Roles of Oceanic Mesoscale Eddy in Rapid Weakening of Typhoons Trami and Kong-Rey in 2018 Simulated with a 2-km-Mesh Atmosphere-Wave-Ocean Coupled Model

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Abstract

In 2018, Typhoon Trami made landfall in Japan and maintained its intensity for a few days, then rapidly weakened after its recurvature. Subsequently, Typhoon Kong-Rey passed through the waters cooled by Trami while rapidly weakening. The region where both typhoons rapidly weakened is a region rich in oceanic mesoscale eddies overlying the Subtropical Countercurrent. To understand the role of a cold-core eddy, in the intensity change of these two typhoons, we examined the similarity and differences between the two typhoons, utilizing numerical simulations with a 2-km-mesh nonhydrostatic atmosphere model and an atmospheric-wave-ocean coupled model. Sensitivity experiments were performed by assuming a significant magnitude on the weakening of Trami during the mature phase; for example, we embedded an artificial cold-core eddy with a magnitude not based on in situ observations to gauge initial oceanic conditions. In contrast for Kong-Rey, nine ensemble simulations for initial atmospheric conditions were conducted instead of different-day initial oceanic conditions. The simulated rapid weakening of two typhoons was related to the low upper-ocean heat content caused by typhoon-induced sea surface cooling (SSC). Most simulations for Trami and Kong-Rey show a tendency of overdevelopment during the mature or weakening phase; the overdevelopment of Trami is caused by insufficiently simulated SSC and the embedded artificial cold eddy, which promoted the SSC; whereas, the overdevelopment of Kong-Rey is related to the failure of track simulation. A reasonable simulated track of Kong-Rey required greater time traveling over the Trami-induced SSC area to enhance weakening by reduction in inner-core moisture transport toward the center near the surface and in the inflow boundary layer on the upshear side. The reductions in downward motion in the center and the associated adiabatic heating were closely related to weakening in both typhoons.

Keywords atmosphere-wave-ocean coupled model; oceanic mesoscale eddy; upper-ocean heat content; rapid weakening


1. Introduction

The western North Pacific region, area where warm and cold mesoscale eddies with a horizontal scale of a few hundred kilometers exist north of 20°N (Hwang et al. 2004; Lin et al. 2008), generates the most tropical cyclones (TCs) in the world (Peduzzi et al. 2012). In this location, interactions between TCs and eddies occur at remarkably high frequencies as
revealed by statistical analysis; from 2001 to 2011, more than 90% of the TCs recorded over the western North Pacific encountered oceanic mesoscaleeddies (Ma et al. 2017). Typically noted in this region, TCs exhibit various behaviors such as rapid intensification, maintenance of TC intensity, and rapid weakening or decaying over the Subtropical Countercurrent. Rapid intensification is generally defined as an increase of 30 kt or more in the maximum 1-m surface wind speed within 24 h over water (Kaplan and DeMaria 2003). However, various researchers have defined rapid weakening differently; as a decrease of 20 kt (DeMaria et al. 2012; Liang et al. 2018; Ma 2020), as a decrease of 30 kt (Wood and Ritchie 2015), as a decrease of 40 kt or more (Ma et al. 2019) in the maximum 1-min surface wind speed within 24 h. Current weather prediction systems have difficulty forecasting rapid changes in TC intensity, although their ability in forecasting intensity has improved (Courtney et al. 2019).

Numerous observational and numerical studies have shown that rapid weakening is closely associated with atmospheric and oceanic environmental conditions; stable stratification in the lower to middle troposphere, induced by vertical wind shear, which is the difference in horizontal winds between 850 hPa and 200 hPa (DeMaria 1996), seawater temperature (Ma 2020), and lower to mid-level intrusion of dry air (Colomb et al. 2019; Zhang et al. 2007) are examples of environmental conditions that impact rapid weakening. Thus, these conditions related to the oceanic environment are deeply involved in the genesis and development of TCs (Gray 1979). Warm oceanic environments, such as mesoscale warm-core eddies, promote rapid intensification of TCs in the western North Pacific (Lin et al. 2008; Wu et al. 2007), and cold oceanic environments such as mesoscale cold-core eddies not only suppress the intensification of TCs but also lead to rapid weakening (Ma 2020; Johnston et al. 2020). However, not all eddies appreciably affect the intensity changes of TCs (Ma et al. 2018). A TC traveling across the ocean may lower the sea surface temperature (SST) along the TC track by as much as approximately 9°C (Sakaida et al. 1998; Walker et al. 2014). The cooling effect of TCs is influenced by atmospheric factors (i.e., the intensity, size, and speed of TCs) and by oceanic factors (i.e., the stratification of the upper ocean and the depth of the upper-ocean mixed layer underneath TCs) (e.g., Price 1981; Wada 2002). To understand the effect of TC-induced sea surface cooling (SSC) in relation to TC intensity predictions is not a straightforward task due to the complexity of the oceanic response to TCs.

Typhoon Trami (2018), which made landfall in Japan, is an example of a TC that traversed the eddy-rich area over the Subtropical Countercurrent (Fig. 1). Trami moved northwestward after its genesis and rapidly intensified over the ocean south of Okinawa before arriving at the surrounding area at (20°N, 130°E). When Trami was stagnant south of Okinawa for the 12 h period from 1200 UTC on 25 September to 0000 UTC on 26 September, the maximum 10-min surface wind speed decreased by 15 kt according to the Regional Specialized Meteorological Center (RSMC) Tokyo best track data (Fig. 2a). Trami then moved to the northeast after rapidly weakening. SSC was induced by Trami primarily around (20°N, 130°E) in its intensification phase and around (22.5°N, 128°E), when the storm was moving slowly or stagnant (Fig. 3a).

The Japan Meteorological Agency (JMA) currently issues TC intensity forecasts through RSMC Tokyo based on the methodology entitled, “Typhoon Intensity Forecast (TIFS) scheme based on Statistical Hurricane Intensity Prediction Scheme (SHIPS)” (e.g., Shimada et al. 2018), which is a statistical scheme with a performance superior to that of dynamical models (Rappaport et al. 2012). The RSMC best track data show that rapid weakening of Trami and Kong-Rey occurred at 20 kt for Trami and 35 kt for Kong-Rey during the period shown in Figs. 2a, b. The track prediction by JMA atmospheric global spectral model at the initial value of 0000 UTC on 26 September was typical of the RSMC best track and the data output used as explanatory variables of TIFS. However, the TIFS forecasted overdevelopment of Trami (Table S1) when the TC traveled northeastward and maintained a central pressure of 950 hPa (Figs. 2c, d). The forecast of overdevelopment in relation to the rapid weakening was a serious error because it signifies that the TC would make landfall in Japan without weakening, and that this landfall would add to the preexisting damage caused by Typhoon Jebi (2018) in Japan (Takemi et al. 2019).

Additionally, a week later after Trami made landfall in Japan, the forecast of rapid weakening for Typhoon Kong-Rey was challenging since Kong-Rey was generated by Rossby-wave propagation (Holland 1995; Carr and Elsberry 1995; Li and Fu 2006) from Trami around (12.6°N, 142.6°E). Kong-Rey followed a track similar to the trajectory of Trami and passed over the cold wake induced by Trami at around (22.5°N, 128.0°E) (Figs. 1b, d, 3b), after which it rapidly weakened (Figs. 2b, d). From 0600 UTC to 1800 UTC on 2 October, the maximum 10-min surface wind decreased by 25 kt (Fig. 2b) and the central pressure
increased by 40 hPa (Fig. 2d), according to the RSMC best track data. Although the track prediction by JMA atmospheric global spectral model of the initial value during the period was reasonable against the RSMC Tokyo best track, TIFS predicted that Kong-Rey would also develop against rapid weakening as Trami’s case, resulting in error of intensity forecasting.

