1. Introduction

Typhoon Lionrock, the tenth typhoon in the western North Pacific of the 2016 tropical cyclone (TC) season, moved along a counterclockwise track after it became the tropical storm from the tropical depression south of Japan on 21 August. After undergoing recurvature due to the proximity of the mid-latitude jet, Lionrock moved north-northeast and then changed the direction north-northwest. It made landfall in the Pacific side of northern Japan on 31 August, which was recorded as the first typhoon landfalling in the region since 1951 when the statistics of typhoons started. In fact, Lionrock caused torrential rain and serious damage in the wide area of northern Japan.

The mechanism of changes in TC intensity just before the landfall is of great concern to East and Southeast Asian countries (Peduzzi et al. 2012; Mei and Xie 2016). The increase in the intensity of TCs is closely associated with the increasing of sea surface temperature (SST) in the coastal ocean (Mei and Xie...
Kossin et al. (2014) indicated that the average location of TC maximum intensity shifts poleward over the last 30 years, and it is related to changes in vertical wind shear and potential intensity (Bister and Emanuel 1998). In addition, SST increased significantly east of northern Japan from 1900 to 2013 (http://www.jma.go.jp/jma/en/Activities/cc.html). The long-term increase in SST may have contributed to heavy rainfall associated with TCs approaching Japan in recent years. Typhoon Man-yi (2013), for example, caused extraordinarily heavy rainfall in the northern Kinki districts, because of preexisting warm-water conditions (Wada 2015).

Using the results of numerical simulations by an atmosphere-wave-ocean coupled model, Wada (2015) reported that Man-yi intensified north of 30°N on the vortex scale of a few hundred kilometers. The intensification process differed from the so-called axisymmetrization that was found in the axisymmetrized inner-core structure of extremely intense TCs projected under global warming (Kanada et al. 2013; Tsuboki et al. 2014). Rather, the processes were explained by downshear reformation, with the asymmetric distribution of convection represented by diabatic heating due to mesovortices that helped reducing the vortex tilt (Nguyen and Molinari 2015). However, the numerical studies on TC-ocean interactions with a coupled atmosphere-wave-ocean model (Lee and Chen 2014; Wada et al. 2014; Wada 2015) indicated that the convection could be suppressed by excessive storm-induced sea surface cooling (SSC) (Price 1981). Even so, it is unclear how and to what extent the asymmetric local convection affects the intensity of TCs through or against the oceanic response to a storm that varies depending on the preexisting oceanic conditions, such as mesoscale ocean eddies (Ma et al. 2013, 2017, 2018).

Lionrock reached the highest intensity at 0600 UTC on 28 August. After recurvature, the SSC clearly appeared on the right side of its track (Fig. 1). The geostationary satellite Himawari-8 (Bessho et al. 2016) observed consecutive deep convections, usually CBs (Riehl and Malkus 1961; Steranka et al. 1986), around Lionrock on 30 August during its decay phase, just before it made landfall in the Pacific side of northern Japan (Fig. 2). The Regional Specialized Meteorological Center Tokyo (RSMC Tokyo) of the Japan Meteorological Agency (JMA) reported no change in the storm intensity during CBs in the best-track dataset. In other words, the decay phase was paused temporarily

Fig. 1. Horizontal distribution of TRMM/TMI daily mean SST on 30 August 2016. Colors indicate water temperature (°C). Circles indicate the location and central pressure of Typhoon Lionrock.

Fig. 2. Horizontal distribution of the Himawari–8 infrared channel (Band 13: 10.4 μm) brightness temperature around Lionrock at 0730 UTC on 30 August 2016.
due to CBs.

To clarify the influence of CBs on changes in the intensity of Lionrock, numerical simulations with a 3 km coupled atmosphere-wave-ocean model were performed, and they are discussed in this work. The focus was on the processes associated with CBs and intensity changes during the decay phase under the condition in which SSC occurred. In the present study, simulated maximum surface wind speed (MWS) is defined at the lowermost level (20 m height) of the nonhydrostatic atmosphere model and the coupled model used in this study. Both MWS and central pressure are used as storm intensity indicators.

The main part of this paper is organized as follows. Section 2 explains the numerical model, experimental design, analysis method, and used datasets of SST and the significant wave heights. Section 3 describes the results of numerical simulations on Lionrock and the ocean response. This section also shows the results of analyses regarding CBs and changes in the intensity. Results and analyses of the sensitivity numerical experiments regarding the oceanic conditions are shown in Section 4. Section 5 discusses the role of the ocean in CBs. Section 6 is devoted to concluding remarks.

2. Data and method

2.1 Model

The atmosphere-wave-ocean coupled model was developed in the JMA Meteorological Research Institute (MRI) (Wada et al. 2010; Wada 2015). The components of the coupled model are a nonhydrostatic atmospheric model (NHM) (Saito 2012), a third-generation ocean surface-wave model (Wada et al. 2010) and a multilayer ocean model developed in the MRI based on Bender et al. (1993). In Section 4, the NHM alone is used to examine the sensitivity to oceanic conditions.

