguchi et al., 2010), it is not straightforward to clearly isolate individual contributions 514 from both factors. Therefore, herein we qualitatively conclude that changes in the stability of the 515 KE path leads to coherent changes in the upper ocean temperature and salinity over the KOCR 516 through modulation of heat and salt transport by large-scale geostrophic current and mesoscale 517 eddies, whereas contributions from freshwater flux and Ekman advection anomalies seemed to 518 be not so important. A comparison of advective terms estimated from the current fields and 519 salinity of the FORA-WNP30 also corroborated the above conclusion (Fig. S1). For a more 520 quantitative and comprehensive assessments of these processes, a closed salinity budget analysis 521 based on a realistic high-resolution ocean model is desirable and could be an interesting topic for 522 future studies.

523



Figure 8. Lead-lag correlation coefficients between February-April averaged SSS 525 anomalies over the KOCR and other variables (black: SST; grey: SSS tendency; 526 red: EKE; blue: freshwater flux; green: low-passed surface zonal velocity; and 527 528 orange: low-passed surface meridional velocity). All oceanic variables (SSS, SST, 529 EKE, surface velocity) are taken from the FORA-WNP30 reanalysis, whereas the J-OFURO3 product is adopted for the freshwater flux. The lag is in units of the 530 month, and positive (negative) values indicate SSS anomaly leads (lags). Correlation 531 coefficients that are significant at the 90% confidence levels on the basis of the 532 bootstrap method are represented by the colored dots. 533

4. Impact of salinity variation

In this section, we assess how these salinity variations can alter the upper ocean's
hydrographic properties and eventually affect the evolution of the SST, which is a key variable
for midlatitude air–sea interaction.

538

539 4.1 Temperature and salinity contributions to density anomalies

540 We calculate the potential density of seawater (denoted as $\rho(T, S)$) using original (i.e. 541 interannually varying) temperature and salinity based on the equation of state by Jackett 542 &Mcdougall (1995). Then, we compute the potential density using original temperature and climatological salinity (represented by \overline{S}), $\rho_T = \rho(T, \overline{S})$, to isolate the effect of salinity variations. 543 544 Differences in potential density between the positive and negative years, $\Delta \rho = \rho_{POS}(T, S) - \rho_{POS}(T, S)$ $\rho_{NEG}(T, S)$, contain both contributions from temperature and salinity differences. For ρ_T , 545 $\Delta \rho_T = \rho_{POS}(T, \bar{S}) - \rho_{NEG}(T, \bar{S})$ are caused only by the temperature difference. Thus, the 546 547 contribution from salinity variations to potential density (= $\Delta \rho_s$) can be estimated by considering the differences between $\Delta \rho$ and $\Delta \rho_T$, i.e. $\Delta \rho_S = \Delta \rho - \Delta \rho_T = \{\rho_{POS}(T,S) - \rho_{POS}(T,\bar{S})\} - \rho_{POS}(T,\bar{S})\}$ 548 $\{\rho_{NEG}(T,S) - \rho_{NEG}(T,\bar{S})\}$). Although the nonlinearity of the equation of state does not allow a 549 550 complete separation of the density signals into temperature and salinity contributions, this 551 method is useful for illustrating the importance of salinity variations. Differences in other 552 density-dependent variables, such as the buoyancy frequency and mixed layer (see below for 553 detailed definitions) can, in the same manner, be decomposed into temperature and salinity 554 contributions. This approach has been widely used for assessments of salinity impacts over the 555 tropical Pacific (Zheng & Zhang, 2012, 2015) and tropical Indian Ocean (Kido & Tozuka, 2017).

556	The positive temperature and salinity anomalies in the KOCR in the positive years (Figs.
557	9a, b; see also Figs. 3c, f) serve compensate each other in forming anomalous densities, with
558	positive temperature anomalies leading to a decrease in sea surface density (SSD) (Fig. 9c),
559	whereas positive salinity anomalies contribute to increases in the SSD there (Fig. 9d). As a result,
560	negative values of the total SSD differences are confined to the southern part of the KOCR
561	(south of 38°N; Fig. 9e). Similar features are also evident in composites from the FORA-WNP30,
562	although surface saltening and associated compensations of the temperature-related SSD signals
563	are weaker than those in the Argo data (Figs. 10a–e).