In the International Best Track Archive for Climate Stewardship dataset (Knapp et al. 2010), best track analyses by the China Meteorological Administration’s Shanghai Typhoon Institute, Hong Kong Observatory, and the U.S. Department of Defense Joint Typhoon Warning Center all indicated rapid weakening of the two TCs, which was consistent with the RSMC Tokyo best track data analysis. These results yield greater understanding of the role of atmosphere-ocean interaction that impacts the intensity change in the two TCs. In fact, it is difficult to quantitatively show the real magnitude of cold-core eddy during the passage of TCs where in situ observations are sparse. More specifically, there are not enough in situ observations to quantitatively separate the TC-induced cold-core eddy from the preexisting cold-core eddy. Furthermore, it is difficult to investigate the quantitative change in intensity of TCs while passing over the cold-core eddy because of very rare in situ observations around the TC center. Therefore, this study focuses on qualitatively understanding the mechanism of rapid weakening of Trami and Kong-Rey during the passage over the artificial cold-core eddy, which has a significant magnitude to possibly affect intensity change. Therefore, in this study, we performed numerical simulations with a 2-km-mesh nonhydrostatic atmosphere model and an atmosphere-wave-ocean coupled model with different atmospheric and oceanic initial conditions. The results of the simulations and sensitivity experiments regarding preexisting atmospheric and oceanic environments, best track analysis data, and satellite observations were used to investigate the various factors that may have affected rapid weakening. The magnitude of the artificial cold-core eddy is hypothetically determined by assuming a significant magnitude that possibly affects the intensity simulation of Trami because of sparse in situ observations utilizing profiling floats (see Section 2.4). The purpose of this study is to understand the potential roles of an artificial cold-core eddy on the intensity change of Trami and Kong-Rey in their simulations particularly during the rapid weakening and subsequent mature phases of Trami and the rapid weakening phase of Kong-Rey as well as explore the similarity and difference of the roles between simulated Trami and Kong-Rey, utilizing a 2-km-mesh nonhydrostatic atmosphere model and an atmosphere-wave-ocean coupled model.

The remainder of this paper is organized as follows: Section 2 describes the data, experimental design, and methods of analysis used. Section 3 presents the results of numerical simulations, sensitivity experiments, and backward trajectory analyses for Trami and Kong-Rey. The results of this study are summarized and discussed in Section 4.

2. Experimental design

2.1 Atmospheric and oceanic data

To establish initial conditions for our modeling experiments before the passage of Trami and Kong-Rey, we identified mesoscale eddies in the study area from the Archiving, Validation, and Interpretation of Satellite Oceanographic (AVISO) dataset of daily sea surface height anomalies (SSHAs), with a horizontal resolution of $0.25^\circ \times 0.25^\circ$ on a Cartesian grid. The resulting maps show a succession of alternating cold-core and warm-core eddies over the Subtropical Countercurrent on 19 September (Fig. 1a) and 29 September (Fig. 1b), although the contrast between them was smaller than it was over Kuroshio, a powerful western boundary current directly south of Japan. The location of mesoscale cold eddies was consistent with the analysis obtained from the actual time of the mesoscale eddy trajectory atlas project (Mason et al. 2014).

We also mapped the distribution of SSHAs on the same dates from the Four-dimensional variational Ocean ReAnalysis (FORA) dataset for the Western North Pacific (FORA-WNP30; https://synthesis.jamstec.go.jp/FORA/e/index.html; Usui et al. 2017). The FORA dataset used in this study is the three-dimensional variational ocean reanalysis dataset for the North Pacific and includes water temperature, salinity, oceanic currents, and sea-level anomaly data with a $0.5^\circ$ horizontal resolution. The arrangement of eddies obtained from the FORA dataset (Figs. 1c, d) was not always consistent with the data obtained from the AVISO observations even though the FORA dataset includes the satellite SSHA data through assimilation. It should be noted that the reference of SSHA differs between AVISO, SSHA, and FORA sea-level anomalies because the latter value is a deviation from the model surface, while the former value is a difference from the mean state. In addition, we used the Merged Satellite and In-situ Data Global Daily Sea Surface Temperatures (MGDSTT) dataset (Kurihara et al. 2006), with a horizontal resolution of $0.25^\circ$ to create the oceanic initial condition for the simulations.
conducted by the atmosphere-wave-ocean coupled model. MGDSST is also included in the FORA dataset because it is used as a boundary condition.

We also used SST data from the Microwave Optimally Interpolated SST (noted as OISST hereafter) gathered on a daily basis and obtained from the Remote Sensing Systems website (https://www.remss.com/). OISST, a daily dataset that merges SST data measured by the TRMM/TMI and AMSR-E satellite radiometers, covers the entire ocean around the earth with a 0.25° horizontal resolution at a depth of approximately 1 m. The OISST values were corrected by using a diurnal model that represents the daytime temperature at 8 AM local time (Gentemann et al. 2003). The resulting SST maps are shown for 27 September (Fig. 3a) and 5 October (Fig. 3b). Trami-induced SSC, calculated as the difference in SST from 23 September, was approximately 6.8°C around (20°N, 130°E) on 27 September and approximately 5.9°C around (22.5°N, 128°E) on 29 September. The areas of SSC persisted during the passage of Kong-Rey. Kong-Rey induced SSC, calculated as the differ-

Fig. 1. Horizontal distributions of daily mean AVISO sea surface height anomaly (m) on (a) 19 September and (b) 30 September 2018 and of MOVE sea-level anomaly at 0000 on (c) 19 September and (d) 30 September 2018. Circles indicate the position of Trami and squares indicate the position of Kong-Rey; colors within the symbols indicate the central pressure (horizontal color bars indicates the value of the colors). Asterisks indicate the location of mesoscale cold eddies obtained from the near-real-time mesoscale eddy trajectory atlas project (Mason et al. 2014). TC data are based on RSMC Tokyo best track data.
Fig. 2. Time series of RSMC Tokyo best track 10-min maximum sustained wind speed (vertical axis: m s$^{-1}$) for (a) Trami and (b) Kong-Rey (black lines and dots) and best track central pressure (vertical axis; hPa) for (c) Trami and (d) Kong-Rey together with 24-h (red open circles), 48-h (orange dots), 72-h (yellow open circles), 96-h (green dots), and 120-h (light-blue open circles) forecasts at intervals of 6 h. A box in each panel indicates the rapid weakening period of Trami or Kong-Rey.

Fig. 3. Horizontal distribution of microwave daily SST on (a) 27 September and (b) 5 October 2018. Circles indicate the position of Trami and squares indicate the position of Kong-Rey; colors of the symbols indicate the central pressure. TC data are based on RSMC Tokyo best track data.
ence in SST from 29 September, was approximately 2.4°C around (22.5°N, 128°E), which was smaller than Trami-induced SSC. OISST was also used as the initial SST condition in all simulations of Trami (Table 1) and some of Kong-Rey’s simulations (Table 2) by replacing MGDSST with OISST.

The atmospheric initial and boundary conditions were obtained from JMA, utilizing 6-hourly global atmospheric datasets for analysis with a horizontal grid spacing of approximately 20 km. In addition, global atmospheric ensemble prediction data with a horizontal grid spacing of 1.25° obtained from JMA were used to add perturbations to these conditions in the case of Kong-Rey.

### 2.2 Tropical cyclone heat potential

The upper-ocean heat content, calculated in the Eq. (1), is known as tropical cyclone heat potential (TCHP), which is related to TC intensity and intensification (Wada and Usui 2007; Wada 2015). TCHP is a measure of the oceanic heat content from the surface to the 26°C isotherm depth (Leipper and Volgenau 1972). In this study, TCHP is defined as

\[
Q_{\text{TCHP}} = Cp\left[\rho_1(T_1-26)h_1 + \frac{1}{2}\rho_2(T_2-26)(D_{26}-h_1)\right],
\]

where \(Q_{\text{TCHP}}\) is a value of TCHP for a grid cell, \(\rho_1\) is the density of seawater in the mixed layer, \(\rho_2\) is the seawater density at the seasonal thermocline, \(T_1\) and \(T_2\) are the water temperature at the top and bottom of the mixed layer respectively, \(h_1\) is the thickness of the mixed layer, \(D_{26}\) is the depth of the 26°C isotherm, and \(Cp\) is the specific heat at constant pressure. In the time series analysis, \(Q_{\text{TCHP}}\) and \(D_{26}\) are obtained by averaging a square area of 26 km (approximately 0.25° on a Cartesian grid) with the simulated TC center as its center. In Eq. (1), TCHP becomes small when \(T_1\) and \(T_2\) approach 26°C or when \(D_{26}\) is relatively insignificant.