The NHM includes various physical processes, such as cloud microphysics, surface and atmospheric boundary layer, and radiation. The cloud microphysics is expressed as an explicit three–ice bulk microphysics scheme based on the works of Ikawa and Saito (1991) and Lin et al. (1983). No cumulus parameterization is used in this study. An assumed resistance law is applied to calculate air-sea momentum, sensible and latent heat (enthalpy) fluxes in the atmospheric surface boundary layer. The exchange coefficients for air-sea momentum and enthalpy transfers over the sea are calculated based on bulk formulas (Kondo 1975) or, when the ocean-wave model is coupled, they are calculated considering the roughness lengths proposed by Taylor and Yelland (2001). The surface boundary layer also includes a sea spray formulation provided by the work of Bao et al. (2000), which was based on the work of Fairall et al. (1994). A turbulent closure model formulated from the work of Klemp and Wilhelmson (1978) and Deardorff (1980) is used as an atmospheric boundary layer scheme. An atmospheric radiation scheme is formulated based on the work of Sugi et al. (1990).

The third-generation ocean surface-wave model (Japan Meteorological Agency 2013) predicts wave spectra as a function of space and time from an energy balance equation composed of the spectral energy input by the wind, nonlinear transfer of spectral energy due to wave–wave interactions, and dissipation of energy due to breaking surface waves and white–cap formation. The wave spectrum in the third-generation ocean-wave model has 900 components, each associated with one of 25 frequencies and one of 36 directions. The frequency of the wave spectrum is divided logarithmically from 0.0375 to 0.3000 Hz. The oceanic surface wave is assumed to be stationary at the initial time.

The multilayer ocean model used in this study uses reduced gravity and hydrostatic approximations, and it is assumed that the water is a Boussinesq fluid (Wada 2015). The model has three layers and four interfaces at maximum. The uppermost layer is a mixed layer, where the density is uniform in the vertical direction. The middle layer is the seasonal thermocline, where the vertical temperature gradient is the greatest. The bottom layer is assumed to be undisturbed by entrainment. Note that it is not necessary in each grid to have the three layers, particularly when the sea bottom is shallow. The number of layers, interfaces, and layer thickness depends on the topology of a grid, so that the reduced gravity approximation is applied for each grid. The four levels are the sea surface, base of the mixed layer, base of the thermocline, and sea bottom at maximum. The model calculates the water temperature and salinity at the surface and at the base of the mixed layer, and the thickness of each layer and the two–dimensional flows in the layers. The degree of entrainment is calculated at the base of the mixed layer by using the multi–limit entrainment formula proposed by Deardorff (1983). The skin SST can be calculated in the multilayer ocean model based on the work of Wada and Kawai (2009). The skin SST is defined as a temperature within the conductive diffusion dominated sub-layer at a depth of approximately 10–20 μm (Kawai and Wada 2007).

The following processes denote the exchanges between the atmosphere, ocean surface, and
multilayer ocean models. Short- and long-wavelength radiation, air-sea sensible and latent heat fluxes, wind stresses, cloud amounts, and precipitation were calculated by the atmospheric model and supplied to the ocean model at every time step in the ocean model. The SST calculated by the ocean model was provided to the atmospheric model at every time step in the ocean model. Surface wind speeds calculated by the atmosphere model were provided to the ocean surface-wave model at every time step of the ocean surface-wave model. Conversely, significant wave heights calculated by the ocean surface-wave model were provided to the atmosphere model to calculate the steepness of the ocean wave for estimating the surface roughness lengths (Wada et al. 2010, 2013a). The ocean currents at the uppermost layer calculated in the ocean model were provided to the ocean surface-wave model at every time step of the ocean surface-wave model. The group velocity of the ocean waves calculated by the ocean model was modified by the ocean currents in the uppermost layer. Contrariwise, the wave–induced stresses calculated by the ocean surface-wave model were provided to the ocean model to modify the surface wind stress. The wave–induced stress was also used for estimating the entrainment rate for breaking surface waves (Wada et al. 2010).

2.2 Experimental design

The horizontal resolution of the NHM and the atmosphere-wave-ocean coupled model was 3 km with a computational domain of 3000 × 3000 km centered at 33.0°N, 137.0°E. There were 55 vertical levels with the intervals ranging from 40 (for the near-surface layer) to 1180 m (for the uppermost layer). The top height was approximately 27 km. A schematic diagram of the experimental design is shown in Fig. 3. Because both the NHM and the coupled model have no scheme for vortex initialization, the integration of Lionrock has to be started at an earlier intensification phase of a storm or before. Therefore, the initial time was 0000 UTC on 23 August 2016, and the integration time was 210 h. Time steps were 3 s in the NHM, 18 s in the multilayer ocean model, and 6 min in the third-generation ocean surface-wave model. Note that the time step of the advection term in the third-generation ocean surface-wave model was 36 s with the iteration of 10 times within 6 min. As a rule, the time step of the ocean surface-wave model was relatively long due to its high computational cost.

The atmospheric initial and boundary conditions were created from JMA 6-hourly global atmospheric analysis data with a horizontal grid spacing of 20 km. The oceanic initial conditions were created from daily oceanic analysis data in the North Pacific with a horizontal grid spacing of 0.5° calculated with the MRI Ocean Variational Estimation (MOVE) system (Usui et al. 2006). The initial depth of the oceanic mixed layer was determined from oceanic reanalysis data by defining the mixed layer as a layer in which the density was no more than 0.25 kg m$^{-3}$ higher than the density at the surface, and having a maximum depth of 200 m (Wada et al. 2014). The base of the thermocline was limited to 600 m, whereas the water depth was limited to 2000 m.

