Intensity Change of Typhoon Nancy (1961) during Landfall in a Moist Environment over Japan: A Numerical Simulation with Spectral Nudging

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ABSTRACT

Intensity change of tropical cyclones (TCs) as they make landfall is closely linked to sustained periods of high surface winds and heavy precipitation. Few studies have investigated the intensity change of intense TCs that make landfall in middle latitudes such as Japan because few intense typhoons make landfall in middle latitudes. In this study, a numerical simulation of intense Typhoon Nancy (1961) was used to understand the intensity change that occurred when Nancy made landfall in Japan. A spectral nudging technique was introduced to reduce track errors in the simulation. During landfall, the simulated storm exhibited the salient asymmetric structure and rapid eyewall contraction. A tangential wind budget indicated that the maximum wind speed decreased concurrent with an increase in surface friction and advection associated with low-level asymmetric flows. Detailed evolution of the storm's warm core was analyzed with a potential temperature budget. In the upper part of the warm core centered at a 12-km height, cooling due to ventilation by asymmetric flows and longwave radiation overcame heating due to condensation and shortwave radiation during the contraction of eyewall clouds. In the lower part of the warm core, adiabatic cooling more than offset warm-air intrusions associated with asymmetric flows and condensational heating. The condensation was supplied by an abundance of moisture due to evaporation from the ocean in the well-developed typhoon based on a moisture budget. Sensitivity experiments revealed that environmental baroclinicity in the midlatitudes, orography, and radiative processes made minor contributions to the weakening. The weakening was instead controlled by inner-core dynamics and interactions with land surfaces.

1. Introduction

A tropical cyclone (TC) has a warm-core structure in its storm center. The warm core corresponds to intense winds in the lower troposphere via the thermal wind

relationship. The unique structure of the intense wind near the surface causes many adverse meteorological impacts, such as a storm surge, surface wind gusts, and heavy precipitation when the storm makes landfall. A full understanding of the dynamics of the changes of the structure and intensity of a storm making landfall is therefore important for quantitative predictions of storm impacts.

In previous studies, idealized simulations with numerical models have been used to investigate the intensity and structural changes of a storm as it makes landfall. Tuleya and Kurihara (1978) have elucidated the process of storm intensity change during landfall based on an energy budget analysis of a simulated, idealized storm. During landfall, the increase in surface friction over land causes a reduction of surface wind speed and enhancement of low-level inflow that results in moisture convergence (i.e., eyewall updraft) within

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the storm. The absence of surface evaporation over land, however, suppresses condensational heating in the eyewall. Eventually, the storm weakens, once the suppression of heating due to the absence of evaporation more than offsets the enhancement of heating due to moisture convergence. Tuleya and Kurihara (1978) have also indicated that a storm can maintain its intensity even during landfall if the suppression of heating due to the absence of evaporation is insufficient to offset the enhancement due to moisture convergence.

Bender et al. (1985), Bender and Kurihara (1986), and Bender et al. (1987) have identified the role of topography in weakening storms making landfall through simulations of idealized storms and realistic distributions of orography in places such as the Philippines, Taiwan, and Japan. Mountainous terrain caused the idealized storms to have clearly asymmetric structures during landfall. When a low-level inflow of a moist air mass to a storm is lifted over mountains, active condensation on the windward side of the mountains produces a dry air mass on the lee side. The orographically induced asymmetry and intrusion of dry air into the inner core of the storm as it makes landfall causes the structure of the storm to change and reduces its intensity.

Several recent modeling studies have been able to produce realistic simulations because numerical models have become more sophisticated and computer systems more powerful. Kimball (2006), who investigated a hurricane making landfall in a dry environment, has proposed that 1) dry air in the free atmosphere causes evaporative cooling outside the eyewall, 2) low-equivalent potential temperature air caused by the cooling is entrained in the low-level inflow of the storm, and 3) cooling of the air causes suppression of the eyewall updraft and weakens the storm. This result implies that weakening of the storm may depend on ground surface conditions, as suggested by the idealized simulations of Tuleya and Kurihara (1978).

Wu et al. (2009) have reported that a numerically simulated typhoon rapidly weakened while passing over the Philippines main island (Luzon Island) but re-intensified after the passage. Weakening of the warm core and eyewall of the storm were caused by strong surface friction and the mountainous terrain of the island. In particular, they pointed out that the eyewall rapidly contracted as the storm approached the island. Structural changes of the eyewall during landfall have been reported in other studies (e.g., Yang et al. 2008, 2011). Yang et al. (2011) have used a numerical simulation to explore the dynamics of the changes of the structure of a typhoon making landfall over Taiwan. They have reported that the storm eyewall rapidly

contracts during landfall and that the primary and secondary circulations of the storm are enhanced by the terrain of Taiwan. These studies suggest that weakening of the intensity and collapse of the structure of TCs are strongly controlled by surface conditions, orographic characteristics, and atmospheric conditions when the TCs make landfall.

Many similar studies have investigated structural changes and processes related to the intensity change of intense storms during landfall in low latitudes, because landfalls of intense TCs are frequent in low-latitude places such as the Philippines, Taiwan, Cuba, Florida, and northern Australia. In contrast, few studies have focused on dynamics of the intensity change of category-4 and category-5 typhoons that make landfall in middle latitudes, such as the main islands of Japan. In general, it is difficult for a storm to maintain its intensity in middle latitudes because the high ocean surface temperatures $>27^{\circ}\text{C}$ (e.g., Evans 1993) required to maintain its intensity are found only at low latitudes (e.g., Emanuel 1986).

However, a statistical analysis by Kossin et al. (2014) has indicated that the globally averaged latitude at which TCs achieve maximum intensity has migrated poleward over the past 30 years. Mei and Xie (2016) have shown that typhoons striking East and Southeast Asia have intensified by 12%–15% since 1977 on the basis of data bias corrected among different datasets. These results suggest that the frequency of intense typhoons that make landfall in middle-latitude countries may increase as the climate warms in the future. Tsuboki et al. (2015), who have used a cloud-resolving model to perform numerical simulations of intense typhoons in a warming climate scenario, have indicated that supertyphoons, which have maximum wind speeds greater than 130 kt ($1 \text{ kt} \approx 0.51 \text{ m s}^{-1}$), may occur at latitudes as high as 30°N because sea surface temperatures (SSTs) will increase as the climate warms. Thus, the intensity change of intense typhoons making landfall in the middle latitudes will be closely linked to storm-induced disasters as the climate warms in the future.

The purpose of this study was to elucidate the processes responsible for the intensity change of intense typhoons that make landfall in middle latitudes. The focus of this study was Supertyphoon Nancy (1961), which has been referred to as “Second Muroto Typhoon” in Japan. On 6 September 1961, Nancy formed to the east of the Marshall Islands as a tropical depression and moved westward until 14 September. To the east of the main island of Okinawa, the storm then turned to the northeast and made landfall around Cape Muroto, Shikoku Island, on 16 September (Fig. 1a).

FIG. 1. (a) Tracks and (b) central pressures of Typhoon Nancy (1961) estimated via observations (black), reanalysis (red), and CReSS (blue). Cross marks are plotted every 6 h, and circles are plotted at each 0000 UTC. In the inset of (a), black stars denote surface observation sites. The observations include aircraft observations made by the U.S. Air Force.

The maximum recorded wind speed and minimum central pressure of Nancy were 95 m s^{-1} and 888 hPa, respectively, according to the best track data provided by the Japan Meteorological Agency (JMA) Regional Specialized Meteorological Center (RSMC) Tokyo. Nancy holds the record as the typhoon with the longest-lasting Saffir–Simpson category 5 winds at 5.5 days (Cerveny et al. 2007). According to the JMA, Nancy is the most intense typhoon (with an estimated central pressure of 925 hPa) at the landfall in Japan after 1951. At landfall, the minimum surface pressure recorded at Cape Muroto was 930.9 hPa. After Nancy made landfall around 0000 UTC 16 September, the central pressure remained lower than 940 hPa as Nancy traveled over Japan (Fig. 1b). The passage of the storm over Japan left 194 people dead and 4972 people injured (JMA 1967). Most of the casualties and damage were caused by Nancy's intense winds (e.g., Yamamoto et al. 1963; JMA 1967).