Figure 9. (a), (b): Differences in composited (a) SST (in °C) and (b) SSS (in psu)
fields between positive and negative year during February-April. The contour
intervals in (a) are 0.4, whereas those in (b) are 0.05. (c)-(e) As in (a) and (b), but
for (e) shows the surface density differences and contribution from (c) SST and
(d) SSS differences (in kg m⁻³). The contour intervals are 0.08 (see the main text

570	for details of the decomposition method). (f): Sea surface density differences
571	estimated from SST differences, assuming a uniform thermal expansion
572	coefficient (i.e. (a) multiplied by –0.22 kg $^{\circ}C^{-1}$ m ⁻³) (in kg m ⁻³). The contour intervals
573	are 0.08. (g), as in (f), but from SSS differences assuming uniform saline
574	contraction coefficients (i.e., (b) multiplied by 0.77 kg $psu^{-1} m^{-3}$). The sum of (f)
575	and (g) is shown in (h). (i)-(k): As in (c)-(e), but for the mixed layer depth
576	differences (in m). The contour intervals are 20. Differences that are significant
577	at the 80% confidence levels on the basis of a two-tailed t-test are green dotted
578	(except for (f)-(h)). All panels are from the Argo data.



Composite differences of SST, SSS, SSD, and MLD from FORA-WNP30

- Figure 10. As in Fig. 9, but from the FORA-WNP30. 581
- 582

580

583 A noticeable feature in the composite field is the fact the spatial pattern of SST's 584 contribution to SSD (Figs. 9c) is somewhat different from that of the original SST differences 585 (Figs. 9a), and such discrepancies are not found in the SSS fields (Figs. 9b, d) (see also Fig. 10 for the FORA-WNP30). Specifically, SST's contribution to SSD is relatively weak in higher 586

latitudes compared to the distribution of the original SST differences. This could be due to lower
background SST and weaker dependence of density on temperature at higher latitudes. To
confirm this argument, we converted SST and SSS anomalies into SSD using a linear relation.
Assuming the linearity of the equation of state, the differences of potential density can be
approximated as follows:

$$\Delta \rho = \Delta \rho_T + \Delta \rho_S \simeq \alpha (T_{POS} - T_{NEG}) + \beta (S_{POS} - S_{NEG}) = -\alpha \Delta T + \beta \Delta S_A$$

592 where α and β represent thermal expansion and saline contraction coefficients, respectively. The 593 value of α significantly increases with temperature and β is almost uniform within the parameter 594 range of the upper ocean in the KOCR (Gill, 1982; Jing et al., 2019). By converting SST and 595 SSS differences into SSD using spatially constant α and β , we can estimate the impact of the background SST distribution (here we choose $\alpha = 0.22$ kg °C⁻¹ m⁻³ and $\beta = 0.77$ kg psu⁻¹ m⁻³). 596 597 Figs. 9f-h show the SSD anomalies estimated through a linear relationship between temperature, 598 salinity, and potential density. The negative values of the SSD differences under the linear 599 relation (Fig. 9h) extend more poleward than those of the actual SSD differences (Fig. 9e), 600 primarily due to larger contributions from SST (Figs. 9c, f). Similar patterns but weaker 601 magnitudes are also evident in composites from the FORA-WNP30, again confirming the 602 importance of nonlinearity in generating SSD anomalies. We note that a recent study 603 demonstrated that differences in the thermohaline properties of mesoscale eddies in the KE and 604 those in the Oyashio region can also be explained by the meridional contrasts in the thermal 605 expansion coefficients associated with the front of background SST (Jing et al., 2019). 606 Changes in SSD can alter the density stratification, hence affecting the MLD, which is an 607 important parameter for controlling the effective heat capacity of the upper ocean. To assess the

608 impacts of temperature and salinity variations on the MLD, we also decompose the MLD

differences between the positive and negative years using the method described above. Here, the
MLD is defined as a depth at which the density increases by 0.125 kg m⁻³ over the SSD.

611 The total differences in MLD in the KOCR are characterized by a complex spatial pattern 612 with meridionally alternating positive and negative anomalies (Fig. 9k). During the positive year, 613 a significant ML shoaling is observed around 38–40°N and 32–35°N, whereas deepening of ML 614 is observed to north of 40°N. Distinct MLD variations associated with changes in the dynamical 615 state of the KE are also pointed out by Oka et al. (2012); they have found that deepening of 616 wintertime ML is observed around 31–35°N and 40–42°N during a stable state of KE (see their 617 Fig. 4). Given that a positive (negative) year generally corresponds to an unstable (a stable) state 618 of KE, our results are fairly consistent with findings of Oka et al, (2012), although some 619 discrepancies are found in the details of their spatial patterns, arguably due to differences in data 620 period and processing methods.