The experimental acronyms are also shown in Fig. 3. The ‘CPL’ experiment was performed only with the atmosphere-wave-ocean coupled model, whereas the ‘FIX’ experiment was performed only with the NHM. The ‘SEN’ experiments consisted of the following two parts. First, a numerical simulation was performed with the coupled model from the initial time to 2100 UTC on 29 August. The following simulation was performed with the NHM started from 2100 UTC on 29 August. The ‘SEN’ experiment was conducted to understand the effects of storm-induced SSC simulated in the CPL experiment on the relationship between CBs and changes in intensity of the storm. In this study, the CPL experiment is considered as a control experiment while FIX and SEN experiments are considered as sensitivity experiments.

2.3 Trajectory analysis

To elucidate the impact of atmospheric conditions on CBs, a backward trajectory analysis was conducted for each parcel. The modeled trajectory was calculated with a second-order Runge-Kutta method. Thirty parcels were released around 37.1°N, 142.6°E from heights of 0.02, 0.2, 0.5, 1.0, 1.25, 1.5, 1.75, 2, and 3 km where the maximum surface wind speed was
simulated. The starting time was 0300 UTC on 30 August, and the start of the integration time was 171 h. The time step was 6 s. For each level, 30 trajectories were traced backward for six hours.

2.4 Data

a. Satellite data

This study used the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) daily SST product with a horizontal resolution of 0.25° (http://www.remss.com/) to show a horizontal distribution of SST (Fig. 1). The data do not include SSTs measured by infrared sensors. In addition, the Group for High Resolution Sea Surface Temperature (GHRSST) Project Data Processing Specification version 2 format specifications (GDS2.0) version 2.1 Level 2P (L2P) Global Skin Sea Surface Temperature from the Advanced Microwave Scanning Radiometer 2 (AMSR2) on the Global Change Observation Mission-Water “Shizuku” (GCOM-W) satellite (http://suzaku.eorc.jaxa.jp/GHRSST/index.html) were used to compare the satellite SSTs with the simulated SSTs at 0000 UTC on 31 August 2016. Hereafter, this SST result is called ‘AMSR2 L2 SST’. The time, 0000 UTC, on 31 August 2016, was selected, because no AMSR2 L2 SST data around the storm during the passage (on 30 August) are available due to the thick clouds.

b. Coastal ocean surface-wave model

Data calculated by coastal ocean surface-wave model operationally used in JMA were considered to compare with the CPL-simulated significant wave heights at 0000 UTC on 30 August, 2016. This is a forecast dataset, but it also includes the data at the initial time, which were used for comparative purposes. The horizontal resolution was 0.5°, and the computational domain covered the area around 20–50°N and 120–150°E.

3. Results

3.1 Track, intensity, and CBs of Lionrock

The track of Lionrock was simulated quite reasonably in the CPL experiments (Fig. 4a) compared with RSMC Tokyo best track. The moving speed of CPL-simulated Lionrock was approximately 8.3 (0000 UTC) – 10.4 m s$^{-1}$ (0600 UTC). The amplitude of vertical wind shear, calculated as the difference of horizontal wind speeds between 200 and 850 hPa, was 7.2 m s$^{-1}$, which fits well the favorable range for TC intensification (Yu et al. 2013). After 1800 UTC on 28 August, during the decay phase, the best-track MWS, defined as the 10 min averaged wind speed at 10 m height, decreased monotonically, except from 0000 to 0800 UTC on 30 August. The CPL experiment successfully simulated the monotonic decrease in MWS.

Fig. 4. (a) Best track of Typhoon Lionrock and CPL-simulated tracks during the integration. The arrow indicates a vertical wind shear vector between 200 and 850 hPa at 0300 UTC on 30 August in 2016. (b) Time series of best-track MWS and simulated maximum surface wind speeds at 10 m height. (c) Time series of best-track central pressure and simulated central pressures.
From 0800 to 1200 UTC on 30 August, the best-track MWS rapidly decreased while a subsequent increase in the CPL-simulated MWS occurred around 1200 UTC. The best-track central pressure retained the value of 965 hPa from 0000 to 0600 UTC on 30 August (Fig. 4c). The CPL-simulated central pressures were very similar to those of the RSMC Tokyo best-track data.

An asymmetric distribution of horizontal wind speeds exceeding 20 m s\(^{-1}\) at 20 m height was simulated at 0300 UTC on 30 August (Fig. 5a). The CPL-simulated horizontal wind speeds were relatively high on the downshear side, corresponding to the right side of the moving storm. This study defines a CB when the vertical velocity in a grid is higher than 7 m s\(^{-1}\) between 3000 and 12000 m heights (mid-to-upper troposphere) within a radius of 300 km from the storm center. Figure 5b indicates that CBs occurred on the downshear-left side at the northern edge of the convergence area. The convergence was calculated from CPL-simulated wind speeds at 20 m height. The location relative to the storm center was determined by surface friction, which depends on the moving speed of a storm (Shapiro 1983; Zhang and Uhlhorn 2012).

### 3.2 Oceanic response to the storm

As described in the Introduction, SSC was induced by Lionrock (Fig. 1). Previous studies have reported that storm-induced SSC is important to suppress the storm intensification, resulting in weakening of the maximum intensity (Bender and Ginis 2000; Emanuel 1999). However, Lionrock paused the decay of the intensity while CBs were enhanced (Fig. 4). This subsection examines the relationship between CBs and changes in intensity of Lionrock during the occurrence of the storm-induced SSC in the decay phase.

