Influence of the Kuroshio Large Meander on the Climate around Japan
Based on a Regional Climate Model

Kazuyo MURAZAKI, Hiroyuki TSUJINO, Tatsuo MOTOI, and Kazuo KURIHARA

Meteorological Research Institute, Japan Meteorological Agency, Japan
Meteorological College, Japan Meteorological Agency, Japan

(Manuscript received 3 October 2013, in final form 20 November 2014)

Abstract

We performed a 20-year numerical experiment over the period 1985 to 2004 using a high-resolution North Pacific Ocean General Circulation Model (NPOGCM) and a 20 km-resolution regional climate model (RCM20) to clarify the impact of the Kuroshio large meander (LM) on the climate around Japan. The NPOGCM reproduced the two primary quasi-stationary states, straight path (SP), and large meander (LM), although the periods during which each state prevailed differed from those indicated in the observational data. The NPOGCM result also showed that the Kuroshio LM causes a cold sea surface temperature anomaly to the south of the Pacific coast of the central Japan. Using the result as a lower boundary condition, a continuous numerical integration was performed by the RCM20. An 8-year composite analysis of the atmospheric circulations of the RCM20 simulation for the Kuroshio LM and SP showed that, in both winter and summer, substantial decreases in the upward surface turbulent heat flux, the frequency of precipitation, and the frequency of steep horizontal gradients in equivalent potential temperature over the ocean are caused by the cold sea surface temperature anomaly. Similar effects are evident over the land area of Japan, although they are less intense, at most 20–50 % of magnitude over the cold sea surface temperature anomaly area, and limited to the coastal region on the Pacific Ocean side in the central part of the country.

Keywords Kuroshio large meander; air-sea interaction; regional climate

1. Introduction

The demand for more detailed regional climate information has increased in recent years because of the considerable effect of regional climate on water resources, the agricultural sector, flood control, renewable energy production, and other human activities. Several studies have used regional climate models (RCMs) to examine the factors affecting regional climate (Eastman et al. 2001; Liang et al. 2005; Yoshimura and Kanamitsu 2009).

Kuroshio is a warm western boundary current that transfers substantial amounts of heat from the western tropical Pacific to the mid-latitude region of the western North Pacific. It flows east-northeastward along the Pacific Ocean side of the Japanese islands and then moves off the Japanese coast to the east at around 35°N, forming the eastward jet known as the Kuroshio Extension. Numerous observational and numerical simulation studies have shown that the Kuroshio has an important impact on the atmosphere in this region. For example, it causes changes in atmospheric structure and cloud formation (Tanimoto et al. 2009; Tokinaga et al. 2009). Wind speed is related to sea surface temperature (SST) anomalies (Nonaka and Xie 2003) and decadal variability in the Kuroshio Extension region (Qiu 2003; Nonaka et al. 2006).

Kuroshio is known for its meandering behavior; it sometimes deviates from its straight path (SP) along
the coast of Japan, and meanders to the southeast, away from the coast. This flow pattern is known as the Kuroshio large meander (LM). Various researchers have studied the mechanism of Kuroshio flow variation (Kawabe 1995; Tsujino et al. 2006; Usui et al. 2008), while the research on the effects of the Kuroshio LM on atmospheric conditions is still ongoing. Xu et al. (2010) examined the effect of the Kuroshio LM on atmospheric parameters during the meander events of 2004–2005, using satellite observations and a numerical model. They performed a month-long simulation for January in 2005 and examined the atmospheric response to the SST cold pool caused by the Kuroshio LM. Nakamura et al. (2012) examined the behavior of cyclone tracks using an observational 40-year wintertime sea level pressure (SLP) dataset. They suggested that cyclone tracks shift southward and that cyclone development rates decrease during Kuroshio LM periods. Hayasaki et al. (2013) used a high-resolution coupled general circulation model (CGCM) to investigate cyclone activity in winter over the Kuroshio and found that their results agreed well with the analysis of observational data reported by Nakamura et al. (2012).

However, thus far, research has focused primarily on winter, and the effects of the Kuroshio LM on climate throughout the year have not been studied well. One of the reasons for this is that the air-sea heat flux is relatively weak in the warmer seasons. It is therefore more difficult to detect the atmospheric impact of the Kuroshio LM, except in winter.