Figures 4a–g show the distribution of daily TCHP calculated in Eq. (1) from 19 to 25 September. TCHP was greater than 80 kJ cm\(^{-2}\) around the track of Trami, which was greater than 50 kJ cm\(^{-2}\), which was a threshold of TC intensification (Wada 2015) and enough for Trami to intensify. Although we do not

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### Table 1. Ensemble sensitivity experiments for Trami.

### Table 2. Ensemble sensitivity experiments for Kong-Rey.
Fig. 4. Horizontal distribution of TCHP at the initial time of the integration using oceanic initial conditions based on oceanic analysis data for (a) 19 September, (b) 20 September, (c) 21 September, (d) 22 September, (e) 23 September, (f) 24 September, and (g) 25 September, and for (h) 23 September with the insertion of an artificial cold-core eddy (Table 1). In all panels, the OISST product is used as SST at the initial time.
know the real magnitude of TCHP during the passage of Trami because of sparse data from in situ observations utilizing profiling floats, we easily notice that the distribution of daily TCHP estimated from ocean data analysis lacks information of preexisting TC-induced mesoscale cold-core eddies shown in Fig. 1a.

2.3 Models

The atmosphere-wave-ocean coupled model used in this study has been widely used to investigate the role of the ocean in TC predictions (e.g., Wada et al. 2018; Wada and Oyama 2018; Oyama and Wada 2019) and to compare simulation results with climatological data for verification (Wada et al. 2020). As shown in Fig. 5, the coupled model consists of a nonhydrostatic atmospheric model (NHM; Saito 2012), which is a Meteorological Research Institute (MRI) third-generation ocean surface-wave model (MRI-III; Japan Meteorological Agency 2013), and an MRI multilayer ocean model based on Bender et al. (1993). The multilayer ocean model includes a diurnally varying SST scheme based on Schiller and Godfrey (2005) with the shortwave absorption and penetration formulation of Ohlmann and Siegel (2000).

The NHM includes microphysics expressed in an explicit three-ice bulk microphysics scheme (Ikawa and Saito 1991, Lin et al. 1983), air–sea momentum, and sensible and latent heat fluxes. Exchange coefficients for air–sea momentum and enthalpy transfers across the ocean are based on the following: bulk formulas (Kondo 1975), or when the ocean-wave model is coupled (Wada et al. 2010), on the basis of rough lengths proposed by Taylor and Yelland (2001), a sea spray formulation with whitecap coverage of approximately 4% under TC conditions (Wada et al. 2018), a turbulent closure model (Klemp and Wilhelmson 1978; Deardorff 1980), and a radiation scheme (Sugi et al. 1990).

The MRI-III predicts wave spectra as a function of space and time from an energy balance equation composed of the spectral energy input by the wind, the nonlinear transfer of spectral energy due to wave–wave interactions, and the dissipation of energy due to breaking surface waves and whitecap formation. The wave spectrum had 900 components, each associated with a wave spectra of 25 frequencies (divided logarithmically from 0.0375 Hz to 0.3000 Hz) and 36 directions [divided every 10° from 0° (north) to 350°].

The MRI multilayer ocean model employs a reduced gravity approximation and a hydrostatic approximation, and it assumes that the water is a Boussinesq fluid (Bender et al. 1993; Wada 2002). The model has as many as three layers and four interfaces. The four interface levels are: the sea surface, the base of the mixed layer, the base of the thermocline, and the seafloor. The uppermost layer is the mixed layer with a thickness of $h_1$ in Eq. (1), in which the density ($\rho_1$) is vertically uniform. The density $\rho_1$ is determined by SST ($T_1$) and salinity, while the density $\rho_2$ is determined by water temperature and salinity at the mixed layer base. At the initial time, the mixed layer depth was determined from the oceanic reanalysis.
data based on the criterion within the mixed layer that the difference in density from the surface \((\rho_1-\rho_2)\) was < 0.25 kg m\(^{-3}\), and the lower limit of the mixed layer was set to 200 m. The definition is the similarly used by Wada et al. (2018). The middle layer is the seasonal thermocline, where the vertical temperature gradient is greatest. \(D_{sh}\) is determined by linear interpolation by using the water temperatures at the mixed layer base and at the seasonal thermocline base where \(T_1\) and \(T_2\) are greater than 26°C. The bottom layer is assumed to be undisturbed by entrainment. The model calculates the water temperature and salinity at the surface and at the base of the mixed layer, the thickness, and the horizontal flows of each layer. The entrainment rate is calculated only at the base of the mixed layer (Dear-dorff 1983).

The exchange processes between the NHM, MRI-III, and MRI multilayer ocean models (Fig. 5) are the same as those in Wada et al. (2018). The time steps of these exchange processes are explained in the following: shortwave and longwave radiation, air–sea sensible and latent heat fluxes, wind stresses, cloud cover, and precipitation calculated by the NHM were used in the multilayer ocean model during each time step. The SST calculated by the multilayer ocean model at every time step was used in the atmospheric model. Surface wind speeds calculated by the NHM were provided to the MRI-III at every time step of the MRI-III. Wave heights calculated by the MRI-III were provided to the NHM to calculate the steepness of ocean waves for estimating the surface roughness length. Ocean currents in the uppermost layer calculated by the multilayer ocean model were used by the MRI-III during each time step. These ocean currents were also used within the multilayer ocean model to modify the calculations of the velocity of ocean waves as a group. Wave-induced stresses calculated by the MRI-III were provided, utilizing the multilayer ocean model to modify the surface wind stress and to estimate the entrainment rate caused by the breaking of waves.

2.4 Configuration of numerical simulations

The configuration of the numerical simulations differed between the two TC cases in two respects: the integration period was from 0000 UTC on 23 September to 0000 UTC on 30 September 2018 for Trami and from 1200 UTC on 29 September to 1200 UTC on 5 October 2018 for Kong-Rey, and the computational domain was approximately 2280 km \(\times\) 3120 km for Trami and 2520 km \(\times\) 3000 km for Kong-Rey. Both simulations used a time interval of 3 s for the NHM, 18 s for the multilayer ocean model, and 6 min for the MRI-III. All NHM simulations had 55 levels in vertical coordinates, with intervals ranging from 40 m for the near-surface layer to 1180 m for the uppermost layer, and a top height of approximately 27 km.

Three sets of sensitivity experiments were conducted in this study. In Fig. 1, there were two mesoscale cold-core eddies along the track of Trami and Kong-Rey around (20°N, 130°E) and (22.5°N, 128°E) although the distribution of TCHP estimated from oceanic data analysis lacks the eddies’ information (Figs. 4a–g). However, as described in Section 2.2, the actual magnitudes of TCHP on the two cold-core eddies are difficult to estimate due to sparse data of in situ observations. Therefore, this study estimated the potential value of TCHP just below the simulated Trami near the area of (22.5°N, 128°E) when the TC started to show a weakening tendency. TIFS was used to determine the weakening tendency that followed the decrease in TCHP and potential value of TCHP. Table S1 shows the results of TIFS initiated at 0000 UTC on 26 September with different values of TCHP (42 kJ cm\(^{-2}\) to 71 kJ cm\(^{-2}\)) from 12 h to 24 h forecast time when Trami was located at (22.5°N, 128°E). The value of 71 kJ cm\(^{-2}\) used in this operation shows the decrease in the central pressure of 6 hPa (from 954 hPa to 948 hPa) from 12 h to 24 h forecast time. When the value of 42 kJ cm\(^{-2}\) was used, the central pressure became 954 hPa, which is higher than 952 hPa when the value was 71 kJ cm\(^{-2}\). Although the difference of 2 hPa was slight, the weakening tendency followed the decrease in TCHP approximately 30 kJ cm\(^{-2}\). Interestingly, the value of 42 kJ cm\(^{-2}\) is lower than 50 kJ cm\(^{-2}\), which is noted as the threshold of TC intensification (Wada 2015). In TIFS, the impact of TCHP on predicted central pressures are small because TIFS forecasts TC intensity changes by a multiple regression equation with 26 explanatory variables, and any changes in the atmospheric explanatory variables caused by the decrease in TCHP are not considered. The relatively slow translation of the actual TC compared to that of the simulated TC may also affect the predicted central pressures in TIFS. Therefore, we first used a modified oceanic initial condition that embedded an artificial cold-core eddy having the minimum TCHP of approximately 42 kJ cm\(^{-2}\), as shown schematically in Fig. 6, to investigate the impact of TC-induced SSC varied by the amplitude of the mesoscale cold-core eddy on TC simulations. In Table S1, the forecast location of Trami in TIFS was around (21°N, 129°E) from 1200 UTC on 26 September to 0000 UTC on 27 September, corresponding to the forecast times from 12 h to 24 h. Based on the result of TIFS, the artificial eddy
hypothetically implanted in the original oceanic initial condition at the initial integration time was represented by negative anomalies of oceanic mixed-layer thickness within a radius of 200 km. The maximum negative anomaly of the mixed layer thickness and seasonal thermocline position was 50 m at the center (21°N, 129°E) and decreased along a quadratic curve toward the outer edge. The location (21°N, 129°E) was determined based on the cold-core eddy distribution on 25 September (Fig. 1a), the results of TIFS in Table S1, and the results of track simulations, although the shape of the cold-core eddy was not circular. The experiment for Trami with an artificial cold-core eddy having the minimum TCHP of approximately 42 kJ cm$^{-2}$ around (21°N, 129°E) is labeled as “AWO_cold” (Table 1), which hypothetically added the artificial cold-core eddy to the 23 September initial conditions. The experiments for Kong-Rey are labeled as “MGDSST_COLD,” and “OISST_COLD” (Table 2), which hypothetically added the artificial cold-core eddy to the 29 September initial conditions. “A” means a simulation by the noncoupled NHM, while “AWO” means a simulation by a coupled NHM-wave-ocean model. Due to the constraint of the ocean model used in our study, we calculated the thickness only in the mixed layer, seasonal thermocline, and the associated water temperature; the vertical profile of the ocean must change unrealistically as referenced in Fig. 6 to reflect the decrease in TCHP of approximately 30 kJ cm$^{-2}$ at the initial integration time as described in the introduction.