Numerous observations by aircraft, surface observatories, radar, and a satellite¹ at that time provided insufficient resolution to clarify the processes by which Nancy weakened. In the present study, we carried out a numerical simulation using a nonhydrostatic regional model with full physics to investigate the change of Nancy's intensity and structure during its landfall in Japan. Results of the simulation were validated with available observations. The storm intensity was quantified in terms of its central pressure and wind speed. Evolution of the wind speed in the simulated storm during the landfall is examined by a tangential wind budget (e.g., Tuleya and Kurihara 1978). In making the landfall in midlatitudes, the storm structure and intensity can be influenced by not only local orography (e.g., Bender and Kurihara 1986; Wu et al. 2009; Yang et al. 2011), but also baroclinicity in midlatitudes [i.e., an extratropical transition (ET); Jones et al. 2003]. Kitabatake (2008, 2011) has indicated that typhoons approaching and making landfall in Japan often experience an ET. Moreover, the baroclinicity can influence the intensity change of TC through the ventilation of warm core in TC (Tang and Emanuel 2010). The vertical structure and amplitude of TC's warm core were quantified in terms of its central pressure and wind speed via hydrostatic and gradient wind balances (e.g., Chen and Zhang 2013). Thus, evolution of the warm core might be useful for understanding of the storm intensity change during the landfall, in addition to evolution of the wind speed in the storm. The potential temperature budget near the simulated storm center was analyzed as in previous studies (e.g., Stern and Zhang 2013; Ohno and Satoh 2015) to investigate the evolution of the warm core during landfall. On the basis of the budget analyses, several sensitivity experiments and moisture budget were performed to examine the mechanisms responsible for the proposed intensity change during the landfall.

2. Methodology

a. Model description

The numerical study of Typhoon Nancy was conducted with the Cloud Resolving Storm Simulator (CReSS 3.4.2), which is a three-dimensional, regional, compressible nonhydrostatic model (Tsuboki and Sakakibara 2002). The CReSS model uses a terrain-following coordinate system in the vertical and calculates the three-dimensional wind velocity components,

¹ In 1960, the world's first meteorological satellite, *TIROS-I*, was launched in the United States.

TABLE 1. List of the model physics.

Physics	Description
Microphysics	Explicit bulk cold rain at a single moment for condensation (Lin et al. 1983)
Surface exchange	Aerodynamic bulk formula from Kondo (1975) for ocean and from Louis et al. (1981) for land
Subgrid scale	1.5-order closure with TKE prediction (Deardorff 1980)
Radiation	Rapid Radiative Transfer Model (RRTM-G; 4.85) described in Mlawer et al. (1997) and Iacono et al. (2003)

pressure perturbation, potential temperature perturbation, turbulent kinetic energy (TKE), and the mixing ratios of water vapor, cloud water, rain, cloud ice, snow, and graupel. The CReSS model has been used to study many aspects of TCs (e.g., Akter and Tsuboki 2012; Wang et al. 2012; Tsujino et al. 2017). Table 1 summarizes the model physics. The CReSS model does not use cumulus parameterization.

b. Spectral nudging technique

This study focused on the landfall of an intense typhoon with Saffir–Simpson category 4 or 5. The simulation was therefore started during an early developmental stage of the typhoon to reproduce the fully developed structure of the storm, and it required an integration time longer than one week (Fig. 1). The long integration time can induce slightly missing evolution of synoptic-scale disturbances in a regional model. The wrong evolution can be induced by strong internal variability embedded in regional features such as topography, surface conditions, and thermal convection in the model (e.g., von Storch et al. 2000; Riette and Caya 2002; Alexandru et al. 2009), and it can amplify errors of typhoon track. Some studies have indicated that spectral nudging is an effective way to improve synoptic-scale disturbances and storm tracks in regional simulations (e.g., Cha et al. 2011; Choi and Lee 2016). A spectral nudging method was therefore used in the present study to reproduce the storm track more accurately.

An arbitrary variable $\psi = \psi(\lambda, \phi, z^*, t)$ in latitude and longitude coordinates (λ, ϕ) can be expressed as a Fourier series as follows:

$$\psi(\lambda, \phi, z^*, t) = \sum_{j=-J_m, k=-K_m}^{J_m, K_m} \hat{\psi}_{j,k}(z^*, t) \times \exp\left(\frac{ij\lambda}{L_\lambda}\right) \exp\left(\frac{ik\phi}{L_\phi}\right),$$

where j and k are zonal (λ) and meridional (ϕ) wavenumbers, respectively; L_λ and L_ϕ are the zonal and meridional sizes (degrees) of the model domain, respectively; z^* is the vertical coordinate in the terrain-following direction; and t is time. The expression $\hat{\psi}$ represents the Fourier coefficient of ψ . The variables J_m

and K_m are the zonal and meridional wavenumbers resolvable in a simulation. Spectral nudging is performed by including an additional forcing term in the prognostic equations of the CReSS model as follows:

$$\frac{\partial \psi}{\partial t} = L(\psi) + \sum_{j=-J_m, k=-K_m}^{J_m, K_m} \eta_{j,k}(z^*, t)(\hat{\psi}_{j,k}^a - \hat{\psi}_{j,k}) \times \exp\left(\frac{ij\lambda}{L_\lambda}\right) \exp\left(\frac{ik\phi}{L_\phi}\right), \quad (1)$$

where L represents the model operator; the superscript a represents a reference such as reanalysis data, used as an initial or boundary condition in the regional simulation; and $\eta_{j,k}$ is a nudging coefficient that depends on horizontal wavenumbers, height, and time. The nudging coefficient is expressed as follows:

$$\eta_{j,k}(z^*, t) = f(z^*, t) \times \begin{cases} \alpha, & |j| \leq J_T, \quad |k| \leq K_T, \\ 0, & |j| > J_T, \quad |k| > K_T, \end{cases} \quad (2)$$

where J_T and K_T are truncating wavenumbers, and α is a constant. In previous studies, J_T and K_T have been upper bounds of wavenumbers forced by the additional term of Eq. (1). In general, large-scale (i.e., lower wavenumbers) disturbances mainly control the typhoon track. The function $f(z^*, t)$ is defined by

$$f(z^*, t) \equiv \beta(t) \times \begin{cases} (z^* - z_b)/(z_t - z_b), & z^* \geq z_b, \\ 0, & z^* < z_b, \end{cases}$$

where z_t and z_b are the model top and bottom height of the forcing, respectively. In the present study, $\beta(t) = 1$.

c. Experimental design

The model domain was 64.8° in the zonal direction $\times$ 51.2° in the meridional direction $\times$ 24.75 km in height (Fig. 1). The horizontal grid spacing was uniformly 0.05° in both the zonal and meridional directions. The vertical grid was a stretching vertical coordinate. The lowest grid spacing was 200 m, and there were 55 vertical grids (Table 2). The integration period was from 0600 UTC 8 September to 1200 UTC 16 September. In the present simulation, the Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) dataset, which was provided by

TABLE 2. Vertical grid points (where any prognostic scalars are defined) over ocean in the present study.

Grid numbers	Heights (m)									
1–10	100.0	313.8	553.9	817.8	1102.9	1407.1	1728.1	2063.9	2412.6	2772.4
11–20	3141.8	3519.1	3902.9	4292.2	4685.6	5082.3	5481.4	5882.1	6283.9	6686.4
21–30	7089.1	7491.9	7894.7	8297.6	8700.8	9104.6	9509.5	9916.0	10 325.0	10 737.2
31–40	11 153.6	11 575.3	12 003.6	12 439.9	12 885.7	13 342.6	13 812.5	14 297.1	14 798.5	15 319.0
41–50	15 860.8	16 426.3	17 018.1	17 630.5	18 249.6	18 868.7	19 487.7	20 106.8	20 725.9	21 345.0
51–55	21 964.1	22 583.2	23 202.3	23 821.4	24 440.5					

the JMA and covers the period from 1958 to the present, was used for the initial and boundary conditions. Table 3 provides details. The simulation used no additional storm initializations such as bogus vortices (e.g., Kurihara et al. 1995).