This study aims to quantify the impact of the Kuroshio LM on the climate around Japan in winter and summer. To represent the narrow structure of the Kuroshio and its variation in terms of oceanic and atmospheric characteristics, it was necessary to use a spatial resolution of 0.2° or higher because the width of the current and the curvature radius of the Kuroshio LM is approximately 1°. The RCM used by Murazaki et al. (2005) and Sasaki et al. (2006) performs better when used with high resolution SST than when used with low-resolution SST. Therefore, we used an updated version of the high-resolution North Pacific Ocean General Circulation Model (NPOGCMv.2) and a 20-km resolution RCM (RCM20) developed by the Meteorological Research Institute (MRI) of Japan. A long-term integration driven by JRA-25 atmospheric forcing was conducted using NPOGCMv.2 that can resolve the Kuroshio and the accompanying SST change. A 20-year integration with the RCM20 was performed using the SST simulated by NPOGCMv.2 as a lower boundary condition and JRA-25 as a lateral boundary condition. In this study, the SST of an ocean model simulation was used because a physically consistent field in the upper ocean, which comprises SST, subsurface water temperature, and oceanic current, can be obtained with high horizontal resolution for the entire JRA25 period without missing value. To the best of our knowledge, such a dataset based only on observations is currently unavailable.

To detect detailed atmospheric responses to the variation in the Kuroshio path, we compared these results with those of another long-term integration simulation performed by Murazaki et al. (2010), which used the RCM20 with observational low-resolution SST data as the lower boundary condition.

This paper is organized as follows. The model settings and the simulation design are explained in Section 2. Simulation results are presented in Section 3. The causes of the larger variation in precipitation rates in summer than in winter and the wind-affecting mechanism of the Kuroshio LM are discussed in Section 4. Finally, summary and conclusions are presented in Section 5.

2. Model and simulation design

NPOGCMv.2 (ocean model) and RCM20 (regional atmospheric model) models were used in this study. NPOGCMv.2 (hereafter, NPOGCM) is an updated version of the model used by Sato et al. (2006), and it is based on the Meteorological Research Institute Community Ocean Model (MRI.COM), which was described by Ishikawa et al. (2005) and evaluated by Ishizaki and Motoi (1999) and Tsujino and Yasuda (2004). The model domain covers the area from 15°S to 65°N and from 100°E to 75°W with 34 vertical layers and a horizontal resolution of 1/4° in longitude and 1/6° in latitude. Around Japan (25–50°N, 120–160°E), the resolution is 1/10° in both the longitude and latitude. The model includes a bulk formula for calculating surface flux developed by Kondo (1975), a vertical mixing scheme based on Mellor and Yamada’s (1982) level 2.5 scheme (Mellor and Blumberg 2004), and a horizontal mixing scheme with biharmonic operators. High-accuracy advection schemes are used for the tracers, UTOPIA (Leonard et al. 1993) for the horizontal and QUICKEST (Leonard 1979) for the vertical direction, as well as for momentum (Ishizaki and Motoi 1999). In addition to the main mixing scheme, two kinds of additional mixing are imposed. First, to reproduce the low SST in summer along the Kuril Islands (Nakamura et al. 2004), a tidal mixing parameterization based on St. Laurent et al. (2002) is used. Second, to reduce the

...
strong fluctuation of the Kuroshio at the southern coast of Japan, a harmonic viscosity is added for momentum dissipation over steep bottom topography, with a viscosity coefficient of $2.5 \times 10^2 \cos \varphi \text{ m}^2 \text{ s}^{-1}$, where $\varphi$ is the latitude.

RCM20 is based on a regional spectral model originally developed by the Japan Meteorological Agency (JMA) as a short-range forecasting model (NPD/JMA 1997). The model resolution used in this study is approximately 20 km with 36 hybrid vertical levels. The model includes the Mellor and Yamada (1982) level-2 scheme for vertical diffusion, the Arakawa and Shubert (1974) cumulus convection and convection adjustment schemes, a short-wave radiation scheme (Lacis and Hansen 1974), a long-wave radiation scheme (Sugi et al. 1990), and a ground surface process (Takayabu et al. 2004). A spectral boundary coupling method proposed by Kida et al. (1991) was used for the long-term simulations. In this method, the large-scale results of the outer model and the small-scale results of the inner nested model are combined in wave number space. This has the advantage of a smooth long-term integration, avoiding contradictions between an outer coarse-mesh model and a nested model with respect to large-scale fields. A detailed description of RCM20 can be found in Sasaki et al. (2000).

We used the Japan 25-year reanalysis (JRA-25) dataset as the atmospheric forcing data for driving the NPOGCM and as a lateral boundary condition for the RCM20. This dataset consists of reanalysis data from 1979–2004 at 6-hourly time intervals. It was produced by the JMA and the Central Research Institute of the Electric Power Industry (CRIEPI) (Onogi et al. 2007). The resolution for the atmospheric data was 1.25° × 1.25°. The JRA-25 dataset also includes the 1.125° resolution SST dataset (hereafter, JRA-SST) based on central in situ observation-based estimates (COBE) of SST produced by the JMA (Ishii et al. 2005; Onogi et al. 2007).