The second sensitivity numerical experiments used oceanic reanalysis data and OISST from different dates as the oceanic initial condition, as listed in Table 1 for Trami, to investigate the impact of oceanic variations on a weekly timescale on TC simulations during the change in the preexisting oceanic condition and simulated oceanic response to a TC. Additionally, the experiments used the oceanic data on the same date but from the different SST datasets, as listed in Table 2 for Kong-Rey, to investigate the impact of the replacement of the SST product on the TC simulations. The four values displayed in Table 1 represent the month and day of daily oceanic data used to create the initial condition. Figure 4h shows the distribution of daily TCHP on 23 September with the artificial cold-core eddy included. In this study, a mesoscale eddy with a TCHP value lower than 60 kJ cm$^{-2}$ and with the horizontal scale of a hundred kilometers is defined as a cold-core eddy. The distribution of daily TCHP on 23 September (Fig. 4e) is also used to calculate the evolution of TCHP around the simulated TC in the A_0923 experiment. Figures 4a–g show that TCHP was higher than 80 kJ cm$^{-2}$ around the track of Trami, which was sufficient for Trami to intensify. These results indicate that the TC simulations were not affected by preexisting mesoscale cold eddies as analyzed in Fig. 1a. From 19 to 25 September, the average FORA TCHP value was calculated in a 2° × 2° squared area centered at (21°N, 129°E), which exceeds 80 kJ cm$^{-2}$ and demonstrates that the variation is small when compared to other TCHP data (Fig. S1). In other words, the variation of TCHP in Figs. 4a–g is primarily caused by the difference in OISST SSTs from different dates. The cold wake was well reproduced by hypothetically embedding the artificial cold-core eddy with a minimum TCHP value of approximately 42 kJ cm$^{-2}$, which is lower than the threshold value (50 kJ cm$^{-2}$, Wada 2015) of TC intensification in the oceanic initial condition and is close to the lowermost value of the TCHP data on 23 September (Fig. S1). The location of the artificial cold-core eddy around (22.5°N, 129°E) (Fig. 4h) is consistent with the SSHA distribution in Fig. 1a, although the value of SSHA is given as an integral of the water density from the sea surface to the sea floor in a water column and does not yield the vertical profile of water temperature and salinity above the 26°C isotherm depth.

Figure 7 shows the distribution of daily TCHP on 29 September in the experiments OISST, MGDSST, and MDGSST_COLD (Table 2). The distribution of daily TCHP on 29 September (Fig. 7a) is used to calculate the evolution of TCHP around the simulated TC in the A_OISST experiment. All three simulations
featured a cold-core area around (22.5°N, 130°E) in which the value of TCHP was almost zero. The area of daily TCHP was lower than 20 kJ cm$^{-2}$ in the OISST experiment (Fig. 7a) and was larger than the area of daily TCHP in the MGDSST experiment (Fig. 7b). This result implies that the ocean’s response to Trami was not well analyzed in daily TCHP used in the MGDSST. For this reason, the same artificial cold-core eddy was hypothetically implanted into the original oceanic initial condition with the assumption that high TCHP tendency continued for a week due to the 5-day assimilation window in the three-dimensional variational ocean reanalysis system (see Figs. 4a–g, S1). Because the same oceanic data except for SST was used in the simulation by the coupled model and FORA, TCHP was relatively high when compared to the other TCHP data (Fig. S1); the same artificial cold-core eddy was also implanted into the original oceanic initial condition in the OISST_COLD experiment.

The third sensitivity numerical experiments only pertaining to Kong-Rey were atmospheric ensemble experiments conducted only by the atmosphere-wave-ocean coupled model; these were conducted with nine different atmospheric initial and boundary conditions created from global objective analysis by adding perturbations derived from the JMA operational global ensemble prediction system to investigate the effect of different atmospheric initial conditions on the TC simulations, particularly the effect on TC intensity simulations across the change in the track simulations.

2.5 Back trajectory analyses

Previous studies have reported that there are two pathways that affect the rapid weakening of TCs, (1) wind shear-induced downward advection of cool air with low equivalent potential temperature in the inflow boundary layer and (2) mid-level inflow of dry air toward the TC center (e.g., Colomb et al. 2019; Wood and Ritchie 2015; Zhang et al. 2007). The importance of moisture transport toward the TC center near the surface due to surface friction and in the inflow boundary layer on TC–ocean interactions has been already pointed out in Wada and Oyama (2018) and Wada et al. (2018), although both studies did not mention the impact of rapid weakening. To elucidate the impact of the ocean on the rapid weakening of Trami and Kong-Rey, we conducted back trajectory analyses for a set of air parcels (Wada and Oyama 2018; Wada et al. 2018), using a second order Runge-Kutta method. A hundred parcels were released from the central position of simulated TCs only at the 90-h integration time in experiments A_0923, AWO_0923 and AWO_cold for Trami and the experiments A_OISST, AWO_OISST_COLD and AWO_OISST_COLD_006 for Kong-Rey. The initial vertical locations of the parcels at the center of simulated TCs were at heights of 0.02, 0.05, 0.2, 0.5, 1.0, and 1.5 km. Note that “006” represents the member index of ensemble simulation for the atmospheric initial condition. The back trajectory analyses ran, in time steps of 30 s, for 18 h from 1800 UTC on 26 September to 0000 UTC on 26 September in the case of Trami, and from 0600...
UTC on 3 October to 1200 UTC on 2 October in the case of Kong-Rey. For each level, 100 trajectories were traced backward for 18 hours from the 90-h integration time.

3. Results

3.1 Track and central pressure simulations

Figure 8 shows the simulation results for the track, central pressure, and maximum surface wind of Trami. The height of simulated maximum surface wind was 20 m, so it was necessary to multiply that value by approximately 1.067 (= 1/[ln(20/z₀)/ln(10/z₀)], where z₀ (typical value of the roughness over the ocean) = 0.0003 m and the unit of 10 and 20 is a meter) to convert it to a 10-m wind speed to compare it with the RSMC best track maximum wind speed. The simulated track is a reasonable match of the RSMC best track, including the irregularity at approximately 19°N, 129°E (Fig. 8a). Trami-induced SSC was adequately simulated, and was centered around (21°N, 129°E) in the experiments, utilizing the atmosphere-wave-ocean coupled model. The maximum decrease in SST induced by Trami from the initial time, which occurred at [17–22.5°N, 127.5–132°E], was approximately 2.5°C in the AWO_0923 experiment and approximately 2.0°C in the AWO_cold experiment. A warmer initial oceanic condition leads to a more intense simulated TC that causes greater SSC. However, the maximum SSC of approximately 6.8°C in the location of (20°N, 130°E), analyzed from daily OISST data, was not simulated in all experiments, possibly due to the weakly simulated intensity of Trami and limitation of the capability of the oceanic model (Yablonsky and Ginis 2009).