On 23 August, the sea water with SST lower than 25°C intruded southward along the eastern coast of northern Japan (Fig. 6a). On 30 August, the SST increased along the coastal areas in northern Japan, but decreased on the eastern side of the CPL-simulated storm (on the right side of the moving direction) (Fig. 6b). Changes in CPL-simulated SST from 2100 UTC 29 to 0300 UTC 30 August (Fig. 6c) was closely related to the atmospheric feedback, as reflected by the effects of the differences in CPL-simulated central pressures and MWSs (Figs. 4b, c).
This study defines the ‘cold’ area as a rectangular area, 36.0°–36.2°N, 142.95°–143.15°E, and the ‘warm’ area as a rectangular area, 36.8°–37.0°N, 142.7°–142.9°E. The reason for setting a narrow cold/warm area is to focus on physical processes in the area where the skin SST difference is large (Fig. 6c). The tendency of the skin SST was derived from the following equation.

\[
\frac{D T\text{_{skin}}}{Dt} = T\text{_{HADV}} + w_e \cdot \Delta T + \delta T\text{_{skin}},
\]

(1)

where \( T\text{_{skin}} \) is the skin SST over the ocean, \( T\text{_{HADV}} \) is the tendency of the mixed-layer temperature, \( w_e \) is the entrainment rate at the mixed-layer base (Deardorff 1983), \( \Delta T \) is the difference of water temperature between the mixed layer and top of the thermocline, and \( \delta T\text{_{skin}} \) is the variation in skin SST due to a skin effect represented by input/output of radiative and air-sea turbulent heat fluxes at the air-sea interface (Kawai and Wada 2007). This equation was based on that in Wada (2002), except the term of \( T\text{_{skin}} \) (Wada and Kawai 2009). The left-hand term is calculated in each model grid as a local tendency.

Figure 7 shows time series of the hourly areal-integrated (total) tendency of the skin SST due to the horizontal advection, entrainment with vertical advection, and skin effect over the cold (Fig. 7a) and warm (Fig. 7b) areas, respectively. The value of each term is obtained based on the output of the multilayer ocean model. The analysis roughly indicates that the effect of net horizontal advection was relatively small compared with the effect of the entrainment with vertical advection and the skin effect. The effect of the vertical advection was represented by the variation in the mixed-layer thickness included in the entrainment term (Wada 2002). The entrainment with vertical advection played a dominant role in producing the storm-induced SSC over the cold area. The cooling, mainly caused by the upwelling behind the storm, reduced the mixed-layer thickness. The reduction helped more efficiently transport the cool thermocline water into the mixed layer (Price 1981).

In contrast, the entrainment played an essential role in producing the warm water within the inner core of the storm (over the warm area). Indeed, the influence of the horizontal advection was insignificant for producing the warm area. However, the convergence of the horizontal currents ahead of the storm produced downward transport of the surface warm water into the mixed-layer base (Wada 2002). Because the translational speed was higher than the phase speed of the first baroclinic mode in the ocean (~3 m s\(^{-1}\)), the horizontal advection in the upper ocean induced
by Lionrock was enhanced on the right side of the moving storm due to the near-inertial currents caused by cyclonic circulation of the storm (Price et al. 1994; Sanford et al. 2007). The mixed layer was also deepened around the storm on the right side due to the entrainment. Therefore, these two processes favored the increase in SST over the warm area (Fig. 6c).

SST could also increase in most areas due to the solar heating during the day after 2100 UTC 29 August (Fig. 7). However, the increase in SST did not cause an enhancement in MWS from 0000 to 0600 UTC on 30 August (Fig. 4b). This was possible because the increase in SST was suppressed by short of the input of solar radiation around a TC covered with thick clouds as well as because of the effect of storm-induced SSC after the passage of the storm (the maximum value of the SSC was \( \sim 0.4^\circ C \) in Fig. 6c).

Figure 8 displays the horizontal distributions of AMSR2 L2 SST (Fig. 8a), CPL-simulated SST (Fig. 8b) at 0000 UTC on 31 August 2016 and SST (Fig. 8c) at the initial integration time, 0000 UTC on 23 August 2016. Data are plotted along a swath where AMSR2 L2 SST data exist. The AMSR2 L2 SST at 0000 UTC on 31 August, approximately 1 day after the passage of the storm, was used. The amplitude of CPL-simulated SSC was relatively small because of a relatively fast-moving speed of CPL-simulated storm (Price 1981). However, the horizontal distribution of CPL-simulated SST was qualitatively reasonable compared with that of AMSR2 L2 SST.

Figure 9a shows the horizontal distributions of CPL-simulated significant wave heights. The distribution of CPL-simulated wave heights shows asymmetry, and the wave heights were relatively high on the right side of the CPL-simulated Lionrock. The asymmetric pattern of CPL-simulated significant wave heights was enhanced on the right side of the moving storm due to the near-inertial currents caused by cyclonic circulation of the storm (Price et al. 1994; Sanford et al. 2007). The mixed layer was also deepened around the storm on the right side due to the entrainment. Therefore, these two processes favored the increase in SST over the warm area (Fig. 6c).
heights was reasonable to that of significant wave heights provided by the coastal ocean surface-wave model (Fig. 9b). However, CPL-simulated significant wave heights were relatively low around the area where the significant wave height calculated by the coastal ocean surface-wave model was maximum. This result is explained by the assumption that the oceanic surface wave in the CPL experiment is stationary at the initial time and thereby, it is difficult to realistically simulate the waves. The initial condition used in the ocean surface-wave model will be developed in a future work.