Previous studies have reported that some parameters in the spectral nudging influence typhoon track, intensity, and precipitation (e.g., Alexandru et al. 2009; Cha et al. 2011; Choi and Lee 2016). Sensitivity of the simulated storm to the parameter values was examined to determine the best values of the parameters in the simulation (see supplemental material). Then the nudging coefficient ($\eta_{j,k}$) parameters were set to $f = 1$ and $\alpha = 2.77778 \times 10^{-4} \text{ s}^{-1}$ (e -folding time of 1 h). The truncating wavenumbers were set to $J_T = 6$ and $K_T = 4$. These values correspond to wavelengths longer than 1500 km. Nudging was imposed on horizontal wind velocities.

On the bottom boundary, the ground and ocean temperature, including the bulk ocean surface temperature, was predicted on the basis of a one-dimensional thermal diffusive equation in the vertical direction (Segami et al. 1989). The ocean model considered thermal diffusive processes near the ocean surface because there were

few ocean observations during Typhoon Nancy. In addition, the ocean model had heat inputs (i.e., radiation heat, and latent and sensible heat) on the surface. The simple ocean model had a vertical resolution of 20 cm and 30 layers that included a part of the ocean mixed layer. In advance, we conducted sensitivity tests to the depth of the ocean model by changing the vertical grid spacing in the ocean model because the effective heat capacity of the ocean can largely influence the storm intensity. The best one in the reproducibility of the storm intensity change during the landfall was selected, in spite of the shallow ocean depth in the model. We assumed that the initial ocean temperature in the model was vertically uniform (i.e., equal to the SST). SST data are included in the JRA-55 [i.e., Centennial In Situ Observation-Based Estimates of SST (COBE-SST) data; Ishii et al. 2005].

d. Budget analyses

Evolution of the wind speed in the simulated Nancy was examined by a tangential wind ($\bar{v}$) budget equation in cylindrical coordinates:

$$\frac{\partial \bar{v}}{\partial t} = \text{TADV} + \text{TEDD} + \text{DISV}. \quad (3)$$

TABLE 3. Summary of the experimental design.

Characteristic	Description
Model domain	64.8° (zonal) $\times$ 51.2° (meridional) $\times$ 24.75 km (vertical)
Grid spacing	0.05° (horizontal) and Table 2 (vertical)
Initial and boundary	JRA-55 (1.25° $\times$ 1.25°) with 6-hourly output (Kobayashi et al. 2015)
Upper boundary	Rigid lid with a sponge layer from a height of 18 km to the model top to prevent reflection of vertically propagating waves
Lateral boundary	Radiation of waves with a constant phase velocity (e.g., Klemp and Wilhelmson 1978; Tsujino et al. 2017)
Integration period	0600 UTC 8 Sep–1200 UTC 16 Sep 1961
Model output	Hourly, and 1-min interval for some analyses (0600 UTC 14 Sep–1200 UTC 16 Sep 1961)
Terrain	Shuttle Radar Topography Mission data with a horizontal resolution of 30 s from the U.S. National Aeronautics and Space Administration
Initial SST	COBE-SST (same resolution as the JRA-55 dataset; Ishii et al. 2005)
Ocean model	One-dimensional thermal diffusive process (Segami et al. 1989)
Cloud radiation	Radiative effects of ice (Ebert and Curry 1992) Categories to calculate the radiation: liquid and ice water clouds, snow, and graupel Fixed absorption coefficient of $0.0602410 \text{ m}^2 \text{ g}^{-1}$ for liquid water Effective radius for 50- μm ice crystals No mixing ratios of carbon dioxide and ozone

Then

$$\begin{aligned} \text{TADV} &= \text{HADV} + \text{VADV}, \\ \text{HADV} &\equiv -\bar{u}(\bar{\zeta} + \bar{f}), \\ \text{VADV} &\equiv -\bar{w} \frac{\partial \bar{v}}{\partial z}, \\ \text{TEDDV} &\equiv -\overline{u'\zeta'} - \overline{u'f'} - \overline{w' \frac{\partial v'}{\partial z}}, \end{aligned} \quad (4)$$

where TADV, TEDDV, and DISV denote axisymmetric and asymmetric advection terms of $\bar{v}$ and momentum source, respectively. TADV is decomposed into horizontal (HADV) and vertical (VADV) advection, respectively. DISV is directly calculated by the model output of friction, turbulent mixing, numerical diffusion, and spectral nudging forcing. The overbar and prime denote the azimuthal average and deviation from the azimuthal average, respectively. The storm center is estimated by the grid with the most axisymmetric distribution of sea level pressure based on Braun (2002) in Tsujino et al. (2017). Radial (r) and vertical (z) wind components are denoted as u and w , respectively. Time is defined as t . Symbols of f and ζ denote the Coriolis parameter and vertical component of relative vorticity vector, respectively. All model variables are output in the terrain following coordinate. Before calculation of any terms in Eq. (3), all variables are linearly interpolated in horizontally constant levels of Table 2.

Analysis of potential temperature budgets can reveal the dynamics of the evolution of a storm eye (e.g., Stern and Zhang 2013; Ohno and Satoh 2015). In the present study, the potential temperature (θ) budget equation in cylindrical coordinates was expressed as follows:

$$\frac{\partial \bar{\theta}}{\partial t} = \text{TADV} + \text{DIABQ}, \quad (5)$$

where TADV and DIABQ denote advection of $\bar{\theta}$ and diabatic heating due to microphysics, radiation, and turbulent processes, respectively. For more detailed analysis, TADV was decomposed as follows:

$$\text{TADV} = \text{HSADV} + \text{HAADV} + \text{VADV}. \quad (6)$$

Then

$$\begin{aligned} \text{HSADV} &\equiv -\bar{u} \frac{\partial \bar{\theta}}{\partial r}, \\ \text{HAADV} &\equiv -\frac{\partial r u' \theta'}{r \partial r}, \\ \text{VADV} &\equiv -\bar{w} \frac{\partial \bar{\theta}}{\partial z} - \frac{\partial \rho_0 w' \theta'}{\rho_0 \partial z}, \end{aligned} \quad (7)$$

where HSADV, HAAHV, and VADV are horizontal symmetric advection, horizontal asymmetric advection, and vertical advection, respectively. The parameter ρ_0 is the horizontally averaged air density.

We expect that the DIABQ will be dominated by condensational heating in the eyewall. A moisture budget over a system scale was calculated for more quantitative examination of sources in the DIABQ. The budget equation was expressed as follows:

$$\begin{aligned} \frac{dQ}{dt} &= - \int_{z=0}^{z=H} \int_{\lambda=0}^{\lambda=2\pi} u \rho_0 q_v R d\lambda dz \\ &\quad + \int_{r=0}^{r=R} \int_{\lambda=0}^{\lambda=2\pi} r(E - P) d\lambda dr, \end{aligned} \quad (8)$$

where symbols of q_v and λ denote mixing ratio of water vapor and azimuthal angle, respectively. Evaporation from the ocean (E) and precipitation (P) are calculated by the model output directly. Then the moisture amount Q is defined as

$$Q \equiv \int_{z=0}^{z=H} \int_{r=0}^{r=R} \int_{\lambda=0}^{\lambda=2\pi} \rho_0 q_v r d\lambda dr dz. \quad (9)$$

The budget calculation was over an annular region with $R = 800$ km from the storm center for the surface to $H = 17$ km during the approaching period (i.e., 0000 UTC 15 September to 0000 UTC 16 September 1961).

3. Results

a. Validation of the simulation

The numerical simulation was validated with available observations and analyses (Table 4). The simulated track closely followed the observed track (Fig. 1a). The root-mean-square error (RMSE) of the track was about 53 km during the simulation period of 8.25 days. The error was smaller than the horizontal grid (1.25°) of the JRA-55 data. The small difference reflects the efficacy of using the spectral nudging method. In actual, the error in no nudging simulation was 469 km (see Table S1 in the online supplemental material).

The temporal evolution of the simulated central pressure agreed reasonably well with the best track analysis (Fig. 1b). The RMSE of the central pressure was about 8 hPa during the simulation period of 8.25 days. The simulated central pressure (895 hPa) at 0000 UTC 12 September was very close to the central pressure of 888 hPa at 0045 UTC 12 September that was observed by an aircraft. The simulated storm made landfall in Japan shortly after 0000 UTC 16 September, and the rapid increase in the storm's central pressure was

TABLE 4. List of validated elements and used datasets.