Figure 8b shows the time series of simulated central pressures along with the RSMC best track central pressure. The A_0923 experiment, which used the uncoupled NHM, failed to simulate either the rapid
increase in central pressure from 0600 UTC on 25 September to 0000 UTC on 26 September or the rapid decrease in the maximum surface wind at the same time (Fig. 8c). The result of ensemble mean (A_mean) shows that the change in the preexisting oceanic condition on a weekly time scale has less effect (the standard deviation is \( \sim 2 \) hPa) on the improvement of the rapid weakening of simulated Trami (Figs. 8b, c). In contrast, the coupled model reasonably simulated this rapid weakening of Trami in the AWO_0923 experiment. This result suggests that Trami-induced SSC played an essential role in the rapid weakening. However, the simulated Trami redeveloped with the coupled model by approximately 5 hPa after 0000 UTC on 26 September. The overdevelopment was suppressed in the AWO_cold experiment in which an artificial cold-core eddy centered at (21°N, 129°E) with the minimum TCHP of approximately 42 kJ cm\(^2\) was embedded in the initial oceanic condition (see Section 2.4). This result suggests that significantly low TCHP at the artificial cold-core eddy played a crucial role in suppressing the redevelopment of Trami after the recurvature; therefore, Trami in the AWO_cold experiment could retain high central pressure and low maximum surface wind speed to some extent after the rapid weakening. The location of the artificial cold-core eddy around (21°N, 129°E) is effective on suppression of the redevelopment of the simulated Trami after the rapid weakening partly because the simulated track shifts eastward when compared to the RSMC best track at around 21°N from 1200 UTC on 26 September to 0000 UTC on 27 September.

Figure 9 shows the simulation results for the track, central pressure, and maximum surface wind of Kong-Rey. Unlike the simulations for Trami, the track simulations showed a westward deflection after 0000 UTC on 3 October 2018. The four different oceanic initial conditions (Table 2) made no difference in the

![Fig. 9. Simulation results and best track data for Kong-Rey. (a) Best track and horizontal distribution of simulated SST at 1800 UTC on 4 October, (b) simulated and best track central pressure, and (c) simulated maximum wind speed and best track 10-min sustained maximum wind speed.](image-url)
track simulations by the coupled model. It should be noted that the track simulations by the uncoupled NHM (not shown in Fig. 9a) were almost identical to the coupled model. Regarding the central pressure simulations, the central pressure was deeper in the experiments A_MGDSST and A_OISST than in the other experiments. The A_OISST experiment showed the deepening and the subsequent weakening of simulated central pressure earlier than the A_MGDSST experiment although both experiments showed similar overdevelopment in the weakening phase. The uncoupled NHM and the coupled model could not simulate the rapid weakening of Kong-Rey after 2 October irrespective of the implant of the artificial cold-core eddy. The differences in the SST distribution at the initial time had a greater impact on the simulations of central pressure and maximum surface wind than low TCHP at the artificial cold-core eddy. In the case of Kong-Rey, the effect of the artificial cold-core eddy in the uncoupled NHM experiments was comparable to that of the ocean coupling without the artificial cold-core eddy.

3.2 Role of the ocean in rapid weakening of simulated Trami

Oceanic thermodynamic environments underneath a TC vary depending on the ocean’s response to the TC, and this variation in turn affects the evolution of the TC (e.g., Wada and Usui 2007). This subsection reports our investigation of the oceanic thermodynamic environment underneath simulated Trami in terms of TCHP. The time series of $Q_{TCHP}$ and $D_{26}$ for Trami (Fig. 10) display the ensemble mean values and standard deviations for experiments AWO_0919 to AWO_0925 instead of the values of the seven individual experiments. The insertion of the artificial cold-core eddy in the AWO_cold experiment led to a marked decrease in $Q_{TCHP}$ and $D_{26}$ underneath the TC from 0600 UTC on 25 September to 0000 UTC on 26 September. The low TCHP values in the artificial cold-core eddy were outside the ensemble spread estimated by the standard deviations, indicating that the impact on TCHP when the cold-core eddy was implanted in the initial oceanic condition was significantly greater than the ensemble spread of TCHP from experiments AWO_0919 to AWO_0925. These results suggest that the experiments without the artificial cold-core eddy cannot simulate the effect of low TCHP underneath Trami as a result of the intensity change after rapid weakening. This effect is probably due to insufficient analysis of cold-core eddy used in ocean analysis data and relatively small TC-induced SSC simulated in the atmosphere-wave-ocean coupled model used in this study. A further decrease in $Q_{TCHP}$ ($< 50$ kJ cm$^{-2}$) and $D_{26}$ ($< 50$ m) occurred in the AWO_cold experiment after 0000 UTC on 26 September, which was caused by implanting the artificial cold-core eddy in the initial oceanic condition and Trami-induced SSC. The $Q_{TCHP}$ lower than the threshold of TC intensification (Wada 2015) hindered the overdevelopment of Trami. Therefore, the artificial cold-core eddy under Trami’s track centered at ($21^\circ$N, $129^\circ$E), which resulted in low $Q_{TCHP}$ and shallow $D_{26}$ by promoting Trami-induced SSC, which helped suppress the overdevelopment.

Figure 11 shows the horizontal distribution of TCHP at 1800 UTC on 28 September in experiments
Fig. 11. Horizontal distribution of TCHP at 1800 UTC on 28 September 2018 in experiments AWO_0919 to AWO_0925 and AWO_cold. Red circles indicate the simulated location of the center of Trami at 3 h intervals. In all panels, the OISST product is used as SST at the initial time.
In each experiment, TCHP decreased along the track of the simulated TC south of 20°N and east of 130°E during the intensification phase. This effect was a result of Trami-induced SSC (Figs. 4, 11), which was mainly caused by upwelling and vertical mixing (e.g., Price 1981). However, none of the experiments except AWO_cold simulated the significant magnitude of mesoscale cold-core eddy around (22.5°N, 128.0°E) even though the simulated Trami overdeveloped and gained strength and force in the ocean. This result suggests that the cold-core eddy did not form significantly in experiments AWO_0919 to AWO_0925 as an oceanic response to Trami but already preexisted in the oceanic environment. The decrease in TCHP due to the ocean response to Trami but already preexisted in the oceanic environment. The decrease in TCHP due to the ocean response to Trami was further promoted around (22.5°N, 128.0°E), which helped suppress the overdevelopment, surrounding the artificial cold-core eddy in the AWO_cold experiment. Therefore, the absence of the cold-core eddy with a significant magnitude in the initial oceanic condition possibly led to the overdevelopment after rapid weakening in experiments AWO_0919 to AWO_0925 even though the coupled model could simulate TC-induced SSC to some extent. In fact, the vertical profile of upper-ocean water temperature in the initial oceanic condition affects the magnitude of TC-induced SSC (Wada 2002). In addition, the multilayer ocean model used in the coupled model cannot adequately simulate TC-induced SSC caused by the upwelling particularly when the speed of a TC motion is relatively slow (less than 3 m s⁻¹) (Wada 2002).

Figure 12 shows the results of the backward trajectory analyses in experiments A_0923, AWO_0923 and AWO_cold along with vertical wind shear averaged for 18 h from the 72-h to 90-h integration time. Vertical wind shear was calculated as the difference in wind speeds between the 13th (~1500 m) and 38th (~14000 m) levels in the annulus between 100 km and 300 km from the TC center.

During the analysis period, the simulated Trami was held stationary. In the A_0923 experiment (Fig. 12a), parcels at the center of Trami at the 90-h integration time arrived via two trajectories: 1) from the middle to upper troposphere and 2) from the planetary boundary layer below 1500 m altitude. Colomb et al. (2019) showed in their numerical simulation of TC Hellen (2014) that the area of high equivalent potential temperature at the eye was separated into levels above 300 hPa and below 900 hPa. The first trajectory from the middle to upper troposphere indicated subsidence of dry air from the middle to upper troposphere at the center of Trami. Therefore, adiabatic heating contributed to the formation of lower-tropospheric warm core (Willoughby 1998) and thereby simulated intensity of Trami. The second trajectory below 1500 m altitude

AWO_0919 to AWO_0925 and AWO_cold (Table 1).