3.3 Air-sea latent heat and lower-tropospheric moisture fluxes

Air-sea latent heat flux is estimated by the speed of the surface wind at 10 m height and considering the difference of moisture between the atmosphere and ocean. In addition, it was assumed that the air near the surface over the ocean is saturated while the air-sea exchange coefficients for enthalpy are based on the Monin-Obukhov similarity theory (Monin and Obukhov 1954). Accordingly, the storm-induced SSC affects the air-sea turbulent heat fluxes, particularly air-sea latent heat fluxes within the inner core. Gener-
ally, the reduction in air-sea latent heat fluxes within the inner core decreases the intensity of the storm (Wada et al. 2014). Figure 10a shows the horizontal distribution of CPL-simulated air-sea latent heat flux at 0300 UTC on 30 August. The maximum of CPL-simulated air-sea latent heat flux was less than 300 W m$^{-2}$ within the inner core of the storm, which was comparable to areal mean latent heat flux in the inner core of TCs during the mature phase (Wada et al. 2014), and it was much lower than the value reported in previous studies (Bao et al. 2000; Cione et al. 2000). The horizontal distribution of CPL-simulated

![Fig. 10. Horizontal distributions at 0300 UTC on 30 August of the (a) CPL-simulated air-sea latent heat flux, (b) CPL-simulated specific humidity at 1040 m height, (c) CPL-simulated horizontal moisture flux at 1040 m height, and (d) CPL-simulated vertical moisture flux at 1040 m height. Contours indicate CPL-simulated sea-level pressures at intervals of 4 hPa. Bold cross marks indicate the location of the start of back trajectory analysis at 0300 UTC on 30 August.](image-url)
specific humidity at 1040 m height (corresponding to the eleventh vertical level and the atmospheric boundary layer in the lower troposphere) showed an asymmetric pattern due to the relatively high specific humidity (over 18 g kg\(^{-1}\)), which extended south of the storm center (Fig. 10b). This is quite different from the horizontal distribution of CPL-simulated air-sea latent heat flux (Fig. 10a), but similar to the horizontal distribution of CPL-simulated horizontal wind speeds (Fig. 5).

Lower-tropospheric horizontal moisture fluxes (\(Q_{FLX}\)) and vertical moisture fluxes (\(Q_{VFX}\)) in a grid were computed as follows.

\[ Q_{FLX} = Q_v \cdot V_H, \]  
\[ Q_{VFX} = Q_v \cdot w, \]

where \(Q_v\) is the product of specific humidity, \(V_H\) is the horizontal momentum flux, and \(w\) is the vertical momentum flux in each grid/level. Although the supply of moisture from air-sea latent heat fluxes was relatively small, the maximum of lower-tropospheric CPL-simulated moisture fluxes (Fig. 10c) was higher than \(800 \times 10^{-3} \text{ g m}^{-2} \text{s}^{-1}\) (Wada et al. 2013b). In addition, the horizontal moisture flux was relatively high in the east-to-north quadrants of the storm on the downshear side, despite the fact that the CPL-simulated specific humidity was relatively high on the southern side of the storm. These findings are mainly based on the dependence of the horizontal pattern of the moisture flux on the pattern of horizontal wind speeds (Fig. 5a).

However, at 0300 UTC on 30 August the area of high horizontal moisture flux corresponded to the area of downward moisture transport (Fig. 10d). On the other hand, the area of upward moisture transport near the location of MWS corresponded to the location of CBs and the frictional surface-convergence area, where the CPL-simulated significant wave height was high (Fig. 9a). This demonstrates that the upward moisture transport, locally enhanced around the convergence area, was related to the enhancement of CBs in the inner core of the storm.

### 3.4 CBs and MWS

In the previous section, it was shown that the CPL-simulated upward transport of moisture fluxes occurred locally around the frictional convergence area, before the storm when the storm-induced SSC occurred behind the storm, and they were related to CBs phenomena. It should be noted that the location of CPL-simulated MWS corresponded to the area of downward moisture flux due to precipitation (Fig. 10d). Franklin et al. (2003) reported that low-level downdrafts and enhanced vertical motion were associated with higher relative surface winds. CPL-simulated hourly rainfall rates were relatively high on the downstream side (moving direction) of the moving storm (Fig. 11), especially at the downshear-left side of the storm (Fig. 3), where there was a more mature mixture of intense convection and stratiform echo (Wada 2015). The phase of high precipitation rates represented by the wavenumber-1 pattern (Fig. 11) was shifted by a quarter cycle from the distribution of CPL-simulated horizontal wind speeds at 20 m height (Fig. 5).