621 A decomposition of these differences demonstrates that the ML shoaling in the KOCR 622 during positive years is limited to the south of 40°N because of the salinity effects (Fig. 9j). This 623 can be explained by an anomalous increase in SSD near the surface associated with positive SSS 624 anomalies there (Figs. 9b, d). Meanwhile, contributions from temperature anomalies dominate 625 those of salinity anomalies to the south (Fig. 9i). Again, the qualitatively same features were 626 found in composites from the FORA-WNP30 (Figs. 10i-k), but the ML shoaling in the southern 627 KOCR was also caused by salinity anomalies (around 36°–38°N). Such MLD changes cannot be 628 simply explained by corresponding SSD anomalies, implying that subsurface salinity anomalies 629 also play an important role in determining the distribution of MLD.

To highlight the vertical structure of the temperature and salinity contributions to density anomalies, we constructed depth-latitude sections of the decomposed anomalies (Figs. 11 and

632 12). The anomalous high temperature and salinity near the climatological thermohaline fronts
633 (Figs. 11a, b) have their peak near the surface (thermocline depth) in the northern (southern) part
634 of the KOCR. These positive temperature and salinity anomalies generate offsetting density
635 perturbations (Figs. 11c, d).

636 Due to the larger thermal expansion coefficients for warmer water, the contributions from 637 temperature dominate the total density field to the south of 38°N, but they become comparable 638 $(\sim 80\%)$ of temperature's contribution; figure not shown) to those of salinity to the north (Fig. 639 11e). The effects of these anomalies on the density stratification can be inferred from the composite of squared buoyancy frequency $N^2(z) = \frac{-g}{\rho_0} \frac{\partial \rho}{\partial z}$ (Figs. 11f–h). Here, we have 640 computed $N^{2}(z)$ from the density field using the central difference scheme with a uniform 641 vertical grid. For the poleward side of the KOCR (north of 38°N), positive differences in $N^2(z)$ 642 643 (i.e., strengthening of the stratification) due to anomalous surface warming (Fig. 11f) were 644 significantly compensated by concomitant increase in salinity and potential density near the 645 surface (Fig. 11g). For the southern side, by contrast, salinity effects are not so large and total 646 differences largely reflect contributions from temperature (Figs. 11f, h). A similar meridional 647 contrast of hydrographic structures can also be seen in the composites from the FORA-WNP30 648 (Fig. 12), although subsurface differences around the KE latitude ($\sim 36^{\circ} - 38^{\circ}$ N) are more 649 prominent compared to the Argo data. Both in the Argo and FORA-WNP30 products, salinity 650 variations in the KOCR have nonnegligible contributions to the density perturbations compared 651 to temperature variations, and they become comparable to those of temperature to the north of 652 38°N as the background temperature decreases poleward. Therefore, these salinity variations 653 have the potential to significantly affect the strength of the vertical mixing and evolution of

- temperature, which will be carefully quantified in the next section using the 1-D ML model
- 655 experiments.











Composite differences of temperature, salinity, density, and N²(z) from FORA-WNP30



670 **4.2 Quantitative assessment using 1-D model**

671 Before proceeding to sensitivity experiments of the 1-D ML model, we first check the 672 performance of our control (CTL) experiment, which is a reference for the sensitivity 673 experiments. In the CTL experiment, the model is driven by atmospheric forcing from the J-674 OFURO3, geostrophic current fields from the FORA-WNP30, and temperature and momentum 675 dynamical correction terms derived from the preliminary experiment, whereas the modeled 676 salinity is strongly nudged toward the values of the FORA-WNP30, as detailed in Section 2.2. 677 Owing to the implementation of the dynamical correction methods, the model aptly captures the 678 spatial pattern of the climatological SST and SSS fields (Figs. 13a-f), although the modeled SST 679 is slightly warmer (cooler) to the north (south) of 40°N compared to that of the FORA-WNP30. 680 Here, we have shown the wintertime mean state as a representative period with strong SST/SSS 681 fronts; however, the climatology of other seasons, such as the March–May averaged field, is also 682 simulated well by the 1-D model. In addition, the climatology of the subsurface temperature and 683 salinity fields, as well as the horizontal currents, is also in good agreement with the reanalysis 684 product (figures not shown). This suggests that our 1-D model can adequately simulate the 685 background oceanic conditions over the western North Pacific. Furthermore, the time evolutions 686 of area-averaged SST anomalies over the KOCR from the FORA-WNP30 and the 1-D model are 687 also compared well (Fig. 13g). Similarly, the SSS variation in the 1-D model also corresponds 688 well with that in the reanalysis product (Fig. 13h), as was expected from the adoption of salinity 689 nudging during the CTL experiment. These conspicuous agreements between the 1-D ML model 690 and the reanalysis product allow us to make further use of it for a more detailed investigation. 691