Fig. 12. Results of back trajectory analyses in the (a) A_0923, (b) AWO_0923, and (c) AWO_cold experiments. Colors indicate specific humidity at the trajectory point at that time. Arrows indicate speed and azimuth of vertical wind shear averaged from the 72-h to 90-h integration time.
indicated that moist air near the surface and in the inflow boundary layer was spirally transported toward the TC center due to surface friction, which is consistent with the inflow boundary layer process (Wada and Oyama 2018; Wada et al. 2018). Figure 12a indicates that near-surface specific humidity in the A_0923 experiment was higher than the other two experiments. The result suggests that the moisture transported in the A_0923 experiment was favorable for overdevelopment of simulated TC under a relatively small vertical wind shear (0.32 m s$^{-1}$).

In the AWO_0923 experiment (Fig. 12b), the relatively dry air at the center of Trami was at a lower altitude than in the A_0923 experiment, implying that adiabatic heating became weak in the lower-tropospheric warm core under the small vertical shear (0.82 m s$^{-1}$), and thereby the intensity of Trami was weaker in this experiment. There was relatively low specific humidity spiraling toward the center and near the surface due to surface friction and in the inflow boundary layer because of the effect of Trami-induced SSC on reduction in air–sea sensible and latent heat fluxes. This same pattern appeared in the AWO_cold experiment (Fig. 12c), where the height of the dry air parcel at the center of Trami was low under the small vertical shear (1.84 m s$^{-1}$) and the area of low specific humidity near the surface was large in the southeast quadrant. The number of air particles transported from the northern quadrant where the artificial cold-core eddy existed was less and the area in which air particles existed was also narrower than in the AWO_0923 experiment.

Figure 13 compares the time series of sea-to-air net heat flux at the air-sea interface within a 50 km radius of the TC center and in the 50–150 km annulus for experiments A_0923, AWO_0923, AWO_cold, and AWO_mean (which averaged the net fluxes from experiments AWO_0919 to AWO_0925 and is shown with error bars representing one standard deviation). The net heat flux is the sum of shortwave and longwave radiation and sensible and latent heat fluxes. In the case of Trami, net heat flux was highest in the A_0923 experiment even in the rapid weakening and subsequent mature phases and clearly reduced in the AWO_cold experiment in these phases within the 50 km radius (Fig. 13a); whereas, the experiments differed little in the 50–150 km annulus except in the A_0923 experiment. In the AWO_0923 and AWO_mean experiments, net heat flux was reduced within the 50 km radius from that in the A_0923 experiment from 24 September although the value was relatively high when compared to the results in the AWO_cold experiment. It should be noted that the rapid weakening of Trami from 25 September could be reasonably simulated even without the cold-core eddy in the AWO_0923 experiment. This result shows that the presence of the cold-core eddy, which has significantly low TCHP, can affect the sea-to-air net heat flux at the air–sea interface, which is important in assessing intensity changes of Trami after rapid weakening through the reduction in air–sea sensible and latent heat fluxes due to TC-induced SSC and thereby reduction in specific humidity near the surface spiraling toward the TC center due to surface friction. The importance of air–sea sensible and latent heat fluxes and surface friction is based on an intensification mecha-
nism of wind-induced surface heat exchange (WISHE) instability (Emanuel 1995; Craig and Gray 1996; Zhang and Emanuel 2016). The overdevelopment of simulated Trami in the A_0923 experiment and its suppression in experiments AWO_0923 and AWO_cold can be explained by the amount of air–sea heat fluxes and inward transport of moisture toward the TC center through the secondary circulation of TC, which are essential mechanisms of WISHE instability.

Figures 14a–c show the horizontal distribution of simulated SST with sea-level pressures at the 84-h integration time in experiments A_0923, AWO_0923, and AWO_cold.
and AWO_cold. In the A_0923 experiment, SST around the TC center was higher than 29°C; whereas, SST underneath the TC was approximately 27°C in the AWO_0923 experiment. Additionally, TC-induced SSC around the TC center from the downshear-left to the upshear side (Fig. 12c) was enhanced by the artificial cold-core eddy in the AWO_cold experiment because the area of the SSC on the right-hand side of the TC track was closer to the cold-core eddy. This result suggests that the artificial cold-core eddy enhances TC-induced SSC.

Figures 14d–f show the horizontal distribution of simulated equivalent potential temperature at the height of approximately 1.5 km at the 84-h integration time. Equivalent potential temperature exceeded 372 K around the TC center and 360 K in the inner core of simulated TC in the A_0923 experiment. In the AWO_0923 experiment, the equivalent potential temperature exceeded 372 K around the TC center became small when compared to the result in A_0923 experiment. Thus, the area of the equivalent potential temperature, exceeding 360 K became small in the inner core. Furthermore, in the AWO_cold experiment, there was no area exceeding 372 K around the TC center, and the area exceeding 360 K in the inner core was only found in the southern quadrant of TC. These results are consistent with those of the backward trajectory analysis in view of the reduction in the amount of moisture near the surface transported toward the TC center.

Figures 14g–i show the cross section of simulated equivalent potential temperature averaged within the radius of 20 km below an altitude of approximately 2 km at the 84-h integration time. The direction of the cross section was in parallel with the direction of vertical shear averaged from the 72-h to 90-h integration time. In the A_0923 experiment, air with low equivalent potential temperature was not flushed downward far enough to enter the inflow boundary layer on the northern (upshear) side (Fig. 14g). In contrast, air with low equivalent potential temperature was efficiently flushed downward into the relatively stable inflow boundary layer on the northern (upshear) side in the AWO_cold experiment (Fig. 14i). The low equivalent potential temperature flushed downward possibly fosters rapid weakening of the simulated Trami (Colomb et al. 2019; Ma et al. 2020) in addition to decreases in moisture transport near the surface due to surface friction and in the inflow boundary layer.

3.3 Role of the ocean in rapid weakening of simulated Kong-Rey

This subsection reports our investigation of the oceanic thermodynamic environment underneath simulated Kong-Rey in terms of TCHP. Table 2 is a summary of numerical simulations. $Q_{\text{TCHP}}$ and $D_{26}$ rapidly decreased after 1200 UTC on 1 October in the experiments OISST, MGDSST, OISST_COLD, MGDSST_COLD, A_OISST, and A_MGDSST. However, the simulated central pressure and maximum surface wind speed did not reproduce the rapid
intensification nor rapid weakening analyzed in the RSMC best track data (Fig. 9). Instead, the differences in SST data at the initial time affected the simulated QTCHP and \( D_{26} \) values, and thus the simulated central pressure and maximum surface wind speed. The experiments using the MGDSST dataset showed a tendency that the simulated TCs were stronger than those using the OISST dataset irrespective of ocean coupling. In the case of Trami, the result is different. This suggests that the impact of TC-induced SSC and the embedded artificial cold-core eddy on the intensity simulation was different between Trami and Kong-Rey. In the case of Kong-Rey, the rapid weakening was not simulated realistically even when TCHP underneath the TC was almost zero. Kong-Rey induced SSC was smaller than Trami-induced SSC (Fig. 3). In fact, TCHP was already lowered by passage of Trami, which led to small TC-induced SSC by Kong-Rey. Under this situation, changes in the systematic difference in the SST between MGDSST and OISST could be greater than the difference in the TCHP resulting from the ocean response to Kong-Rey or the hypothetically implant of artificial cold-core eddy. Figure 16 shows the time series of net heat flux at the air–sea interface at the ocean the atmosphere (a) within a radius of 50 km from the center of Kong-Rey and (b) in annulus from 50 km to 150 km from the center of Kong-Rey in the experiments OISST, MGDSST, OISST_COLD, MGDSST_COLD, A_OISST, and A_MGDSST.

Fig. 16. Time series of net heat flux at the air–sea interface from the ocean to the atmosphere (a) within a radius of 50 km from the center of Kong-Rey and (b) in annulus from 50 km to 150 km from the center of Kong-Rey in the experiments OISST, MGDSST, OISST_COLD, MGDSST_COLD, A_OISST, and A_MGDSST.