Back trajectory analysis, as described in Section 2.3, indicated that the spiral transport of moist air into the area with CPL-simulated horizontal wind speeds exceeding 35 m s\(^{-1}\) at 20 m height was limited to the near-surface boundary layer where the CPL-simulated
significant wave height was high (Figs. 5, 9b, 12). The source of moist air was not the warm area where the entrainment with downward vertical advection and the input of solar radiation caused an increase in CPL-simulated SST (Fig. 6c). The area of relatively strong upward vertical velocity appeared locally, where the primary circulation of Lionrock was confluent with the spiral inflow of moist air due to surface friction (the “C” in Fig. 12). The correspondence of the confluent zone to the area where CBs occurred is consistent with the frictional convergence of fast-moving storms (Shapiro 1983; Zhang and Uhlhorn 2012). However, the confluent zone did not match the area including the location of CPL-simulated MWS, but the area of strong upward motion of air particles. CPL-simulated moisture was provided spirally by the

Fig. 12. Trajectory analysis every 6 s for 6 h starting at 0300 UTC on 30 August 2016. “C” indicates the confluent zone between the primary circulation of the storm and spiral inflow in the surface boundary layer. The color scale indicates the specific humidity. The purple dashed line indicates the convergence area at 20 m height at 0220 UTC on 30 August 2016. Solid arrows indicate the frictional inflow and tangential wind at the surface. Dashed line indicates the moving direction of simulated storm.

Fig. 13. Vertical cross sections along A–B line at 0300 UTC on 30 August 2016 of wind direction (vectors) and (a) wind speeds (colors, m s\(^{-1}\)), (b) rate of change of potential temperature (colors, K s\(^{-1}\)), and (c) pressure gradient in the latitudinal direction (colors, Pa km\(^{-1}\)). Cross marks indicate the location of MWS. The star mark in (a) represents the location of low-level jet.
region within the near-surface boundary layer.

In the CPL experiment, the low-level jet was simulated on the downstream side of the location of MWS around the convergence area at 0300 UTC on 30 August 2016 (star mark in Fig. 13a). Above the downward edge of low-level jet, the upward wind velocity became locally strong in the mid-to-upper troposphere, demonstrating the occurrence of CBs. CPL-simulated moist air was transported upward from the confluent zone (Fig. 13a), and condensational heating caused the increase in potential temperature inside the convective cell (Fig. 13b). The location of CPL-simulated MWS corresponded to the upstream side of the convective cell. Diabatic heating led to local increase in lower-tropospheric pressure gradient on the downshear-right ahead of the location of MWS (Fig. 13c). Therefore, the lower-tropospheric pressure gradient helped slightly increases in CPL-simulated MWS and decreases in CPL-simulated central pressure around 0300 UTC on 30 August.

4. Sensitivity experiments

Section 3.2 revealed that the variations in water temperature, particularly storm-induced SSC, play a crucial role in the occurrence of CBs as well as the intensity change of the storm. If the storm-induced SSC hardly occurs due to warm preexisting conditions with thick oceanic mixed layer, how could warm oceanic features affect MWS and central pressure of simulated Lionrock during the decay phase? Therefore, this section focuses on the effect of the ocean on CBs and intensity change of the storm.

As described in Section 2.2, two sensitivity experiments, FIX and SEN, were performed to understand the effects of ocean coupling and corresponding changes in the oceanic conditions. The storm track was simulated reasonably well in all three experiments (Fig. 14a), although the FIX-simulated track was slightly to the west of the best track. During the decay phase after 1800 UTC on 28 August in the FIX and SEN experiments, the increases in simulated central pressures occurred from 0000 to 1200 UTC 30 August were relatively small (Fig. 14b). However, the simulated MWS in the FIX and SEN experiments rapidly increased from 2100 UTC on 29 August to 0900 UTC on 30 August (Fig. 14c). The amplitude of the rapid increase in the simulated MWS was higher in the SEN experiment than in the FIX experiment.

An asymmetric distribution of horizontal wind speeds exceeding 20 m s\(^{-1}\) at 20 m height at 0300 UTC on 30 August was also simulated in the FIX and SEN experiments (Fig. 15). The comparison between the FIX-simulated area and CPL- and SEN-simulated areas suggests that the longer execution time of the coupled model led to reduction in the area where the horizontal wind speeds exceed 20 m s\(^{-1}\) at 20 m height. According to FIX and SEN experiments, the CBs frequently occurred at the northern edge of the convergence area (Fig. 15). The location of CBs was a little far from the location indicated by the FIX-sim-

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Fig. 14. As in Fig. 4, except for the results of FIX-, CPL-, and SEN-simulations during the decay phase.
ulated convergence area and MWS (Fig. 15a) while it matched the location suggested by the SEN-simulated experiment (Fig. 15b). Interestingly, the relative location of CBs to vertical wind shear in the FIX and SEN experiments is quite different from that indicated by the CPL experiment. In fact, the location of CBs was determined based on the direction of vertical wind shear. The number and distribution of CBs are closely related to the storm-track simulation, particularly in the FIX experiment. This indicates that the number and distribution of CBs are sensitive to oceanic conditions and are considered to affect the storm-track simulation.

As mentioned above, the vertical wind shear varied greatly between the CPL experiment and the FIX and SEN experiments. The difference of the shear direction was $134.6^\circ$ (FIX) and $143.8^\circ$ (SEN) counterclockwise with respect to the shear direction in the CPL experiment. The amplitudes of vertical wind shear, however, were similar $-5.2$, $7.2$, and $5.8$ m s$^{-1}$ in FIX, CPL, and SEN experiments, respectively. A subsequent increase in simulated MWS was noticed around 1200 UTC in all experiments (Fig. 14c). However, the amplitudes of decreases in the central pressure simulated in the FIX and SEN experiments were much smaller than expected from the amplitude of

Fig. 15. As in Fig. 5, except for the (a) results of FIX-simulation and (b) SEN-simulation.
increases in the simulated MWS based on the typical MWS–central pressure relationship (Koba et al. 1990). In the CPL experiment, the obtained wind-pressure relationship was normal.