Element	Dataset	Authorization
Track and central pressure	Best track data	RSMC-Tokyo (JMA)
Pressure pattern and synoptic disturbances	Weather maps	National Institute of Informatics
Horizontal distributions of meteorological elements	Technical reports	Yamamoto et al. (1963) and JMA (1967)
Local surface pressure and wind speed	Surface observatories	JMA

reproduced reasonably well during the period of landfall from 0000 to 0600 UTC 16 September. We therefore believe that the simulation can be used to analyze the process associated with the weakening during landfall. We defined the period from 0000 UTC 15 to 0000 UTC 16 September as the approaching period and compared that period with the period of the landfall.

The simulation results during the approach and landfall were compared with weather maps (Fig. 2) on some isobaric surfaces, and with surface maps (Fig. 3) to check synoptic fields around the simulated storm. During its approach, Nancy almost maintained an upright structure from the lower troposphere to the middle troposphere, except for slightly northward tilting with height (Figs. 2a–c). This structure was also reproduced in the simulation (Figs. 2d–f). The fact that synoptic-scale disturbances such as the subtropical high over the North Pacific and the trough around northeastern China were reproduced rather well in the simulation, despite the long-term integration, indicates that the spectral nudging technique was effective in maintaining synoptic-scale disturbances in the model.

Moreover, the simulation qualitatively reproduced a horizontal distribution of sea level pressure (Figs. 3d–f) comparable to the analyzed distribution (Figs. 3a–c). A clear rainband stretching to the northeast on the north side of the simulated storm corresponded to cold fronts analyzed over the Sea of Japan (Fig. 3). The location of the cold fronts was approximately stationary during the approach and landfall of Nancy. This result suggests that the simulation could reproduce synoptic disturbances both at the surface and in the upper atmosphere.

The detailed distributions of the simulated dynamic and thermodynamic fields were compared with the observed fields to validate the storm-scale structure of the simulation. [Yamamoto et al. \(1963\)](#) have used surface observations to determine the horizontal distributions of minimum sea level pressure and severe wind associated with Nancy (Figs. 4a,c). For the horizontal distribution of the minimum sea level pressure, hourly model output was compared with the observations (Fig. 4b). For the simulated maximum wind speed, we compared the maximum of the observed 10-min-average wind speed with that of the 10-min average of the simulated 1-min wind speed output (Fig. 4d).

The distribution of the simulated minimum sea level pressure agreed rather well with the observed minimum sea level pressure (Figs. 4a,b). The distribution of the simulated maximum wind speed was qualitatively consistent with observations (Figs. 4c,d). In particular, severe wind was reproduced around coastal regions on the east side of Shikoku and the southwest sides of the Kii Peninsula, Osaka Bay, and Wakasa Bay. Severe wind was also reproduced along the storm track. However, the simulated maximum wind speed was weaker than the observed maximum wind speed where the latter was exceptionally high.

Finally, we directly compared the temporal variations of the simulated and observed characteristics of Nancy at some surface observatories. At the Muroto-Misaki observatory, which was the closest observatory to the storm center when Nancy made landfall (see subfigure of Fig. 1a), the minimum sea level pressure of 930.4 hPa was observed at 0039 UTC 16 September. The simulated minimum sea level pressure of 923.8 hPa was recorded at 0100 UTC 16 September. At the Osaka regional headquarters (see subfigure of Fig. 1a), the minimum sea level pressure of 937.0 hPa was observed at 0429 UTC 16 September. The simulated minimum sea level pressure of 933.0 hPa occurred at 0500 UTC 16 September. The temporal variations of the simulated and observed sea level pressures (Figs. 5a,c) as well as the temporal variations of the simulated and observed wind speeds (Figs. 5b,d) were very similar. However, the simulated maxima at both locations were weaker than the observed maxima as the storm approached.

It is likely that the difference between the simulated and observed maximum wind speeds apparent in Figs. 4c,d and 5b,d can be explained by a diagnosis of surface winds and the relatively coarse resolution of the model simulation. However, the simulation could reasonably reproduce the horizontal distribution and temporal variation of sea level pressure during landfall. The reasonable simulation could therefore be used to analyze the structure of Nancy.

b. Storm-scale structure

In this subsection, we show in detail the storm-scale axisymmetric and asymmetric structural changes during Nancy's approach and landfall. From an axisymmetric

FIG. 2. (left) Weather charts analyzed by the JMA at 1200 UTC 15 Sep 1961 for the (a) 850-, (b) 700-, and (c) 500-hPa levels and (right) horizontal distributions of geopotential heights in the simulation at (d) 850, (e) 700, and (f) 500 hPa.

FIG. 3. (a)–(c) Surface weather charts analyzed by the JMA and (d)–(f) horizontal distributions of simulated sea level pressure and precipitation at (top) 0000 UTC 15 Sep, (middle) 1200 UTC 15 Sep, and (bottom) 0000 UTC 16 Sep 1961. Contours and colors are sea level pressure (hPa) and precipitation rate (mm h^{-1}) in (d)–(f).

FIG. 4. Comparison of (a),(c) observed (OBS) and (b),(d) simulated (SIM) (top) minimum sea level pressure (hPa) and (bottom) maximum wind speed. Observations of (a) the minimum sea level pressure and (c) 10-min averaged maximum wind speed during 14–16 Sep 1961 are based on surface observations [adapted from Figs. 2 and 7 of Yamamoto et al. (1963)]. Dash-dotted lines denote estimated tracks of Nancy in the observation and simulation. Dots in (a) and (c) denote surface observations sites. In (a), dashed lines with time indicate that the minimum pressure on the lines was estimated at that time. See the text for details of estimations of (b) the minimum sea level pressure and (d) maximum wind speed in the simulation.

perspective, there were updrafts of 0.2 m s^{-1} , strong tangential winds exceeding 50 m s^{-1} , and strong low-level inflows of 12 m s^{-1} at a radius of 50 km during the approach (Figs. 6a,e). In addition, there was a layer with weak outflows above the inflow boundary layer (Fig. 6e). This might be a common feature of mature typhoons using the CReSS model as also shown in another study

by Tsujino et al. (2017). There was a large peak of temperature anomaly, which is defined as departure from azimuthally averaged temperature at a radius of 400 km, at an altitude of $\sim 13 \text{ km}$ at the storm center. The maximum anomaly exceeded 10 K, and the large peak expanded within a radius of 50 km. The peak height in Nancy was higher than that in typical category 5 TCs

FIG. 5. Temporal variation of (a),(c) sea level pressure and (b),(d) 10-min surface wind speed during 14–16 Sep 1961 at (top) Muroto-Misaki and (bottom) Osaka. Crosses and solid lines denote observations (OBS) (every 3 h) and simulations (SIM) (every 1 h), respectively (the observed minimum sea level pressure and maximum wind speed are from http://www.data.jma.go.jp/obd/stats/data/bosai/report/1961/19610915/19610915_b2.html).

based on observations (e.g., Wang and Jiang 2019) and another model (Ohno et al. 2016).

The eyewall updraft of the storm rapidly contracted, and the amplitude of the temperature anomaly peak (i.e., warm core) rapidly decreased as Nancy made landfall (Figs. 6b–d). The size of the warm core also decreased. Eventually, the eyewall updraft became located underneath the warm core (Figs. 6c,d). The front edge of the low-level inflow also contracted, and it reached the storm center as the eyewall updraft rapidly contracted (Figs. 6f–h). At 0420 UTC 16 September, the inflow reached a local peak that exceeded 12 m s^{-1} at radii of about 50–150 km (Fig. 6g). Beginning at the same time, the maximum of the tangential wind gradually weakened from 45 to 40 m s^{-1} for about 1 h. Eyewall contraction associated with landfall has been reported in previous studies (e.g., Wu et al. 2009; Yang et al. 2008, 2011).