Initially, for the simulation using NPOGCM, 26-year-monthly-averaged datasets of surface wind, surface air temperature (SAT), radiation, precipitation, and SLP were generated by JRA-25 as forcing data. The 20-year integration was then performed as a spin-up. Using the spin-up results as an initial condition and the 6-hourly JRA-25 atmospheric data as the forcing data, the NPOGCM simulation was run from 1 January 1979 to 31 December 2004. The last 20 years of the resulting SST and sea ice data were used as lower boundary conditions for RCM20. Initial and lateral boundary conditions were provided by the 6-hourly data from JRA-25. A 20-year simulation using RCM20 was run from 1 January 1985 to 31 December 2004 (hereafter, OGCM run).

It is difficult to identify atmospheric responses to variations in SST using the OGCM run results alone because several factors influence the atmosphere such as the monsoon, global warming, and the westerly jet. Therefore, we used the long-term integration results from RCM20 (JRA run; Murazaki et al. 2010), using the JRA-SST as the lower boundary condition, as a control run. The only difference between the OGCM and JRA runs was the lower boundary conditions. The JRA run used an observation-based SST with a resolution of 1.125°, whereas the OGCM run used the NPOGCM high-resolution simulation result with a resolution of 1/10° around Japan. We compared these two runs to detect detailed atmospheric responses to variations in the Kuroshio’s path.

To characterize atmospheric responses to the Kuroshio LM more clearly, we investigated the seasonally averaged field of surface turbulent heat flux and equivalent potential temperature (EPT). We also examined transient events such as the frequency of steep gradients in EPT and high-intensity rainfall events.

The EPT of a volume of air is greater when its temperature is high or the air contains a large amount of water vapor. Heavy rain is closely related to EPT at the 850 hPa level at the time of heavy rainfall is not suitable because water vapor from the sea is stored in a lower convective mixed layer up to ~1 km deep over the sea. He compared EPT at the 500 m level, SLP, and wind vectors in heavy rainfall areas, and concluded that inflow from the south of an air mass with high EPT at the 500 m level plays an important role in the generation and development of cumulus clouds. Because RCM20 does not predict conditions at the 500 m level, we used predicted EPT at the 950 hPa level instead as air pressure at 500 m is generally close to 950 hPa (Kato 2009).

We also investigated horizontal and vertical EPT gradients over the ocean around the Kuroshio LM. The vertical EPT gradient serves as an index of atmospheric stability, which is closely related to atmospheric convection. The horizontal EPT gradient increases where two different air masses meet. Weather fronts are expressed more clearly by the horizontal EPT gradient than the temperature gradient, especially in the rainy season. The occurrence of a
strong EPT gradient indicates that active front zone exists there (Ninomiya and Akiyama 1992; Zhou et al.
2004).

3. Results

3.1 Reproducibility of the Kuroshio LM

We compared the 20-year averages of SST for winter (December–February) and summer (June–August) from JRA-25 and NPOGCM with the fine resolution (0.25° × 0.25°) merged satellite and in situ data global daily SST (MGDSST) data (Kurihara et al. 2006) (Fig. 1), and found that JRA-SST does not sufficiently resolve the detailed features of the ocean currents (Figs. 1a, d). MGDSST effectively showed the fine-scale SST distribution associated with the Kuroshio flow along the Pacific coast of Japan (Figs. 1b, e). NPOGCM showed the transport of warm water to the southern coast of Japan by the Kuroshio in detail (rectangle in Figs. 1c, f), however, this model also had some biases. During winter, the 22 °C isotherm predicted by NPOGCM reached the Kii Peninsula (33°N 136°E; Fig. 1c), whereas those indicated by the MGDSST and JRA-SST data lay south of 33°N (Figs. 1a, b). In summer, the 28 °C isotherm predicted by NPOGCM (Fig. 1f) was north of those indicated by the MGDSST and JRA-SST data (Figs. 1d, e).

Next, the reproducibility of variations in the Kuroshio’s paths is demonstrated in Fig. 2. In this study, we used the southernmost latitude of the Kuroshio front to identify the presence of the Kuroshio LM. According to the definition given by Kawai (1972), the Kuroshio LM is expected to occur when the southernmost latitude of the Kuroshio front is stably located south of 32°N, i.e., between 136°E and 139°E. The position of the Kuroshio front is approximated by a water temperature of 15 °C at a depth of 200 m (Kawai 1972). On the sea surface, weather significantly affects water temperature. We therefore used subsurface temperature at a depth of 200 m, rather than SST, to identify the Kuroshio LM. It is well-known in the oceanographic community that this method gives results similar to those given by other methods such as the report on tidal level differences between two points on both the sides of Cape Shiono-Misaki at the southern tip of the Kii Peninsula (JMA 2007).