Figures 16a, b show the time series of areal mean horizontal (Fig. 16a) and vertical (Fig. 16b) moisture fluxes in the lower troposphere in all experiments. The mean moisture fluxes were averaged for an area between 500 and 1500 m heights, where the MWS exceeds 20 m s$^{-1}$ within a radius of 300 km from the storm center. The corresponding values provided by the FIX experiment for the horizontal (0.42–0.44 g m$^{-2}$ s$^{-1}$) and vertical (0.00030–0.00045 g m$^{-2}$ s$^{-1}$) fluxes as well as the value for air-sea latent heat flux (125–145 W m$^{-2}$) were relatively high from 1800 UTC 29 to 0000 UTC 30 August during the decay...
phase. However, the mean fluxes decreased rapidly as land became included in the averaged area after 0000 UTC 30 August. The areal mean air-sea latent heat flux in the CPL experiment was higher than that in the SEN experiment from 0000 UTC to 0600 UTC 30 August. This result is consistent with the increases in CPL-simulated SST shown in Fig. 6c while the areal mean horizontal and vertical moisture fluxes in the SEN experiment were higher than those in the CPL experiment.

Figure 16c illustrates the time series of areal mean vertical wind velocity between 500 and 1500 m heights with the number of CBs between 3000 and 12000 m heights, where the MWS exceeds 20 m s$^{-1}$ within a radius of 300 km from the storm center. From 0000 UTC to 0600 UTC 30 August, the areal mean vertical wind velocity was comparable between the three experiments. However, the number of CBs increased around 0300 UTC 30 August in the FIX (374 at 0250 UTC to 744 at 0350 UTC) and SEN (272 at 0250 UTC to 832 at 0410 UTC) experiments. Because CBs was located on the downstream side in all the three experiments (Figs. 5, 15), the increase in the number of CBs led to rapid increases in MWSs near the surface (Fig. 14) based on the mechanism described in Section 3.3. After 0300 UTC 30 August, rapid increases in MWSs resulted in increases in the maximum value of frictional convergence at convergence area of around 20 m height, particularly in the FIX and SEN experiments (Fig. 17). In other words, the increase of the surface frictional convergence does not enhance CBs without increasing the vertical moisture flux (Figs. 16, 17).

Figure 18 shows the time series of areal mean horizontal and vertical moisture fluxes between 2000 and 14000 m heights, where the MWS exceeds 20 m s$^{-1}$ within the radius of 300 km from the storm center. The range of heights for calculating the areal mean moisture fluxes includes the CBs’ definition range i.e., between 3000 and 12000 m heights. These mean fluxes were calculated where the vertical wind velocity is upward and downward, respectively. The results indicate that the mean horizontal fluxes quantitatively dominated the moisture in the inner core of the storm compared with the mean vertical moisture flux. The amount of the mean CPL-simulated horizontal moisture flux at the area with positive vertical wind velocity was the smallest. In other quarters, the fixed SST conditions in the FIX and SEN experiments resulted in increases in areal mean horizontal moisture fluxes. Note that CBs had little effect on the amount of the mean vertical moisture flux. This suggests that CBs occurred locally, so that the simulated MWS rapidly increased at a relatively narrow area, but the simulated central pressure decreased only slightly.

An important result obtained from these sensitivity experiments is related to the effect of the execution time of the atmosphere-wave-ocean coupled model, which appears in the horizontal and vertical moisture fluxes rather than air-sea latent heat flux. An asymmetric storm with a relatively fast translation in mid-latitude is expected to rapidly increase MWS temporarily near the sea surface on the upstream side under a favorable oceanic condition due to the vertical moisture fluxes. Hence, the number of CBs could increase around a surface frictional convergence area. Figure 19 displays the mechanism regarding the relationship between CBs and changes in the storm intensity. The mechanism consists of the following two steps. [1] Surface convergence, increases in horizontal and vertical moisture fluxes, formation of the low-level jet and enhancement of CBs (Fig. 19a). [2] Locally occurrence of CBs, latent heating, increases in lower-tropospheric pressure gradient and temporal maintenance/increase of MWS (Fig. 19b). In addition, low-level downdrafts accompanying precipitation and enhanced vertical motion that occurred on the downstream side are related to relatively high surface winds, as shown by the asymmetric pattern of horizontal wind speeds and in the paper of Franklin et al. (2003).
5. Discussion

The comparison between the FIX- and CPL-simulated areas with high winds (see Figs. 5, 15) indicated that the execution time of the coupled model was one of the factors that determined the high-wind area. Hill and Lackmann (2009) suggested that one factor controlling a storm size is the environmental relative humidity, to which the intensity and coverage of precipitation occurring outside the storm core is strongly sensitive. A relatively dry atmospheric environment was simulated in the CPL experiment compared with the FIX-simulation due to reduction in air-sea turbulent heat fluxes even though the diurnal...
cycle of SST was simulated by the coupled model. The dry environment obtained from the simulation integrated for a several days by using the coupled model leads to reduction in hourly precipitation, particularly outside the inner core of the storm due to suppression of distant rainbands (Houze 2010). Therefore, the lateral extension of the wind field is suppressed. Nonetheless, from the comparison between the SEN and CPL experiments, the effect of the ocean coupling processes on the storm size was relatively small when the execution time of the coupled model was relatively short i.e., less than 24 h. Nevertheless, the CBs could be simulated on the downstream side of a fast-moving storm even in the CPL experiment. This suggests that the dynamic and associated thermodynamics in the atmosphere, predominantly in the lower troposphere, play an important role in the simulations of CBs.