During Nancy's approach, its updraft had an annular structure within a radius of 100 km, and the updraft reached the upper troposphere at a height of $\sim 12 \text{ km}$ (Figs. 7a,b). During landfall, the annular structure was destroyed, and a salient, nonaxisymmetric structure appeared (Figs. 7c–f). In particular, a stronger updraft became localized in the rear side of the storm's direction of motion. That updraft had an outward tilting structure at 0420 UTC 16 September. At 0540 UTC 16 September, the tilting structure became more upright as the storm penetrated farther inland. The strong, asymmetric

updraft was enhanced both on the rear side of the storm and to the left of the direction of motion of the storm at 0540 UTC 16 September (Fig. 7e). The strong updraft to the southwest of Nancy penetrated within a 40-km radius of the storm track as the typhoon's axisymmetric structure rapidly contracted (Figs. 6d and 7f). Then storm-scale vertical wind shear (VWS), defined as difference of area-averaged winds between 200- and 850-hPa levels within 400 km from the storm center (Li et al. 2015), had northward direction (Figs. 7c,e). The strong updraft was approximately located in the left-hand side of the VWS vector. The relationship between the inner-core updraft and storm-scale VWS is partly consistent with idealized experiments in Li et al. (2015). The reduction of the intensity of the simulated storm could therefore have been influenced by both axisymmetric and nonaxisymmetric structures.

c. Tangential wind budget analysis

First, changing processes in azimuthal average of tangential wind speed $\bar{v}$ during the landfall (from 0000 to 0600 UTC 16 September 1961) were briefly shown in Fig. 8 based on a tangential wind budget. Consistent with Figs. 6f–h, the storm experienced clear decreases in the wind speed beyond a radius of 30 km (Fig. 8a). In particular, there were drastically decreases near the surface and in layers from 2 to 4 km near the radius of maximum wind speed (RMW). In addition, the storm exhibited a clear increase in $\bar{v}$ within the radius of 30 km.

FIG. 6. Radius–height cross section of (a)–(d) azimuthally averaged vertical velocity (contour lines; $m s^{-1}$) and departure (color shading; K) from azimuthally averaged temperature at a radius of 400 km in the simulation and (e)–(h) azimuthally averaged radial velocity (color shading; $m s^{-1}$) and tangential velocity (contour lines; $m s^{-1}$). Black dashed lines represent the 40-km radius of the warm-core.

FIG. 7. (left) Horizontal and (right) vertical distributions of vertical velocity around the storm. In the left column, color shading and contour lines denote vertical velocity at a height of about 5 km and terrain height, respectively. The contour interval is 200 m. Black stars denote the storm center. Direction and strength of the storm-scale VWS are shown by black vectors and values (m s^{-1}), respectively. Green dashed lines are identical to crossing lines in the right column. In the right column, vertical cross sections of vertical velocity are through the storm center (black line) along the green dashed lines from A to A'. Times and dates are (a), (b) 0940 UTC 15 Sep, (c), (d) 0420 UTC 16 Sep, and (e), (f) 0540 UTC 16 Sep 1961. The warm-core size of the 40-km radius is represented by the black circle in the left column and the black dashed lines in the right column.

The increase indicated inward moving of the RMW associated with the eyewall contraction during the landfall (Figs. 6f–h). The budget analysis was almost similar to the actual change (Figs. 8a,b), except for slight

difference near the surface. The budget result is useful for understanding of the evolution of $\bar{v}$ although the difference can be mainly induced by error of vertical interpolation during the landfall.

FIG. 8. Radius–height cross sections of (a) actual change of axisymmetric tangential wind $\bar{v}$, (b) sum of the right-hand side in the budget equation of $\bar{v}$, (c) axisymmetric advection, (d) asymmetric advection, (e) external forcing, (f) horizontal advection in (c), and (g) vertical advection in (c) during landfall (color shading; m s⁻¹ for 6 h). Contour lines denote temporally averaged $\bar{v}$ during the landfall (m s⁻¹). The external forcing is calculated by the direct output of surface friction, turbulent mixing, numerical diffusion, and forcing due to the spectral nudging in the CReSS model.

Keper (2001) proposed that strong low-level inflow can lead to jet of tangential wind through inward advection of the absolute angular momentum within the TC boundary layer. In actual, Kitabatake and Tanaka (2009) observed the low-level jet in a realistic typhoon case. The simulated Nancy also exhibited large positive contribution to increase in $\bar{v}$ due to inward advection of angular momentum associated with strong axisymmetric inflow below a height of 2 km (Figs. 8c,f).

On the other hand, large surface friction over the land and asymmetric advection contributed to the decrease in $\bar{v}$ (Figs. 8d,e). As a result, the decrease totally overcame the inward momentum advection, in contrast to the previous studies.

Concentration of the decrease in $\bar{v}$ near the RMW at the low levels drives the decrease in vertical shear of $\bar{v}$. The vertical shear decrease can modulate radial gradient of temperature in the warm core through the thermal

wind relationship. Dynamics of the warm-core evolution will be examined in the next subsection.

d. Potential temperature budget analysis

To explain the evolution of the warm core during landfall, the potential temperature budget based on Eq. (5) was analyzed within a 40-km radius of the storm center, which was approximately the inner edge of the strong eyewall updraft before landfall.² The actual change of $\bar{\theta}$ was rather similar to the sum of the right-hand side of Eq. (5), except for small differences at altitudes of 1–6 km (Fig. 9a). The actual changes were almost zero within this range of altitudes, and this layer therefore had little influence on the intensity change during landfall. The change of $\bar{\theta}$ was negative in two layers associated with altitudes of 6–12 and 12–17 km and centered at 12 km, the peak altitude of the warm core in Fig. 6.

A positive contribution from diabatic heating (DIABQ) associated with the eyewall was almost balanced by the negative contribution from total advection (TADV) in the lower layer (Fig. 9a). The positive DIABQ was composed mainly of condensation heating in the eyewall. Note that longwave radiative heating partly contributed to the positive DIABQ at altitudes of 10–12 km (Fig. 9c). The radiative heating was due to absorption of upward longwave radiation in the eyewall clouds. The negative TADV was controlled mainly by vertical advection (i.e., adiabatic cooling; VADV) associated with the eyewall updraft (Fig. 9b) to an altitude of 13 km. During landfall, dry-air intrusion into the eyewall weakened the storm (e.g., Tuleya and Kurihara 1978; Bender et al. 1985; Bender and Kurihara 1986; Kimball 2006). This weakening corresponded to a rapid collapse of the warm core by adiabatic cooling in the absence of a positive DIABQ because of the intrusion of dry air. However, the adiabatic cooling was almost cancelled by the positive DIABQ in the case of the simulated Nancy. The weakening therefore differed from the weakening observed in previous studies (Tuleya and Kurihara 1978; Bender et al. 1985; Bender and Kurihara 1986; Kimball 2006). Moreover, a positive contribution by horizontal advection due to asymmetric flow (HAADV) maintained the lower layer of the warm core (Fig. 9b). The positive contribution was significantly different from the pattern in the eye of a typical TC (e.g., Stern and Zhang 2013; Ohno and Satoh 2015). The contributions were due to

the rapid contraction of the simulated eyewall during landfall (Figs. 6 and 7).

In the upper layer, DIABQ acted as a negative contributor to the $\bar{\theta}$ budget at altitudes of 13–15 km and positive contributor at altitudes above 15 km (Fig. 9a). The negative DIABQ at altitudes of 13–15 km was due mainly to microphysical processes. The positive contribution at altitudes above 15 km was due to a combination of microphysical and radiative processes. In particular, the positive DIABQ was contributed by condensation and shortwave radiative heating (Fig. 9c). The radiative heating was caused by absorption of downward solar radiation in the eyewall clouds because landfall occurred during daytime (0000–0600 UTC 16 September). However, the positive heating was almost cancelled by the large cooling due to longwave radiation (Fig. 9c). The cooling was caused by upward emission of longwave radiation in the eyewall clouds. Above an altitude of 15 km, the large negative TADV overcame the positive DIABQ (Fig. 9a). The large negative TADV was composed mainly of negative HAADV, unlike the negative TADV in the lower layer (Fig. 9b). The negative HAADV was caused by outward advection of a warm air mass associated with asymmetric radial flows u' mainly in the warm core.