700	FORA-WNP30 (black) and 1-D ML model (red). The correlation coefficient between
701	them is shown in the lower right. (h) As in (g), but for SSS anomalies.

703 While both temperature and salinity variability are dominated by the same 3-D 704 mechanisms as discussed above, here we intend to clarify how and to what extent salinity has 705 potential to modify dynamical and thermodynamical processes in the upper ocean. Motivated by 706 the fact that many previous studies on the KOCR have not explicitly considered the salinity 707 effects mainly due to limited salinity observations, we explore this issue with an artificial 708 experiment with the 1-D model. To explicitly depict the role played by salinity variability, we 709 treat salinity as a "forcing" in the 1-D ML model and see "responses" of other related variables, 710 such as the vertical diffusion coefficients and temperature. Based on this concept, we have 711 designed another set of experiments that nullify the salinity's roles by artificially suppressing its 712 fluctuation (referred to as the climatological salinity (Sclim) experiment). In this experiment, we 713 initialize and force the model as in the CTL experiment, except that salinity used for the initial 714 and restoring conditions was replaced by corresponding climatological values. With sufficiently 715 strong relaxation, salinity variations (except for the seasonal cycle) and associated changes in 716 density stratification and related processes are eliminated. As the same temperature and 717 momentum dynamical correction were used in both the CTL and Sclim experiments, the 718 collective impacts of salinity anomalies on the vertical mixing process and associated changes in 719 temperature can be adequately measured by considering difference between the CTL and Sclim 720 experiments (Kido & Tozuka, 2017). This framework provides useful insights regarding the 721 significance of salinity effects, although it has an inevitable limitation due to its absence of three-722 dimensional responses (e.g., possible changes in temperature advection associated with salinity-

induced current anomalies are not included). Because oceanic anomalies during negative years
are close to a mirror image of those during positive years, we only conducted the Sclim
experiment for the seven positive years appeared in the FORA-WNP30 (see Table 1).

726 The observed anomalous surface warming and saltening over the KOCR during the late 727 winter to spring of the positive years (Figs. 14a, b) were well reproduced in the CTL experiment 728 (Figs. 14c, d), further confirming its satisfactory ability to simulate the observed variability. To 729 isolate the effects of salinity anomalies, we calculated the differences in the SST and SSS fields 730 between the CTL and Sclim experiments for all selected positive years and their composites, as 731 shown in Figs. 14e, f. The spatial pattern of the SSS differences between the two experiments 732 (Fig. 14f) closely resembles the composite SSS anomalies (Fig. 14d; see also Figs. 3f and 4f), 733 suggesting that the targeted SSS anomalies over the KOCR were successfully removed in the 734 Sclim experiment. The SST differences were characterized by negative (positive) values over the 735 northern (southern) part of the KOCR, suggesting that the inclusion of salinity anomalies during 736 positive years led to cooling (warming) in those parts (Fig. 14e). This implies that salinity 737 variations during the positive years serve to dampen (amplify) the concomitant SST warming 738 over the northern (southern) part of the KOCR and hence, inhibit the poleward intrusion of 739 anomalous warming. The areas with cooler (warmer) SST in the CTL than the Sclim experiment 740 roughly coincide with the regions where salinity anomalies contribute to the weakening 741 (strengthening) of density stratification and deepening (shoaling) of the mixed layer, indicating 742 that changes in the vertical process may hold the key. These differences are commonly seen in 743 all positive years and the maximum differences in SST between the two experiments reaches 744 1.0°C, which constitutes 20%–40% of the total SST anomalies there (Figs. 14c, e). Therefore,

- salinity variations in the KOCR can exert significant effects on the evolution of SST therein by
- 746 modulating the 1-D vertical process.