The number and distribution of CBs are closely related to the storm-track simulation (Figs. 5, 12). This implies that the forecast errors regarding the storm track were attributed to the errors of CBs' forecasts. To accurately predict the number and distribution of CBs, this study suggests the importance of formation of surface frictional convergence area and upward moisture transport ahead of the storm. In fact, the relative position of CBs to vertical wind shear differs quite well between the three experiments. It is true that the asymmetry of hourly precipitation pattern is closely related to the vertical wind shear (Ueno and Bessho 2011). This study demonstrates that the distribution of CBs is associated with the direction of vertical wind shear, as well as the moving direction of the storm.

It is worth noticing that the change of drag coefficient scheme in the atmosphere model did not affect the track simulation (Wada et al. 2013a). The result shows that the surface friction itself may not affect the storm-track simulation. However, it is important to accurately simulate the distribution and amount of air-sea turbulent heat fluxes, horizontal and vertical moisture fluxes, as well as resulting CBs. Their influences regarding the ocean coupling processes are also important for improving the storm-track simulation. Nakano et al. (2017) reported that the replacement of the surface boundary layer scheme in a 7 km mesh nonhydrostatic global model improved the storm-track prediction. The replacement may have the same effect as a relatively long execution time of the coupled model in the storm-track simulations represented by the difference between the FIX and CPL experiments. Otherwise speaking, it is possible to change the location of CBs by replacing the surface boundary layer scheme in the atmosphere model, but the sophistication of the surface boundary layer scheme is beyond the scope of this study. However, this discussion is useful for the improvement of the storm-track predictions.

6. Conclusions

Typhoon Lionrock, the tenth tropical cyclone (TC) in the western North Pacific during the 2016 TC season, traced an unusual counterclockwise track. A special attention was paid to the landfall in the Pacific side of northern Japan, which happened for the first time since 1951, the start of the statistics of typhoons, while Lionrock kept the intensity before the landfall. Despite the fact that SSC was clearly induced on the right side of the moving storm, Lionrock underwent consecutive deep convections (convective bursts, CBs) before making landfall on 31 August in the Pacific side of northern Japan. This study was focused on the effect of CBs occurred before the storm made landfall on the intensity change of the storm during the decay phase under the oceanic conditions of the storm-induced SSC. To clarify the relation between CBs occurred in the mid-to-upper troposphere and changes in the storm intensity, numerical simulations were conducted with a 3 km mesh coupled atmosphere-wave-ocean model.

It was shown that the CBs resulted in a slight decrease in simulated central pressure and an increase in simulated MWS due to the processes illustrated in Fig. 19. Thus, the fast-moving storm-induced SSC behind the storm resulted in decreases in air-sea latent heat flux after the passage, whereas the SST increases near the storm center due to high entrainment with downward vertical advection. Moisture was horizontally advected spirally in the near-surface boundary layer due to the surface friction from the surrounding region where the significant wave height was high. The moist air horizontally transported from the outside to the inner core was confluent with the primary circulation of the storm. The upward motion locally enhanced by low-level jet was occurred around the confluent zone, and the associated vertical moisture flux in the lower troposphere resulted in CBs in the mid-to-upper troposphere ahead of the storm (Fig. 19a). The condensational heating associated with CBs resulted in lower-tropospheric pressure gradient even during the decay phase. The local increase in the pressure gradient led to a slight increase in simulated MWS due to storm-induced SSC. In addition, low-level downward moisture flux accompanying precipitation and enhanced vertical motion occurred on the downstream
side are related to relatively high surface winds and resulting asymmetric pattern of horizontal wind speeds (Fig. 19b).

The sensitivity experiments regarding the effects of the execution time of the coupled model and corresponding changes in the oceanic conditions showed an asymmetric storm with a relatively fast translation in mid-latitude, which is expected to rapidly increase MWS near the sea surface on the upstream side under a favorable oceanic condition. This is caused by the vertical moisture fluxes and the number of CBs that could increase around a surface frictional convergence area ahead of the storm. In addition, the results of the three sensitivity experiments suggest that the number and distribution of CBs are sensitive to oceanic conditions and are considered to affect the storm-track simulation and MWS. The role of ocean coupling and surface boundary layer in estimates of air-sea turbulent heat fluxes, horizontal and vertical moisture fluxes, and occurrence of CBs for improving the storm-track prediction were discussed.

Previous studies have not addressed the relationship between CBs and changes in the storm intensity in mid-latitude under the occurrence of storm-induced SSC simulated by an atmosphere-wave-ocean coupled model. Hence, this study proposes a possible mechanism (Fig. 19) to explain this relationship. Accordingly, the occurrence of CBs due to relatively high upper-ocean temperatures at higher latitudes will result in local, rapid increases in MWS that will increase the risk of natural disasters. To diminish the risk caused by local extreme atmospheric phenomena, it is claimed that a short-term weather prediction based on a high resolution atmosphere-wave-ocean coupled model including more accurate atmospheric, oceanic-wave, and oceanic initial conditions will be needed in future.

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