In the present study, the TADV was directly estimated by the model output. We examined accuracy of the decomposition of the TADV term into the HSADV, HAADV, and VADV terms (Stern and Zhang 2013; Ohno and Satoh 2015). Sum of the HSADV, HAADV, and VADV estimated by the azimuthally averages and anomalies of the wind components and potential temperature (the black dashed line in Fig. 9b) was quite following the TADV term (the black solid line in Fig. 9b) in all levels. Thus, the decomposition is sufficiently accurate for the above discussion on the HSADV and HAADV terms.

e. Moist environment

On the basis of energy budget analyses, Tuleya and Kurihara (1978) have proposed that weakening of diabatic heating in the eyewall of a TC due to suppressed evaporation can cause the storm's intensity to decrease during landfall. In their study, the small amount of moisture near the surface in the landfall area inhibited diabatic heating with the updraft. However, the mechanism responsible for the intensity change of the simulated Nancy during landfall differed from the process that they described. In the case of Nancy, adiabatic cooling was almost balanced by diabatic heating associated with the sustained updraft of the eyewall during landfall. We then asked why the eyewall updraft was maintained during landfall.

² We examined the potential temperature budget analysis for different size (radii of 20, 30, and 40 km). The DIABQ value decreases as the size decreases in the lower troposphere. However, we focus on warm air mass in the eye before the landfall. Thus, we consider that the difference is minor for our discussion about the warm core in the upper troposphere.

FIG. 9. Vertical profiles of (a) sum (TADV + DIABQ, black) in Eq. (5), actual change of $\bar{\theta}$ (yellow), DIABQ (red), and TADV (blue) based on $\bar{\theta}$ budget analysis, (b) details (HSADV + HAADV, red; VADV, yellow; HSADV, pink; HAADV, blue) of the TADV

On 14 September, there was a large amount (60 mm) of precipitable water vapor (PWV) within a radius of about 1000 km from the storm center (Fig. 10a). Closely associated with the region of high moisture content was a strong surface wind that exceeded 10 m s^{-1} . The high-moisture region was the result of active evaporation from the ocean due to the strong surface wind. The area of high moisture was linked to another high-moisture region to the south that extended from the South China Sea to the Philippine Sea. The air mass in the southern high-moisture region could be supplied to the storm by synoptic-scale flow of the subtropical high. The high moisture region around the storm was maintained despite the storm's making landfall (Figs. 10c,d). The prolonged period of high humidity around the storm is consistent with the large, positive diabatic heating (DIABQ) apparent in Fig. 9c. This unique situation—the abundant supply of moisture during landfall—differed from the conditions that prevail when a TC weakens in a dry environment over land as in the scenario proposed by Kimball (2006).

Bender and Kurihara (1986) have indicated that a moderate typhoon as it approaches Japan is weakened due to orographically induced, asymmetric intrusion of dry air. Typhoon Nancy, however, was characterized by relatively high humidity near the surface before landfall (Fig. 11a). During landfall, the water vapor near the surface remained large, and the circulation around the storm remained highly symmetric, although there was a gradual decrease in water vapor around the storm (Figs. 11b–d). Although the airflow crossed the mountains to the west of the storm center, the water vapor of the air mass remained large in the lee of the mountains. The implication is that the influence of orographically induced dry-air intrusion was minor in the case of Nancy because the water vapor was high around the storm. The large water vapor around the storm center could have been maintained by the supply of moist air from the warm sector of the quasi-steady cold front over the Sea of Japan (Figs. 3, 10, and 11). This possibility might be a unique feature of intense typhoons that penetrate into middle latitudes.

Moreover, the storm flow caused dry air to move over the East China Sea, and that air mass was surrounding the storm by 15 September. However, the storm moved to the south of the front, and the region of moist air

(black solid), and (c) longwave (blue) and shortwave (red) radiative heating in DIABQ. Accuracy of asymmetric components in the TADV is verified by the black dotted line (Diag TADV) in (b). The budget results are shown as 6-h averages within a radius of 40 km during landfall.

FIG. 10. Horizontal distributions of precipitable water vapor (color shading; mm) and surface wind speed (contour lines; m s^{-1}) at (a) 1200 UTC 14 Sep, (b) 1800 UTC 15 Sep, (c) 0000 UTC 16 Sep, and (d) 0600 UTC 16 Sep 1961.

within 200 km of the storm center was protected by the front before and after Nancy made landfall (Figs. 10b–d). We believe that the cold front was able to prevent penetration of the dry continental air mass in the storm inner core. This mechanism is somewhat similar to the prevention by outer rainbands of dry-air intrusion into a storm making landfall (Kimball 2006).

The moisture budget of Eq. (8) was calculated for more quantitative examination. Actual change of the volume-averaged precipitable water and sum of the budget equation were -4.27 and -3.89 mm during the approaching period, respectively. The precipitation (-6.34 mm) and lateral advection (-1.55 mm) contributed to reduction of the moisture. The negative contribution fully overcame the evaporation (3.99 mm). The moisture reduction indicated that large amount of water vapor (moisture reservoir) in the mature stage before the approaching was consumed in the approaching stage. Although the

dry-air intrusion from the continent resulted in the negative lateral advection, contribution of the lateral advection was much smaller than that of the precipitation because of inward transport (8.86 mm) of the moisture mainly associated with the subtropical high. Thus, the evaporation from the ocean due to the high wind speed around the well-developed storm and the moisture supply associated with the subtropical high can maintain eyewall updrafts (and the moisture reservoir) during the approaching and landfall periods.

f. Effect of radiative heating

Radiative processes partly controlled the weakening of the warm core in the upper layer, where the $\partial\theta/\partial t$ tended to be negative and there was large radiative cooling in the upper clouds associated with contraction of the eyewall (Fig. 9c). We used a sensitivity experiment to quantitatively estimate the contribution of

FIG. 11. Water vapor mixing ratio (color shading; g kg^{-1}) and horizontal wind (vectors) at 100-m height above the surface at (a) 0900 UTC 15 Sep, (b) 0100 UTC 16 Sep, (c) 0300 UTC 16 Sep, and (d) 0500 UTC 16 Sep 1961. Contour lines denote terrain height. The star indicates the storm center.

radiation to the reduction of storm intensity. The experiment (NORAF) used the same parameter values as the original experiment (CNTRL) but omitted radiative heating of the atmosphere (Table 5). We examined the effect of radiative heating in the upper cloud layer. The surface temperature was therefore still heated by radiation.³

³ In CNTRL, radiation heated not only the atmosphere but also the ocean and ground surface in the one-dimensional thermal diffusion model (Table 3). Although radiative heating was still calculated in NORAF, the heat was not input into the atmosphere but considered in the ground and ocean surface as in CNTRL.

The storm central pressure was approximately the same in NORAF and CNTRL, except during landfall (Fig. 12). The central pressure had decreased in NORAF (i.e., the storm had intensified) by around 0300 UTC 16 September. The maximum difference of the central pressure between CNTRL and NORAF was about 4 hPa at 0400 UTC 16 September. The removal of the longwave radiative cooling in the upper layer of the warm core induced deepening of the central pressure, which is consistent with the budget analysis. However, the rate of storm weakening during landfall was very similar in NORAF and CNTRL. We therefore concluded that the effect of radiative heating on Nancy's intensity

TABLE 5. List of experiments of the sensitivity of the simulated storm intensity to radiation and terrain.

Expt name	Initial time	Remarks
NORAF	0600 UTC 15 Sep 1961	No heating of atmosphere by shortwave and longwave radiation
NOTRN	0600 UTC 8 Sep 1961	Removal of terrain distribution from the land of Japan
NOJPN	0600 UTC 8 Sep 1961	Replacement of the land of Japan with ocean

was minor during landfall, although it could enhance the weakening for 1–2 h.

g. Orography and baroclinicity in the middle latitudes

The potential temperature budget revealed that the simulated warm core was weakened by outward advection (i.e., ventilation) associated with eyewall contraction during landfall. However, TC structure and intensity can often be influenced by local orography (Bender and Kurihara 1986) and baroclinicity in mid-latitudes (i.e., an ET; Jones et al. 2003). Kitabatake (2008, 2011) has indicated that typhoons approaching and making landfall in Japan often experience an ET. Moreover, the baroclinicity can influence the intensity change of TC through the ventilation (Tang and Emanuel 2010). One of most interest is effect of the baroclinicity on the ventilation. In reality, orography and baroclinicity can simultaneously influence storm structure and intensity. Some sensitivity experiments were therefore performed to isolate the influences of 1) local orography and 2) “pure” baroclinicity. The latter effect refers to the influence of prescribed baroclinicity in the middle latitudes, in the absence of any interaction with the land surface and mountainous terrain of Japan, on the simulated storm structure and intensity as the storm approached Japan.