749	Figure 14. (a), (b): Composite of (a) SST (in $^\circ C$) and (b) SSS (in psu) anomalies
750	during February-April during positive years from the FORA-WNP30. (c), (d): As in
751	(a) and (b), but from the CTL experiment of the 1-D ML model. (e), (f): Composite
752	of the difference in SST (e) and SSS (f) between the CTL and Sclim experiments
753	of the 1-D ML model during March-May. The green dotted regions indicate
754	differences that are significant at the 80% confidence levels on the basis of a
755	two-tailed t test. The purple and orange boxes in (e) denote the northern and
756	southern box, respectively.

758 What causes such distinct SST differences between the two experiments? Given the 759 configurations of our 1-D ML model experiments, there are two possible explanations for these 760 differences. First, changes in the MLD due to salinity anomalies (cf. Figs. 9i-k and 10i-k) can 761 alter the effective heat capacity of the upper ocean and affect the sensitivity of SST to 762 atmospheric heat flux. Second, changes in vertical stratification due to salinity anomalies may 763 modulate the strength of vertical mixing and turbulent heat transport in the upper ocean. To 764 assess the first hypothesis, we carried out a detailed mixed layer heat budget analysis based on 765 the output from the 1-D ML model (see the supplementary material for details), and it was found 766 that the MLD changes due to salinity anomalies have the opposite effect. These results indicate 767 that the SST differences between the two experiments cannot be simply explained by those in the 768 atmospheric heat flux or MLD, implying that modulations in vertical mixing and associated heat 769 transport hold the key.

To confirm the above statement and further illuminate the related physical processes, we next check the time evolution of other mixing-related parameters, such as the density stratification and vertical mixing coefficient from each experiment. The time-depth plots of the area-averaged composited temperature, salinity, squared buoyancy frequency ($N^2(z)$), and vertical diffusion coefficient of temperature (κ_T : see Eq. 1) from both 1-D ML model experiments are shown in Figs. 15 (the northern box) and 16 (the southern box).

776 For the northern box, the seasonal cycle of temperature and salinity variation, such as the 777 gradual deepening of the mixed layer in winter and rapid shoaling during spring, are reproduced 778 in both experiments, (Figs. 15a, b, d, e). Differences between the CTL and Sclim experiments 779 demonstrate that the positive salinity anomalies near the surface begin to develop in winter, peak 780 in spring, and subsequently decay in summer (Fig. 15f). These salinity anomalies serve to 781 weaken upper ocean stratification at 100–150 m depth (Fig. 15i) and then lead to the 782 strengthening of the vertical mixing there (Fig. 151). As a result, the vertical heat exchange 783 between the surface and subsurface layer during late winter to early spring is greatly enhanced, 784 giving rise to cooler SST and a warmer subsurface temperature in the CTL experiment (Fig. 15c), 785 supporting the hypothesis proposed above. We again note that temperature differences between 786 the CTL and Sclim experiments are caused only by changes in the vertical diffusion because 787 both experiments adopt the same amount of dynamical corrections and shortwave radiation. Thus, 788 the chain of physical processes described above is adequately represented in our experimental 789 framework. The maximum SST differences were found during April-May and then subducted 790 below the seasonal thermocline during summer and fall. Differences in the subsurface 791 temperature (i.e., warmer temperatures in the CTL experiment) at 150-200 m depth also persist 792 through summer and remain until fall, even though no salinity signals survive until this season.

Therefore, the salinity-induced temperature perturbations can persist longer than the salinity
variations themselves and hence have the potential to affect the low-frequency variation of upper
ocean.

796 The 1-D ML model also satisfactorily reproduces the key features of temperature and 797 salinity variation within the southern box (Figs. 16a-f), as in the northern box. Significant 798 positive salinity signals are also evident in the difference between the two experiments, but their 799 maximum peak is found at 100–150 m depth rather than near the surface (Figs. 16d–f). 800 Consequently, the density stratification is strengthened (Fig. 16i) and vertical mixing near the 801 thermocline is substantially more suppressed in the CTL experiment than in the Sclim 802 experiment (Fig. 161). Therefore, the vertical entrainment of subsurface cold water is 803 significantly reduced, and eventually leads to a warmer SST (and slightly lower thermocline 804 temperature) in the CTL experiment.