The uncoupled NHM was relatively small in the weakening phase, even though the intensity simulated by the uncoupled NHM was greater in the coupled model. The results suggest that the rapid weakening is not determined only by the net heat flux at the air–sea interface. In other words, the combined or separate use of the artificial cold-core eddy having significantly low TCHP not based on in situ observations and the coupled model did not enable the rapid weakening of Kong-Rey to be successfully reproduced, which appears to have been different from Trami. It should be noted that the simulated tracks shown in Fig. 9a are different from the RSMC best track data over the cold-core eddy around (22.5°N, 128°E). Does SST rather than TCHP really affect the intensity changes in the case of Kong-Rey, even when the track simulation is close to the RSMC best track?

Figure 17 shows the results of nine ensemble simulations based on the OISST_COLD experiment (OISST_COLD_001 to OISST_COLD_009). In Fig. 17a, the nine simulated tracks have a large spread owing to the differences in their initial atmospheric conditions. Nevertheless, none of nine ensemble simulations reproduced the minimum central pressure (Fig. 17b) and few reproduced the maximum wind speed (Fig. 17c) of Kong-Rey; some members simulated rapid weakening after 0600 UTC on 2 October to some extent, whereas other members still produced overdevelopment. For example, the experiments OISST_COLD_002 and OISST_COLD_005 produced overdevelopment when their simulated TCs moved on the south side of the track in the OISST_COLD experiment. The OISST_COLD_006 experiment produced...
a track that was closer to the RSMC best track than the track in the OISST_COLD experiment (Fig. 17a).

Figure 18 shows the distribution of TCHP at 1200 UTC on 3 October and the simulated positions every 3 h in the experiments OISST_COLD (Fig. 18a) and OISST_COLD_006 (Fig. 18b). The location and size of the area in which TCHP was lower than 20 kJ cm$^{-2}$ demonstrated little difference in the two experiments, although the simulated tracks were different over that area. The TC-induced SSC around (21°–24°N, 127°–131°E) was approximately 2.6°C in the OISST_COLD experiment and approximately 1.8°C in the OISST_COLD_006 experiment. This result suggests that low TCHP underneath the TC in the OISST_COLD_006 simulation resulted not from the ocean response to the TC, but from initial oceanic conditions determined by the TC location. It should be noted that the value of SST underneath the TC was relatively low in the OISST_COLD_006 experiment (approximately 26.0°C) compared to that in the OISST_COLD experiment (approximately 26.6°C). The amount of TC-induced SSC varies directly with the TC intensity; when the simulated TC intensity under relatively low SST underneath the TC is weaker, the TC-induced SSC is smaller. The influence of initial oceanic conditions on TC-induced SSC is rather small in the case of Kong-Rey.

Figure 19 shows the results of backward trajectory analyses in experiments A_OISST, OISST_COLD and OISST_COLD_006 along with vertical wind shear averaged from the 72-h to 90-h integration time. In contrast to Trami’s case, the averaged vertical wind shear was calculated in the annulus between 300 km and 500 km from the TC center. Unlike the case of Trami (Fig. 12), air parcels at the center of simulated Kong-Rey at 0600 UTC on 3 October were generated only from the boundary layer below 1500 m altitude in the experiments OISST_COLD and OISST_COLD_006.
In other words, adiabatic heating caused by dry subsidence appeared above 1500 m only in the A_OISST experiment. In the OISST_COLD experiment, the effect of adiabatic heating on increasing the lower-tropospheric warm core was suppressed, but the amount of moisture transport near the surface due to surface friction was not reduced compared to the results in the A_OISST experiment. In fact, moist air with specific humidity greater than 24 g kg\(^{-1}\) appeared on the south side (downshear-left side) in the OISST_COLD experiment (Fig. 19b) and was transported to the upshear-left side, whereas specific humidity
became relatively low from the downshear-left to upshear-left side in the OISST_COLD_006 experiment (Fig. 19c). This result conversely supports the importance of lower-tropospheric moisture (Rios-Berrios et al. 2016a) and left-of-shear convection (Rios-Berrios et al. 2016b) on TC intensity changes in the intensification phase. In addition, the averaged vertical wind shear was greater in the OISST_COLD_006 experiment (5.56 m s$^{-1}$) than in the OISST_COLD experiment (3.23 m s$^{-1}$) although both values were greater than the averaged vertical wind shear of Trami.

Figure 20 compares the two experiments in a time series of net heat flux at the air–sea interface within the 50 km radius and in the 50–150 km annulus. Unlike the case of Trami (Fig. 13), the difference in net heat fluxes appeared in both areas, which is consistent with the result of backward trajectory analysis showing that the horizontal distribution of air parcels is wider for Kong-Rey (Fig. 19) than for Trami (Fig. 12). The difference in net heat fluxes between the experiments is caused not by the difference between MGDSST and OISST and that in initial oceanic conditions but by the difference in the simulated tracks, TC size, and oceanic conditions where the simulated TC exists. In other words, insufficient TC weakening was associated with relatively low SST, small TC-induced SSC, and air-sea heat flux regardless of low TCHP. In the OISST_COLD_006 experiment, the stronger TC-induced SSC and reduction in sea-to-air net heat fluxes were, the more pronounced was the TC weakening.

The reasons for the rapid weakening of Kong-Rey in the actual atmospheric environment can be accounted by its size (Fig. 19) and its track passing over the relatively low TCHP area caused by passage of Trami (Figs. 3, 18). Because the low TCHP region is relatively narrow along the track of Trami (Figs. 3, 9) due to relatively small size (Fig. 12), slight deflection of the Kong-Rey track can make a substantial difference in the SST and TCHP along the track unless Kong-Rey-induced SSC occurs significantly. With regard to the track, the relatively fast translation speed of Kong-Rey when compared to translation speed of Trami affected the variation in oceanic response to the TC. Figures 21a, b show the horizontal distribution of simulated SST with sea-level pressures at the 84-h integration time in the experiments OISST_COLD and OISST_COLD_006. The TC center was far from the location of artificial cold-core eddy and the simulated SST around the TC center was approximately 26°C in the OISST_COLD experiment (Fig. 21a). Since the track simulation was closer to the best track in the OISST_COLD_006 experiment, the location of the TC center was closer to the artificial cold-core eddy. Although TC-induced SSC in the OISST_COLD_006 experiment was smaller than the results in the OISST_COLD experiment, the decrease in SST appeared on the downshear side (on the right-hand side of the track) was greater in the OISST_COLD_006 experiment than around the TC center in the OISST_COLD experiment. Since simulated Kong-Rey traveled over the area at relatively low latitudes and passed through an area where the SST was relatively high, overdevelopment occurred in the OISST_COLD experiment. In other words, the relatively warm SST field over the track of simulated Kong-Rey and small TC-induced
SSC possibly hindered the rapid weakening.

Figures 21c, d show the horizontal distribution of equivalent potential temperature at the height of approximately 1.5 km at the 84-h integration time. The equivalent potential temperature within the TC inner core region was higher in the OISST_COLD experiment than that in the OISST_COLD_006 experiment, which is consistent with the result of backward trajectory analysis (Fig. 19). The reduction in moisture transport near the surface and in the inflow...
boundary layer toward the TC center over the area of TC-induced SSC results in low equivalent potential temperature around the TC center in the OISST_COLD_006 experiment.

Figures 21e, f show the cross section of equivalent potential temperature below approximately 2 km altitude at the 84-h integration time, which is an averaged diameter of 20 km. In the OISST_COLD_006 experiment, the size of the relatively high equivalent potential temperature in the inner core and in the inflow boundary layer was relatively small when compared to the results in the OISST_COLD experiment. In fact, air with low equivalent potential temperature can be more efficiently flushed downward into the relatively stable inflow boundary layer (Lee and Chen 2014) on the upshear side in the OISST_COLD_006 experiment (Fig. 21f), whereas the stable layer is formed by TC-induced SSC and the associated reduction in moisture fluxes near the surface and in the inflow boundary layer spiraling toward the TC center.