An experiment (NOTRN) in which the coastlines of the islands of Japan were unchanged but the terrain was flatten was performed to examine topic 1. In the other

experiment (NOJPN), the islands of Japan were replaced with ocean to help isolate the influence of baroclinicity on the storm structure and intensity in the absence of interactions with the land surface and mountainous terrain of Japan. The spectral nudging settings used in the CNTRL simulation were used in NOTRN and NOJPN. The nudging allowed the baroclinicity to produce similar structures in CNTRL, NOTRN and NOJPN. We could therefore examine topic 2 by comparing the storms in NOJPN and CNTRL. The NOTRN and NOJPN simulations were started at the same initial time as the CNTRL simulation (unlike the NORAF simulation) to dampen dynamical inconsistency between the simulations due to the removal of orography (Table 5).

The storms were weaker (about 20 hPa) in NOTRN and NOJPN than in CNTRL throughout the period from 14 to 16 September (Fig. 13). The difference of the central pressure appeared after one day from the initial time⁴ (not shown). The storms in NOTRN and NOJPN were similar during 14–16 September but began to differ after landfall (16 September). The difference between the NOTRN and NOJPN simulations and the CNTRL simulation during approach reflected the absence of the mountains of Japan in NOTRN and NOJPN. However, the reductions of intensity and the rates of change of storm intensity in the NOTRN, NOJPN, and CNTRL simulations were quite similar to one another during approach. We could therefore use the sensitivity experiments to examine the influences of local orography and baroclinicity on the intensity change during landfall.

4. Discussion

a. Process of intensity change during landfall

Rapid contraction of the eyewall of a TC can be caused by increase in surface friction and low-level inflow and

FIG. 12. Temporal variation of the central pressure of Nancy in the CNTRL (black) and NORAF (red) simulations. The vertical dashed lines denote the period of landfall.

⁴ Although the reason of the difference is beyond the scope of the present study, one possibility might be influenced by the tiny difference of the storm track in the intensification stage among the three experiments. The tiny track difference might lead to different location and organization of storm inner-core convection. The tiny track difference is mainly induced by slight difference of the spectral nudging among the three experiments due to removal of landmass in Japan.

FIG. 13. Temporal variation of the storm central pressure in the CNTRL (black), NOTRN (blue), and NOJPN (red) simulations. The vertical dashed line denotes the time of landfall. Crosses denote the central pressure based on the JMA.

orographic modification of the low-level inflow during landfall, as reported in previous studies (e.g., Wu et al. 2009). The present tangential wind budget indicated that the weakening of the Nancy's wind speed during landfall was driven by increase in the surface friction and asymmetric advection of tangential wind in low levels during landfall (Fig. 8). We consider that the decrease in the low-level wind can cause the upper-level warm-core cooling through the thermal wind relationship. In other words, the potential temperature budget analysis can indicate details of the adjustment process.

As the eyewall contracts during landfall, the eyewall updraft supplies diabatic heating and adiabatic cooling to a typhoon's warm core. During landfall, the warm core also undergoes significant radiative heating and cooling of the upper troposphere at altitudes above 15 km (Fig. 9). The radiation is caused by collocation of eyewall clouds with the warm core because of eyewall contraction during landfall.

A very important feature of the warm-core weakening is the negative contribution of horizontal asymmetric advection (HAADV). HAAHV is composed of a θ' flux divergence due to asymmetric flows in Eq. (7). Thus, at a radius of 40 km,

$$\begin{aligned} u' < 0 \quad \text{and} \quad \theta' > 0, \\ \text{or} \\ u' > 0 \quad \text{and} \quad \theta' < 0, \end{aligned} \quad (10)$$

and HAAHV is positive.⁵

⁵ Figure 9a is the areal average of Eq. (5) within a 40-km radius of the storm center. Therefore, HAAHV was proportional to the correlation between u' and θ' at a radius of 40 km (i.e., $\overline{u'\theta'}|_{r=40\text{km}}$) from the storm center based on the divergence theorem.

Figure 14 shows plan views of u' and θ' around the storm and azimuthal profiles of them at the radius of 40 km. At altitudes less than 12 km, $u' < 0$ and $\theta' > 0$ to the north and east of the 40-km radius during landfall (Figs. 14a–h). This distribution is consistent with the distribution of vertical flow (Fig. 7). In the north and east sides of the eyewall, updraft associated with contraction of the eyewall was limited below an altitude of 9 km, and subsidence was dominant around an altitude of 9 km (Figs. 7d,f). Adiabatic warming associated with the subsidence then induced $\theta' > 0$ in the sides. From the south to the west of the 40-km radius, $u' > 0$ and $\theta' < 0$ were the dominant conditions. There was a tall updraft as the eyewall contracted. Adiabatic cooling associated with the updraft then induced $\theta' < 0$ at altitudes below 9 km. A positive θ' was entrained into the warm core because of asymmetric inflow ($u' < 0$). However, a negative θ' was detrained because of the asymmetric outflow. The pattern satisfying Eq. (10) indicated intrusion of warm air into the warm core that contributed to intensification of the warm core.

There was also a clearly positive (negative) θ' region to the northeast (northwest) of the storm center beyond a radius of 150 km (Figs. 14a–d). The regions corresponded to cold and warm sectors in the cold front over the Sea of Japan. However, the regions were clearly separated from the warm-core asymmetry within a radius of 40 km. The implication is that the asymmetry of the synoptic-scale temperature structure differed from that of the storm-scale structure.

At altitudes of 12–17 km, $u' > 0$ was mainly collocated with $\theta' > 0$ to the west and south sides of the storm center within a radius of about 40 km (Figs. 14i–p). The signs of u' and θ' were both negative toward the east and north sides of the storm center. This large asymmetric outflow was linked to the updraft of the tall eyewall in the rear of the storm track (Figs. 7d,f). The HAAHV did not satisfy Eq. (10) in that area, and the outflow ($u' > 0$) in the rear of the storm track transported warm air around the storm center beyond a radius of 40 km.

This pattern meant that there was a ventilation effect in the warm core (e.g., Tang and Emanuel 2010). Note that this HAAHV differed from the original concept of warm-core ventilation because of the vertical shear or baroclinicity of the environmental wind, as proposed by Tang and Emanuel (2010). The negative HAAHV in the simulated Nancy was caused by radial flow associated with eyewall contraction at the scale of the inner core. The inner-core asymmetry was associated with the landfall. Then the ventilation of the warm core due to the asymmetry can be an adjustment process associated with the wind speed change

FIG. 14. Horizontal distribution of θ' (color shading; K) and storm-relative asymmetric component of radial velocity (contour lines; m s^{-1}) at (a)–(d) 9- and (i)–(l) 15-km altitudes around the storm. Solid and dashed contour lines denote outflow and inflow about the storm center, respectively. The 40-km radius around the warm core is represented by the black circle. Azimuthal distributions of θ' (blue solid line), u' (blue dashed line), and $-\theta'u'$ (red line) at the radius of 40 km are shown at (e)–(h) 9- and (m)–(p) 15-km altitudes. In the azimuthal views, symbols E, N, W, and S denote east, north, west, and south, respectively.

due to the asymmetric advection of tangential wind during landfall.

b. Orography and baroclinicity in the middle latitudes

The structural changes of Nancy associated with ET were assessed with a cyclone phase space (CPS) diagram (Hart 2003). Figure 15 shows the temporal variation

among the three experiments of three parameters on the CPS diagram: symmetry B , lower thermal wind V_L^T , and upper thermal wind V_U^T as in Hart (2003). In addition, the parameters based on the JRA-55 were also shown as a reference in Fig. 15.

Before landfall, the three storms had typical TC structures (i.e., almost symmetric and warm-core structure).