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808 Figure 15. (a)-(c): Time evolution of composited temperature in (a) CTL and (b)



- 810 minus the Sclim experiments) averaged over the northern box (142°E-151°E, 38°N-
- 811 40°N: See the purple box in Fig. 14e). The contour intervals in (a) and (b) are 1,
- 812 whereas those in (c) are 0.1. The differences that are significant at the 80%

- 813 confidence levels on the basis of a two-tailed t-test are green dotted in (c). (d)-
- 814 (f): As in (a)-(c), but for salinity (in psu). The contour intervals in (d) and (e) are
- 815 0.05, whereas those in (f) are 0.03. (g)-(i): As in (a)-(c), but for the squared
- buoyancy frequency (in s^{-2}). The contour intervals are 3× 10⁻⁵. (j)-(l): As in (a)-(c),
- 817 but for the vertical diffusion coefficients (in m² s⁻¹). The contour intervals are
- 818 **5×10**⁻².
- 819



Figure 16. As in Fig. 15, but for the southern box (142°E-151°E, 35°N-37°N: See

822 the purple box in Fig. 14e).

5. Summary and discussion

824 Using observational datasets and an eddy-resolving ocean reanalysis product (FORA-825 WNP30), in this study, we investigated low-frequency variations of upper ocean salinity in the 826 KOCR and examined their mechanism and possible effects on the mixed-layer processes, with a 827 specific focus on variability during the boreal winter to spring. From a gridded dataset based on 828 the Argo profiles and an eddy-resolving ocean reanalysis product, we identified a coherent 829 interannual to decadal variation in temperature and salinity in the KOCR, with anomalous 830 saltening (freshening) tending to be accompanied by significant warming (cooling) across the 831 same region. Based on the area-averaged SST and SSS anomalies in the KOCR, we selected 832 several typical years for such events, and a positive (negative) year is defined as one with a 833 significant increase (decrease) in both SST and SSS in the area. A close inspection of the three-834 dimensional structures of composite temperature and salinity anomalies reveals that such 835 anomalies are concentrated near the surface in the northern part of the KOCR, but strong 836 anomalies are found at 200–400 m depth in the southern part of the KOCR. Such meridional 837 differences in the vertical structures of thermohaline anomalies reflect the distribution of the 838 climatological temperature and salinity fronts, which was also pointed out by earlier works 839 (Nakamura & Kazmin, 2003; Nonaka et al., 2006).

The mechanisms of these salinity variations were then explored based on the basis of a composite and lag correlation analysis of the related physical variables. We found that the dynamical stability of the KE was the predominant factor behind the observed variations. During positive years, accompanying an unstable state of the KE, an increase in the SSH and anomalous anticyclonic circulation could be observed in the northern part of the KE. Further, associated northeastward current anomalies directing toward the KOCR enhanced the poleward transport of

846 warm and saline water originating from the subtropics, leading to significant surface warming 847 and saltening in the KOCR. At the same time, the intensification of mesoscale eddy activity and 848 eddy-induced advection in the upstream of the KE also contribute to the generation of positive 849 temperature and salinity anomalies in the KOCR. The increase in surface evaporation due to 850 positive SST anomalies also serves to maintain the positive SSS in the KOCR, but its 851 contribution is relatively small compared to the ocean dynamical effects mentioned above. 852 Similar anomalies, but with opposite polarities (i.e., features with a stable state of the KE) also 853 contribute to anomalous cooling and freshening during negative years.

854 To quantify the effects of these salinity variations on the density of seawater and vertical 855 stratification, we decomposed the density anomalies into contributions from temperature and 856 salinity anomalies. During the positive years, positive SSS anomalies in the northern part of the 857 KOCR lead to an increase in surface density and serve to deepen the mixed layer in that region. 858 Similarly, the positive subsurface salinity anomalies in the southern part of the KOCR enhance 859 the vertical stability and contribute to the shoaling of the mixed layer by increasing the density at 860 that depth. These salinity-induced density perturbations compensate for the concomitant 861 temperature-induced density perturbations and significantly reduce the amplitude of total 862 anomalies. Salinity contributions to density anomalies increase poleward and become 863 comparable to those of temperature in the northern part of the KOCR, and this can be explained 864 by the weaker dependence of density on temperature due to lower background temperatures in 865 that region. These results suggest that salinity variations in the KOCR have substantial impacts 866 on the local density fields and may exert considerable effects on dynamical and 867 thermodynamical processes.