With respect to simulating the timing of the Kuroshio LM, the model results indicate that the current was in a highly stable SP phase from 1987–1996, and subsequently moved to a prevailing LM phase (Fig. 2). However, the JMA observational data (JMA 2007) show that the Kuroshio was in the SP state for most of the study period, except for three relatively short LM phases: December 1986 to July 1988, December 1989 to December 1990, and July 2004 to August 2005. Therefore, the model did not successfully predict the timing of LM phases. Tsujino et al. (2013) suggested that the net depth-integrated Kuroshio transport off southern Japan, whose mean quantity and variability are affected by the basin-scale wind stress field, strongly influences the path selection of the Kuroshio. Specifically, a weak (strong) Kuroshio favored LM (SP). In the NPOGCM simulation, the net Kuroshio transport was about 40 Sv in the SP period (1989–1996) whereas it was 20 Sv in the LM period (1997–2004) (not shown). This is consistent with the suggestion of Tsujino et al. (2013), even though some positive feedback effects should be taken into account. On the basis of the observations, Imawaki et al. (2001) estimated the net Kuroshio transport to be 42 Sv as a seven-year average in the period from 1992 to 1999, during which the Kuroshio was largely in the SP state. Thus, the net Kuroshio transport of NPOGCM was consistent with the observational estimate in the SP state. The net Kuroshio transport should have been larger in the prevailing LM phase in the NPOGCM simulation, although further investigations are required to confirm this conjecture. However, Tsujino et al. (2013) also pointed out that mesoscale eddies are involved in the transition of the Kuroshio path, implying that it would be generally still difficult for the current OGCMs to get the Kuroshio LM timing right. In this study, we will focus on the atmospheric responses to the LM that operate through heat and water vapor exchanges, and will not pursue the mechanism causing the timing of the Kuroshio LM.

The SST distribution patterns characteristic of the two states of the Kuroshio are compared in Fig. 3. When the system switches from the SP to the LM state, SST decreases in the area where the Kuroshio dominates in its SP state, and increases in the area covered by the LM path. The effects of the variation between the two states were not obvious in the JRA-SST data (Figs. 3a, d, g). The MGDSST data, however, showed a striking decrease in SST around 33°N 138°E and an increase around 35°N 142°E (Figs. 3b, e, h). The NPOGCM SST predictions also showed the effects of the variation in state, but suggested that the Kuroshio extended further south than indicated by the MGDSST data (Figs. 3c, f, i). As a result, the cold pool caused by the Kuroshio LM was relatively overestimated by NPOGCM (Fig. 3i).

The Kuroshio LM state simulated by NPOGCM
Fig. 1. Horizontal distributions of 20-year-averaged SST from the JRA-25 (a, d), from the MGDSST (b, e), and simulated by NPOGCM (c, f). Distributions are for winter (December–February; a, b, c) and summer (June–August; d, e, f). The isotherm interval is 1°C. Rectangles in (c) and (f) indicate the area where warm northward SST flows occur, which are associated with the Kuroshio.
was overestimated in terms of both area and duration (Figs. 2, 3). However, because the main purpose of this study is to quantify the impact of the Kuroshio LM on the climate around Japan, the NPOGCM run does not necessarily need to accurately replicate the Kuroshio LM period. The simulation may be considered successful if it reproduces the two characteristic paths of the Kuroshio. The exaggerated area of the LM is preferable because it will cause RCM20 to predict stronger atmospheric responses to the LM than the MGDSST data would. The stronger atmospheric response can then be taken as an upper bound of possible atmospheric responses to the Kuroshio LM. In fact, the LM did occupy a considerably larger area in the early 1960s and late 1970s (Kawabe 1995). Furthermore, because variations in the MGDSST data may be dampened by the application of optimum interpolations, these data might be insufficient to resolve the detailed features of intense ocean currents such as the Kuroshio. The SST predicted by NPOGCM, however, is at least based on the physical law of the model. Therefore, we used the SST fields simulated by NPOGCM when running RCM20.

### 3.2 Atmospheric response to Kuroshio LM using RCM20

#### a. Variations in the frequency of steep gradients in equivalent potential temperature and rainfall intensity

With respect to the difference in total surface turbulent heat flux (latent plus sensible heat flux, defined positive upward, hereafter, surface heat flux) in winter between the LM and SP states predicted by OGCM, there was a region with negative values (cold pool), reaching minima of approximately −80 W m\(^{-2}\) between 33\(^\circ\)N 134\(^\circ\)E and 34\(^\circ\)N 141\(^\circ\)E. This is the northern side of the Kuroshio front during LM phases (Fig. 4a). Regions with positive values (warm pools), reaching a maxima of approximately 60 W m\(^{-2}\), were spread out along the upstream Kuroshio extension front region between 34\(^\circ\)N 140\(^\circ\)E and 36\(^\circ\)N 146\(^\circ\)E. In the JRA run, the area of moderate positive values of less than 32 W m\(^{-2}\) was spread out widely over the ocean (Fig. 4c).