4. Summary and discussion

This study conducted numerical simulations with a 2-km-mesh nonhydrostatic atmosphere model and an atmosphere-wave-ocean coupled model to investigate the possible impacts of an oceanic mesoscale cold-core eddy on the rapid weakening of Typhoons Trami and Kong-Rey in 2018. The magnitude of the artificial cold-core eddy is hypothetically determined by assuming a significant magnitude with a minimum TCHP value of approximately 42 kJ cm$^{-2}$ that potentially affects the intensity simulation of Trami because of sparse in situ observations, utilizing profiling floats. The purpose of this study is to understand the potential roles of an artificial cold-core eddy on the intensity change of Trami and Kong-Rey in their simulations particularly during the rapid weakening and subsequent mature phases of Trami and the rapid weakening phase of Kong-Rey, utilizing a 2-km-mesh nonhydrostatic atmosphere model and an atmosphere-wave-ocean coupled model. Additionally, this study explores the similarity and difference of the roles between simulated Trami and Kong-Rey.

The results of numerical simulations, sensitivity experiments, and backward trajectory analyses support the following conclusions.

1. Rapid weakening of Trami is reasonably simulated by the coupled model. However, most simulations of Trami show a tendency of overdevelopment during the mature phase. The overdevelopment of simulated Trami is caused by insufficient TC-induced simulated SSC around (21°N, 129°E), which is different from the location around (22.5°N, 128°E) because the simulated TC track shifts eastward at 21°N. From the result of sensitivity simulation with an artificial cold eddy at (21°N, 129°E) embedded in the oceanic initial condition having a significant magnitude of TCHP of approximately 42 kJ cm$^{-2}$, ~30 kJ cm$^{-2}$ lower than the original analysis, the decrease in TCHP due to the ocean response to Trami is further promoted around the artificial cold-core eddy, which helped suppress the overdevelopment in the simulation. Even though the significant magnitude of artificial cold-core eddy is not based on in situ observations, utilizing profiling floats, the absence of the cold-core eddy with significantly low TCHP in the initial oceanic condition possibly leads to overdevelopment after the rapid weakening of Trami.

2. Unlike Trami’s case, overdevelopment of Kong-Rey during the rapid weakening and subsequent mature phases cannot be suppressed even when the same artificial cold-core eddy as Trami’s experiment is embedded in the initial oceanic condition. The simulated Kong-Rey traveled over the Trami-enhanced cold-core eddy area, resulting in a relatively small TC-induced SSC. Insufficient TC weakening simulated by the coupled model is associated with a relatively small TC-induced SSC and a reduction in sea-to-air net heat fluxes rather than low TCHP beneath the simulated Kong-Rey. The stronger the SSC and reduction in sea-to-air net heat fluxes are due to the improvement of track simulation, the more pronounced is the weakening of simulated Kong-Rey. Low TCHP, small TC-induced SSC, and reduction in sea-to-air net heat fluxes are related to the amount of moisture transported near the surface and in the inflow boundary layer in the TC’s inner core. Possible mechanisms through which a significant cold-core eddy may affect the intensity changes of both Trami and Kong-Rey during the mature or weakening phase are noted in the following. The reduction in moisture near the surface spiraling toward the TC center is caused by TC-induced SSC, which is promoted by the cold-core eddy with significantly low TCHP in the initial oceanic condition particularly in the case of Trami. Air with low equivalent potential temperature in the lower troposphere is more efficiently flushed downward into the relatively stable inflow boundary layer on the upshear side, where the stable layer is formed by TC-induced SSC and the associated reduction in moisture transport near the surface due to surface friction and in the inflow boundary layer. This
result is related to rapid weakening of both Trami and Kong-Rey.

This study identified three factors—mesoscale cold-core eddy in the initial oceanic condition, TC-induced SSC, and errors of the TC track—that affected the intensity change of Trami and Kong-Rey particularly during the rapid weakening and subsequent mature phases of Trami and the rapid weakening phase of Kong-Rey. These three factors are consistent with the results of Chen et al. (2017) who examined the impact of ocean observations on atmosphere-ocean coupled model forecasting errors of Hurricanes Isaac (2012), Hilda (2015), and Matthew (2016). Researchers have addressed the first factor through improvements in the oceanic analysis system from innovative data assimilation practices, advances in observation technology, and the increase in satellite data and associated datasets (e.g., Domingues et al. 2019; Tomita et al. 2019; Wada et al. 2020). Halliwell et al. (2017) conducted the ocean observing system simulation experiments to investigate the impact of ocean observation data on oceanic analysis fields, such as TCHP and thermal stratification, for TC prediction in the air–sea coupled TC prediction system. The results suggested observational approaches to improve such factors as dynamic structure and thermal stratification.

However, such innovations are not necessarily reflected in the TC prediction system in the western North Pacific due to the following physical processes related to SST variations. In a cold-wake area, the water temperature at the air–sea interface increases to some extent by the input of solar radiation after the passage of a TC; thus, the upper ocean around the cold-core eddy appears to become warm. Beneath this warm ocean skin, the cold-core structure remains intact so that TC-induced SSC easily recurs when a subsequent TC passes over the area (Lin et al. 2008; Ma et al. 2018). Therefore, overdevelopment of a TC is predicted, particularly when the uncoupled NHM is used, because SST remains high during the simulation. Therefore, an atmosphere-ocean coupled model with greater accuracy of initial oceanic conditions should be used for TC prediction (Ito et al. 2015).

Satellite sea surface height (anomaly) data are indispensable for ocean data assimilation and daily ocean data analysis; however, they are not sufficient to determine the three-dimensional oceanic structure. Particularly in deep waters, it is not simple to separate this anomaly in the upper ocean from anomalies in the deeper ocean. To obtain the three-dimensional oceanic structure, it is necessary to increase the frequency of in situ observations (i.e., profiling floats) and to rede-sign observational strategies based on the results of ocean observing system simulation experiments. Increasing the frequency of profiling float observations also makes it possible to capture the ocean response to TCs in greater detail (Wada et al. 2014).

In this study, we used ensemble sensitivity experiments for initial oceanic conditions (for Trami and Kong-Rey) and for initial atmospheric conditions (for Kong-Rey) to evaluate the uncertainty of simulation results. Previous sensitivity experiments for initial oceanic conditions have suggested that oceanic variability on a weekly time scale affected the intensity simulations of TCs Hai-Tang (2005) (Wada and Usui 2010), Choi-wan (2009) (Wada et al. 2013), and Haiyan (2013) (Wada et al. 2018). In addition, sensitivity to initial oceanic conditions has been clearly detected in the area of secondary circulation (Wada et al. 2018). However, the aforementioned method of ensemble sensitivity experiments for initial oceanic conditions does not change the initial atmospheric condition. In that sense, ensemble sensitivity experiments for initial atmospheric conditions are useful for investigating the uncertainty of TC intensity simulations, particularly when the TC track forecasting has failed.

Sensitivity ensemble experiments for initial atmospheric conditions were used to study the method of ensemble sensitivity analysis (Ren et al. 2019), predictability of TC track and intensity (Nystrom et al. 2018), TC intensification and non-intensification (Judt et al. 2016; Munsell et al. 2013, 2017; Leighton et al. 2018), TC inner-core structure during rapid intensification (Judt and Chen 2016; Munsell et al. 2018), and sensitivity of rapid intensification to deep-layer vertical wind shear (Rios-Berrios et al. 2016a, b; Tao and Zhang 2019). However, relatively few studies have addressed the impact of TC–ocean interactions on the uncertainty of TC simulations by using an atmosphere-ocean coupled model with perturbed initial atmospheric conditions. Furthermore, previous studies have not addressed the rapid weakening phase of TCs except the study discussed in Ma et al. (2020). In this study, we used ensemble sensitivity experiments for initial atmospheric conditions only to demonstrate the role of TC track simulation in the simulated intensity for Kong-Rey although we conducted various sensitivity experiments for initial oceanic conditions for both Trami and Kong-Rey. In the future, by increasing the number of TC cases and ensemble members and by conducting a cluster and composite analysis similar to the study noted by Takamura and Wada (2020), it will be possible to evaluate the contribution of initial
oceanic conditions and TC structural change in midlatitude for greater accuracy in TC intensity simulations.

Supplements

Table S1 shows the results of Trami’s intensity guidance in the sensitivity experiments regarding TCHP by using TIFS. The initial time is 0000 UTC 26 September in 2018. Figure S1 shows the time series of TCHP from 19 September to 5 October 2018 obtained from a few oceanic analysis products and Text S1 explains the results of sensitivity experiments regarding TCHP by using TIFS and timeseries of TCHP from 19 September to 5 October 2018 obtained from a few oceanic analysis products.

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