868 Based on the results from the density decomposition, we have assessed the impacts of the 869 density changes associated with salinity anomalies upon the strength of vertical mixing and 870 evolution of the upper ocean temperature by carefully designing and conducting a series of 871 sensitivity experiments using the 1-D ML model. These sensitivity experiments demonstrated 872 that salinity anomalies during positive years serve to cool SST in the northern part of the KOCR 873 by up to -1.0° C, whereas in the southern part, they cause SST warming of the same amplitude. 874 By analyzing other variables from the 1-D ML model, it was found that changes in density 875 stratification due to salinity anomalies indeed modulate the strength of vertical mixing and 876 induce significant responses in the upper ocean temperature. More specifically, positive SSS 877 anomalies in the northern part of the KOCR reduce the density stratification and strengthen the 878 vertical mixing in that region, resulting in significant near surface cooling and subsurface 879 warming in that region. In the southern part, by contrast, positive subsurface salinity anomalies 880 during positive years stabilize the upper ocean column and suppress the vertical exchange of heat 881 within the mixed layer, leading to near surface warming and (weaker) subsurface cooling. Thus, 882 surface and subsurface salinity anomalies in the KOCR suppress the poleward expansions of co-883 occurring temperature anomalies.

The close linkage between the dynamical states of the KE and SST variations over the KOCR has been documented in several previous studies (Masunaga et al., 2016; Qiu et al., 2017; Sasaki & Minobe, 2015; Sugimoto et al., 2014). However, little has been discovered concerning the concomitant salinity variations and associated mechanisms. In this regard, we have shown, primarily based on a lagged correlation analysis, that ocean dynamical processes, especially the modulation of heat and salt transport by large-scale circulation and mesoscale eddies, are closely related to the wintertime temperature and salinity variations in the KOCR. Although these

conclusions are physically consistent and in accordance with many previous studies, they are still
based on statistical relationships and more physical approaches are required to confirm their
validity. An accurate salinity budget analysis as well as coordinated sensitivity experiments
using high-resolution ocean general circulation models (OGCMs) would be helpful in addressing
these issues.

896 In this study, we demonstrated that salinity has the potential to play an active role in low-897 frequency variations in the KOCR by modulating the upper ocean temperature via density 898 change. These results have important implications for the study of climate variability in the 899 North Pacific, as the SST variability in the KOCR affects atmospheric circulation, as discussed 900 in Section 1 (Frankignoul et al., 2011; Taguchi et al., 2012). Due to the strong internal variability 901 of the atmosphere and ocean, how and to what extent these atmospheric responses to SST 902 anomalies feed back onto the ocean is still a matter for debate, but salinity may be involved in 903 such feedback processes, provided that it exerts strong impacts on SST. An important caveat of 904 this study is that our estimates of salinity impacts on SST based on the 1-D ML model disregard 905 changes in three-dimensional advective processes produced by salinity anomalies. As salinity 906 anomalies may also alter circulation in the upper ocean and the associated transport of heat and 907 momentum, well-designed OGCM and data assimilation experiments are necessary to 908 incorporate and assess the significance of such effects. Finally, strong salinity fronts are also 909 found in other WBCs such as the Gulf Stream, Agulhas Current, and Antarctic Circumpolar 910 Current (Kida et al., 2015; Ohishi et al., 2019), and similar salinity variations may also be 911 evident in these regions. The applications of our approach to other WBCs and comparisons to the 912 KOCR results will provide further insight into the physical processes operating in the WBCs and 913 their roles in midlatitude climate variability.

914 Acknowledgments

915 We would like to thank Tomoki Tozuka, Eitarou Oka, and Shota Katsura for their 916 insightful comments and suggestions, and Norihisa Usui for his technical advice on the use of 917 FORA- Constructive comments from two anonymous reviewers have significantly improved an 918 early version of the manuscript. Data used were obtained as follows; the gridded Argo product 919 the Institution from Scripps of Oceanography is from http://sio-920 argo.ucsd.edu/RG Climatology.html; the FORA-WNP30 data is from 921 http://synthesis.jamstec.go.jp/FORA/e/index.html; and the J-OFURO3 surface flux data is from 922 https://j-ofuro.isee.nagoya-u.ac.jp/en/. The present study is supported by KAKENHI 923 JP19H05701, JP19H05702, JP18H03726, and 21K13997.

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