Surface heat flux in summer is relatively weak, and the variation in surface heat flux between the LM and SP states was correspondingly weak with minima of approximately −46 W m\(^{-2}\) in the cold pool (Fig. 4b). This pattern corresponds well with the pattern of variation in SST (Fig. 3i). In the JRA run (Figs. 4c, d), differences were even smaller (less than 16 W m\(^{-2}\)) and were almost uniformly spread over the ocean. There was no statistically significant difference (p < 0.1) in surface heat flux variations in the simulation area from SP to LM states between the OGCM and the JRA run, with the exception of the areas surrounding the Kuroshio LM (not shown). Therefore, the SST variations caused by the Kuroshio LM in the OGCM run are likely to be responsible for the differences in surface heat flux.

Variations in surface heat flux from the ocean to the atmosphere affect both temperature and water vapor in the atmospheric boundary layer. In the OGCM run in winter, EPT in the cold pool was strongly reduced during the LM state relative to the SP state (Fig. 5a), however, the JRA run showed only a weak negative area north of 32\(^\circ\)N and a weak positive area south of 32\(^\circ\)N (Fig. 5c). For the summer months, a weak negative area appeared near 33\(^\circ\)N 140\(^\circ\)E in the OGCM run (Fig. 5b). Given the fact that EPT in this area increases in the JRA run as a result of natural climate variability (Fig. 5d), the variation of EPT due to the Kuroshio LM probably shows a similar pattern in both winter and summer when the natural variability is subtracted.

The variation in surface heat flux caused by the Kuroshio LM affects the variation in both EPT distri-
bution and the horizontal EPT gradient related to atmospheric front activity. The LM causes the winter frequency of horizontal EPT gradients exceeding \(0.1\) K km\(^{-1}\) (hereafter, steep \(\nabla \theta_e\)) at the 950 hPa level to decrease over an area stretching from the northern edge of the cold pool to the southern edge of the warm pool (33°N 134°E to 34°N 147°E; Fig. 6a). The frequency of steep \(\nabla \theta_e\) increases in an area that spreads out from the southern edge of the cold pool. These variations are more pronounced in summer than in winter (Fig. 6b). The larger variations in summer may be related to the fact that the frequency of steep \(\nabla \theta_e\) in summer is more than double of that in winter. For example, at a location of large variation in the surface heat flux (33.6°N 138°E), the seasonal average of steep \(\nabla \theta_e\) frequency based on 3-hourly
Fig. 4. Variations in seasonally averaged total surface turbulent heat flux (W m$^{-2}$) between the Kuroshio LM and SP states (LM minus SP), as simulated by the OGCM run in (a) winter (December–February) and (b) summer (June–August), and by the JRA run in (c) winter and (d) summer. The line from A to B in Fig. 4a indicates the cross-section illustrated in Figs. 7 and 9.

Fig. 5. Variations in seasonally averaged EPT (K) between the Kuroshio LM and SP states (LM minus SP), as simulated by the OGCM run in (a) winter (December–February) and (b) summer (June–August), and by the JRA run in (c) winter and (d) summer.
Sampling is about 60 (46) out of 240 data times per month for the SP (LM) state in summer. In winter, these figures are about 23 and 16 per month for the SP and LM state, respectively. In the JRA run, however, variation in the frequency of steep $\nabla \theta$ was relatively small (Figs. 6c, d). These distributions were similar to the results of the OGCM run (Figs. 6a, b), except for the areas surrounding the Kuroshio LM.

To analyze the influence of the Kuroshio LM on the atmosphere in more details, we examined the rate of variation in frequency of steep $\nabla \theta$ predicted by both the OGCM and JRA run along the line 33.6°N, where variation in surface heat flux was the greatest (Fig. 4a).

The rates of variation in the frequency of steep $\nabla \theta$ predicted by the two runs diverged near

![Fig. 6. Variations in seasonally averaged frequency of horizontal EPT gradients ($\nabla \theta$) that exceed 0.1 K km$^{-1}$ (times month$^{-1}$) between the Kuroshio LM and SP states (LM minus SP), as simulated by the OGCM run in (a) winter (December–February) and (b) summer (June–August), and by the JRA run in (c) winter and (d) summer.](image)

![Fig. 7. Rate of variation, calculated as (LM–SP)/SP and expressed as a percentage, in the frequency of horizontal EPT gradients ($\nabla \theta$) that exceed 0.1 K km$^{-1}$ in (a) winter and (b) summer, as simulated by the JRA run (dashed line) and the OGCM run (bold solid line), along 33.6°N (line A–B in Fig. 4a).](image)
134°E and converged near 146°E in winter, whereas they diverged near 136°E and converged near 145°E in summer (Fig. 7). The difference between the two runs was approximately 35% in winter and 30% in summer. Frequency decreased in a relatively large area that extended eastward from the cold pool. The following may constitute a possible explanation.