FIG. 15. Temporal variation of the CPS parameters of (a) B , (b) V_L^T , and (c) V_U^T that characterize the structure of Nancy in the CNTRL (black), NOTRN (blue), and NOJPN (red) simulations, and the JRA-55 (yellow). Horizontal dashed lines denote thresholds for classification of types of typhoons, based on Hart (2003). Low (high) B indicates a cyclone with symmetric (asymmetric) structure. Positive (negative) V_L^T and V_U^T indicates a cyclone with warm (cold) cores in lower and upper layers, respectively. TC is categorized as $B < 10$, $V_L^T > 0$, and $V_U^T > 0$. The vertical dashed line denotes the time of landfall.

After landfall, the parameter B in the three storms rapidly increased (Fig. 15a). This increase indicated that each storm lost its symmetric structure within a radius of 500 km as a result of the baroclinicity of middle latitudes. The evolution of B in each storm was mainly associated with large-scale structure (i.e., baroclinicity) maintained in the spectral nudging. In actual, the evolution of B in the JRA-55 was also similar to those in the three storms.

The values of V_L^T and V_U^T in the CNTRL and NOTRN storms rapidly decreased after landfall (Figs. 15b,c). Figure 16 shows warm-core and eyewall structure during the landfall in the sensitivity experiments. The decrease corresponded to decrease in the temperature anomaly in the warm core (Figs. 6d and 16b) and an increase in

the storm central pressure (Fig. 13). On the other hand, the parameters V_L^T and V_U^T in the NOJPN storm did not drastically change during landfall. In addition, the NOJPN storm maintained a typical TC structure, and the central pressure changed very little after 0000 UTC 16 September. Note that values of V_L^T and V_U^T in the JRA-55 were different from that in the CNTRL storm although tendencies of V_L^T and V_U^T in the JRA-55 were qualitatively similar to those in the three storms. The values in these parameters can highly depend on model resolution as already pointed out by Hart (2003).

The similarity of V_L^T and V_U^T between CNTRL and NOTRN indicated that the storm structure and intensity could be controlled mainly by the increase in surface friction (without a mountainous terrain) and flows at the scale of the storm during landfall. In fact, the Nancy simulated in NOTRN was influenced by the land surface of Japan but no terrain. The eyewall updraft in the storm then rapidly contracted during landfall (Figs. 16a,b). Evolution similar to that of CNTRL was consistent with the previous study of Wu et al. (2009). Moreover, the differences of V_L^T and V_U^T between CNTRL and NOJPN indicated that there was minor influence of “pure” baroclinicity in the middle latitudes on the reduction of the simulated storm intensity during landfall. In fact, the simulated eyewall updraft in NOJPN occurred at almost the same radius before and after landfall in the absence of the land surface and terrain in Japan (Figs. 16c,d).

5. Summary and conclusions

The intensity change of TCs that make landfall is related to sustained periods of severe surface wind and heavy precipitation over land. Full understanding of the changes of structure and intensity during landfall is important for more accurate prediction of severe phenomena. There have been fewer studies of the process in midlatitude countries such as Japan than in low-latitude places such as Taiwan, the Philippines, Cuba, Florida, and northern Australia. In this study, a realistic simulation of very intense Typhoon Nancy (1961) with a nonhydrostatic regional model was used to study the intensity change of an intense typhoon making landfall in Japan. Nancy made landfall on 16 September over Shikoku Island and passed over the western part of Japan, including Osaka. The simulation, which extended over a period of 8.25 days, could reproduce rather well the structure and intensity of Nancy. The results were validated with a variety of available observations over Japan.

During landfall, the structural changes of the simulated storm were apparent in axisymmetric and

FIG. 16. As in Fig. 6 (left), but for the sensitivity experiments: (a),(b) NOTRN and (c),(d) NOJPN. Rows show snapshots at (top) 1000 UTC 15 Sep and (bottom) 0400 UTC 16 Sep 1961.

nonaxisymmetric perspectives. In the axisymmetric structure, a rapid decrease after landfall of the speed of tangential winds at low altitudes within the storm coincided with an increase in friction. As the surface friction increased, low-level inflow also increased over the land. The decrease in the low-level wind speed drives decrease in vertical shear of tangential wind. The vertical shear decrease can modulate radial gradient of temperature in the warm core through the thermal wind relationship. Thus, evolution of the warm core can be

an adjustment associated with the evolution of the dynamical fields during landfall. The rapid contraction of the eyewall after landfall was consistent with previous studies (e.g., Wu et al. 2009; Yang et al. 2011). The eyewall was clearly asymmetric during landfall. A tall and intense eyewall updraft formed toward the rear of the storm's direction of motion. However, a small and weak updraft was apparent only below an altitude of 9 km toward the front of the storm. Weakening of the warm core during landfall was explained by a potential

FIG. 17. Conceptual summary of the process of weakening during landfall in the simulated Nancy. Blue and red components have negative and positive contributions to the warm core. Gray arrows denote secondary circulations in the front (right) and rear (left) sides of the storm. See the text for details.

temperature budget analysis within a radius of 40 km from the storm center.

Figure 17 is a schematic of the storm intensity and inner-core structure changes comprehensively. Coincided with rapid contraction of the eyewall (the gray lines in Fig. 17b), asymmetric momentum advection and increase in the surface friction during landfall caused decrease in the low-level tangential wind. The eyewall rapidly contracted and partly collocated with the warm core (black dashed ellipses in Fig. 17b) during landfall. Diabatic heating was almost balanced by adiabatic cooling in the warm core (Fig. 9a). The heating and cooling were supplied by the contracted eyewall updraft. Previous studies have proposed some weakening associated with the absence of a supply of water vapor (e.g., Tuleya and Kurihara 1978; Kimball 2006) and intrusion of dry air at low altitudes during landfall (Bender and Kurihara 1986). However, much moisture was supplied by enhanced evaporation (i.e., moisture reservoir) over the ocean because of the severe surface winds of the intense storm before landfall in the case of Nancy (region delineated by red dashed lines in Fig. 17a), and that moisture was carried by synoptic-scale flows with the subtropical high (red bold arrows in Fig. 17a). Release of a large amount of latent heat was then maintained

in the storm eyewall, even though a dry air mass from the Eurasian continent (blue bold arrow and region delineated by blue dashed line in Fig. 17a) gradually enveloped the storm. These features associated with this intense typhoon clearly differed from previous studies.

The negative $\bar{\theta}$ tendency at all altitudes weakened the simulated storm during landfall. In particular, we found two large areas of the negative tendency in the lower and upper layers of the warm-core peak at an altitude of about 12 km. Intense adiabatic cooling overcame condensation heating in the eyewall clouds in the lower layer. Then warm-air intrusion to the warm core acted to maintain the warm core because the asymmetric radial inflow coincided with the clearly asymmetric structure of the eyewall.

However, there was no longer large adiabatic cooling in the area of much weakening in the upper layer (Fig. 9b). Instead, there were two large contributions to weakening from 1) radial advection due to asymmetric flow and 2) longwave radiative cooling associated with contraction of the eyewall clouds. The radial advection corresponded to ventilation of the warm core (e.g., Tang and Emanuel 2010). However, the ventilation was induced mainly by collocation of the eyewall updraft with the warm-core air due to contraction of the eyewall as a result of an interaction between the storm-scale flow and land surface during landfall. The warm air was ventilated by the upper-level outflow associated with the contracted updraft (not “pure” environmental baroclinicity). In fact, there was no eyewall contraction in the NOJPN simulation (Figs. 16c,d), and the storm intensity decreased very little (Fig. 13). The ventilation during landfall differed from the warm-core ventilation by the environmental baroclinicity in Tang and Emanuel (2010), although the storm penetrated into midlatitudes. Radiative cooling (NORAF) and local orography (NOTRN) were minor contributors to the reduction of the storm’s intensity (Figs. 12 and 13).

In the case of Nancy, the intensity change was controlled mainly by dynamics and interactions with the land surface (not mountainous terrain) at the storm scale (or inner-core scale) rather than by environmental baroclinicity. This process might also explain the intensity change of TCs making landfall in low latitudes with insignificant baroclinicity. However, this suggestion is based on one case. It is still unclear whether this process can explain the weakening of other intense typhoons as they make landfall.

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