In general, the meridional EPT distribution makes a substantial contribution to \( \nabla \theta \). When the Kuroshio is in the LM state, \( \nabla \theta \) over the northern edge of the cold pool, around 33°N and 134–140°E, decreases because the lower EPT in the cold pool reduces the differences in EPT between the north and south. Moreover, it decreases in the extended eastward zone around 34°N and 140–146°E because the increase in the EPT over the warm pool weakens the contrast between this air and the humid air from the south.

The variation in the frequency of high intensity rainfall events, defined as more than 1 mm of precipitation within three hours, is shown in Fig. 8. The JRA run predicted a weak decrease in the cold pool area (Fig. 8c). The OGCM run, however, predicted variation corresponding to the variations in surface heat flux during winter months (Fig. 5a), i.e., the frequency decreased in the cold pool but increased in the warm pool (Fig. 8a). The maximum frequency variation was about 8 times per month. In summer, the pattern was similar to winter (Fig. 8b), with the addition of a region of strong decline in the frequency of high intensity rainfall events over the much-weakened cold pool. As shown in Figs. 4 and 6, there were few differences between the OGCM run and the JRA run outside the area surrounding the Kuroshio LM.

Figure 9 shows the rate of variation in the frequency of high intensity rainfall events, as predicted by the OGCM run and the JRA run, along the line A–B shown in Fig. 4a. The rates of variation predicted by the two simulations began to diverge near 135.5°E and again converged near 143°E, which corresponds almost exactly with the cold pool area. The difference between the two runs was approximately 30% for summer and 20% for winter.

Fig. 8. Variations in the seasonally averaged frequencies of high intensity rainfall events (times month\(^{-1}\)) between the Kuroshio LM and SP phases (LM minus SP), as predicted by the OGCM run for (a) winter (December–February) and (b) summer (June–August), and by the JRA run for (c) winter and (d) summer.
b. The influence of the Kuroshio LM on the Japanese climate

As discussed above, differences between the predictions of the OGCM run and the JRA run were most striking around the cold pool area and the Pacific Ocean side of central Japan, but were hardly apparent in other regions of Japan (Figs. 5, 6, 8). To obtain a quantitative estimate of the influence of variations in the Kuroshio’s path on Japan’s climate, we focused on three land areas along the Pacific side of Japan (Fig. 10) and compared the averages of four climate parameters with those from the cold pool area. To reduce the effects of natural variations unrelated to variations in the path (e.g., global warming and decadal-scale climate variations), we examined the deviation of the predictions of the OGCM run from those of the JRA run. For this analysis, area A was defined as the region where the Kuroshio turns southward during LM phases, area B as the region where the Kuroshio’s path is the furthest from land, and area C as the region near the open ocean where the Kuroshio turns north-eastward in both the LM and SP phases. Model run duration was divided into 4 seasons and results were compared between seasons.

With respect to seasonal variation in area-averaged SAT between SP and LM phases, area A showed a statistically significant (p < 0.1, all seasons) reduction in SAT of approximately 0.1–0.2 °C (Fig. 11a), which is 20–30 % of the reduction over the oceanic cold pool area. There was also a small but significant (p < 0.1) reduction of approximately 0.1 °C in spring (March, April and May) and summer (June, July and August) in area B, but no statistically significant differences in area C were observed.

Air temperature at the 850 hPa level exhibited a similar pattern of cooling in all seasons (Fig. 11b). In area A, the variation was smaller than that of SAT in every season, but in areas B and C the degree of variation in temperature at the 850 hPa level was relatively small.

Area-averaged SLP showed positive variations in the range of 0.03–0.07 hPa in all the seasons (Fig. 11c), which were 30–50 % smaller than those in the
cold pool. The most dramatic variations were observed in spring, similar to air temperature, and the degree of variation was the greatest in area A. In area C, there were no significant variations except in spring.

The area-averaged seasonal mean frequency of high intensity rainfall events decreased by 1–2 times per month in spring and autumn during LM phases (Fig. 11d). With respect to SAT and SLP, in most cases the variations caused by the Kuroshio LM were the largest in area A, which is the nearest to the cold SST pool. In contrast, the frequency of high intensity rainfall events showed larger variations in area C or B in every season. The variation over the oceanic cold pool was two or three times larger than that over the three land areas. In summer, the variation over the oceanic cold pool was substantial; however, despite this no significant variations occurred in the land areas.

4. Discussion

4.1 The mechanism causing the variation in intense rainfall frequency in summer

Surface heat flux variation was greater in winter than in summer (Fig. 4); however, the variation in rainfall frequency was greater in summer than in winter. Here, we investigate the cause from the perspective of the vertical profiles of EPT.

Figure 12 depicts the vertical profiles of the difference in temperature and EPT between the LM and the SP states averaged over a rectangular area from 32°N to 34°N and from 138°E to 140°E (the cold pool area; see also Fig. 10). In comparison with the JRA run, which serves as a reference for natural variation, the
OGCM run predicted that temperature near the surface would be relatively colder in the LM than in the SP state. The decrease in the surface temperature of the LM state in the OGCM run was larger by approximately 0.6 K in winter (Fig. 12a) and approximately 0.4 K in summer (Fig. 12b) than that in the JRA run. The variation in EPT was more pronounced than that in temperature in both summer (approximately 1.2 K) and winter (approximately 1.0 K; Figs. 12c, d).

We examined the vertical EPT gradient by considering the difference between the 950 hPa and 700 hPa levels in the cold pool area. If the vertical EPT gradient is positive ($\frac{\partial EPT}{\partial p} > 0$), the lower atmospheric layer is unstable, and if it is negative ($\frac{\partial EPT}{\partial p} < 0$), the lower atmospheric layer is stable. We took the difference in the frequency of occurrence of a range of vertical EPT gradient values (divided into categories 0.02 K hPa$^{-1}$ apart), which were calculated at 3-hourly intervals, between the two states (LM minus SP), for both the OGCM and JRA runs (Fig. 13). To elucidate the effect of the LM on vertical EPT gradients, we compared the frequency variation predicted by the OGCM run, in which the LM is relatively exaggerated, with that predicted by the JRA run. In winter, the frequency variation in the OGCM run was more negative than that in the JRA run when the vertical EPT gradient was above −0.01 K hPa$^{-1}$ (Fig. 13a). However, when the vertical gradient was below −0.01 K hPa$^{-1}$, the frequency variation in the OGCM run was more positive. This suggests that atmospheric stability increases when the Kuroshio is in the LM state.
Summer results are separated by month (Figs. 13b, c, d) because intra-seasonal variation is substantial in summer as a result of the Baiu Front and the Pacific subtropical high. In June, the variation in the frequency of vertical EPT gradients in the OGCM run in comparison with those in the JRA run tended to be more negative for vertical EPT gradients of more than 0.01 K hPa\(^{-1}\) and more positive for gradients of less than 0.01 K hPa\(^{-1}\) (Fig. 13b). Frequency variation signals were very weak in July (Fig. 13c), but in August frequency change in the strongly unstable range (0.06 K hPa\(^{-1}\) or more) was strongly negative in the OGCM run (Fig. 13d). Therefore, although the frequency of vertical EPT gradients decreased for positive (unstable) gradients in both summer and winter in the cold SST pool of the Kuroshio LM, frequency in the strongly unstable range decreased substantially in summer. This implies that the depression of near-surface EPT was responsible for the reduced frequency of high-intensity rainfall events in summer.

4.2 The wind-affecting mechanism of the Kuroshio LM

In the OGCM run, the LM state caused a marked variation in wind at the 950 hPa level over the cold pool in comparison with the SP state (Figs. 14a, b). This variation was weak to absent in the JRA run (Figs. 14c, d).

Two mechanisms have been proposed to explain the effects of SST anomalies on wind: vertical mixing (Wallace et al. 1989; Nonaka and Xie; 2003) and pressure adjustment (Lindzen and Nigam 1987; Minobe et al. 2008). These mechanisms have been mainly
studied over the oceanic western boundary currents such as the Gulf Stream and the Kuroshio. The vertical mixing mechanism states that increased static stability around a cold SST anomaly induces reduced vertical mixing, thereby inhibiting momentum transport from above into the lower boundary layer. This results in a decrease in wind speed near the surface. The pressure adjustment mechanism states that increased SLP around the cold SST anomaly causes a pressure gradient between the areas of higher and lower SLP, inducing a variation in surface wind.

Xu et al. (2010) investigated these two mechanisms by examining the variations between atmospheric states in the LM state (from November 2004 to March 2005) and the artificially smoothed SP state. We investigated how these mechanisms may have contributed to variations in winds over the cold SST pool.

Following the horizontal atmospheric momentum equations of Xu et al. (2010),

\[ \frac{Du}{Dt} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + fv - \frac{\partial u'w'}{\partial z} + D_u, \]  
\[ \frac{Dv}{Dt} = -\frac{1}{\rho} \frac{\partial p}{\partial y} + fu - \frac{\partial v'w'}{\partial z} + D_v, \]

where \( u \) and \( v \) indicate horizontal wind; \( w \) indicate vertical wind; \( u' \), \( v' \), and \( w' \) represent the turbulent fractions of \( u \), \( v \) and \( w \), respectively; and \( \rho \) and \( f \) indicate air density and the Coriolis parameter, respectively. \( D_u \) and \( D_v \) indicate horizontal mixing.

We focus on the first and third terms on the right hand side of equations (1) and (2), which are related to pressure adjustments and vertical mixing. For
analyzing the equations, we used the 950 hPa level as the typical height of a boundary layer, although this height may slightly vary between the LM and the SP states.

In winter, the pressure gradient term revealed a clear divergence field above the cold pool (Fig. 15a), which corresponds with the results of Xu et al. (2010). This suggests that increased SLP caused by the cold pool can create a divergent flow. In summer, a similar but less pronounced distribution appeared (Fig. 15b). Although the pressure adjustment mechanism does cause a variation in the wind over the cold pool, this effect is strengthened when the seasonal wind flows in the same direction, and is weakened when it flows in the opposite direction. For example, the winter seasonal wind around Japan is north-westerly, so the wind blows more powerfully on the lee (or south) side of the cold pool.

The vertical mixing term was quite small in comparison with the pressure gradient term (not shown), although the vertical mixing term may have been underestimated at 950 hPa owing to the effect of surface friction. The analyses presented here are consistent with the recent findings by Liu et al. (2013), who showed that the pressure adjustment mechanism tends to dominate on monthly to seasonal time scales whereas the vertical mixing mechanism dominates on shorter (2 to 6 days) time scales (Booth et al. 2010).

5. Summary and conclusions

We conducted a long-term integration of a high-resolution ocean model (NPOGCM), which can resolve eddies and currents in detail. NPOGCM effectively...
reproduced the two main states of the Kuroshio current, although the simulated Kuroshio LM was somewhat exaggerated relative to observational data.

Using the NPOGCM results as a lower boundary condition, we performed a 20-year simulation for 1985–2004 with RCM20. To detect variations in the atmospheric response to the LM and the SP states of the Kuroshio, we used the composite 8-year mean difference of the RCM to examine atmospheric parameters in regions where SST varied considerably between the two states.

Our simulation indicates that the cold SST pool may have two distinct effects on atmospheric convection and rainfall frequency. First, atmospheric properties, such as EPT, change as a result of decreases in latent and sensible heat fluxes. A decreasing vertical gradient in EPT (taking pressure as the vertical coordinate, i.e., positive downward) over the cold SST pool creates increased static stability in the lower boundary layer, which is closely related to convection activity. Furthermore, our experiment shows that the frequency of horizontal steep $\nabla h \theta_e$ decreases at the northern edge of the cold pool, suggesting that the frequency of active weather fronts decreases there.

Second, SLP increases substantially most likely as a result of the cold SST pool. The increase in SLP causes a divergent area over the cold pool and thereby changes surface wind strength via the pressure adjustment mechanism. This wind divergence is likely to increase vertical downward wind speed.

These two effects—the decreased frequency of active weather fronts and the positive SLP anomaly—create unfavorable conditions for precipitation. Although rainfall events are also related to a range of other factors, such as upper cold low, potential vorticity, and pressure patterns, the cold SST pool of the LM could be an additional factor in changing the atmospheric environment toward a state unfavorable for precipitation. Although the Kuroshio LM does not necessarily have a causative role in severe weather events, it is possible that it causes unfavorable convection conditions.

Our simulations also revealed a decrease in surface air temperature and precipitation frequency, and an increase in SLP in the coastal region on the Pacific Ocean side of Japan during the Kuroshio LM phases. Moreover, we detected a decrease in temperature at the 850 hPa level, which is far above the sea surface, although this was relatively small in comparison with the variations in surface air temperature. This implies that variations in Kuroshio’s path primarily affect the climate of the Pacific Ocean near Japan, as demonstrated by the modified atmospheric parameters over the cold pool. However, the variations over the land area of the Pacific Ocean side of Japan caused by the Kuroshio LM were at most 20–50 % of the magnitude of those over the cold SST pool area. Atmospheric responses to the LM were negligible over other areas of Japan.

In the simulations created for this study, atmospheric responses to the Kuroshio LM, which were unclear in the observational data, were clearly evident. These results will contribute to the studies of regional climate system. Further investigation is required to clarify whether this signal can be detected in the observational data.

Acknowledgments

The authors thank Dr. I. Takayabu and Prof. A. Kito for their valuable suggestions. We would like to thank two anonymous referees and Dr. Y. Tanimoto for their insightful and constructive comments. We are also grateful to Dr. N. Usui who provided MGDSST datasets. This research was supported in part by the MRI special research program “Comprehensive projection of climate change around Japan due to global warming” and supported by the Grant-in-Aid for scientific research #22106009 by the Ministry of Education, Culture, Sports, Science and Technology, Japan.

References


