Extratropical Transition of Typhoon Tokage (0423) and Associated Heavy Rainfall on the Left Side of Its Track over Western Japan

by

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Abstract

The evolution of Typhoon Tokage (0423), which made landfall in western Japan and underwent extratropical transition (ET) in October 2004 associated with strong winds and heavy rainfall on both the right and left sides of the storm track, is diagnosed using a gridded regional analysis (RANAL) data set produced by the Japan Meteorological Agency. When Tokage makes landfall, it is located in the right entrance of an upper-tropospheric jet streak and downstream of a synoptic-scale upper trough. The lack of an intense short-wave trough interacting with Tokage is consistent with the fact that the storm did not reintensify after the completion of ET. The strong cyclonic circulation in the lower troposphere associated with Tokage strengthens a preexisting lower-tropospheric front over western Japan, and Tokage eventually transformed into a frontal cyclone in the baroclinic zone. The conditional symmetric instability related to the approach of Tokage to the upper-tropospheric jet stream, as well as the conditional instability in the lower-tropospheric air on the cooler side of the frontal zone over the warm sea surface of the southern Sea of Japan, are considered to contribute to the heavy precipitation along the northern coast of western Japan, which is located on the left side of the storm track. Further study is needed to clarify the impact of air-sea interaction on the intensity and structure of a tropical cyclone in the midlatitude.

1. Introduction

A tropical cyclone (TC) moving into the midlatitudes often evolves into an extratropical cyclone. This extratropical transition (ET) of a TC occurs in nearly every ocean basin that experiences tropical cyclones (Jones *et al.*, 2003). Harr *et al.* (2000) classified ETs in the western North Pacific (WNPAC) basin into two categories in relation to the principal circulation in the midlatitude. Kitabatake (2008) classified WNPAC ETs into three categories on the basis of the frontal pattern at the time of the completion of the ET. These studies imply that not only the characteristics (structure, storm track, etc.) of ET cyclones but also their associated rainfall and wind distributions may vary from case to case in response to their midlatitude environment. Fujibe and Kitabatake (2007) examined the surface wind distributions of TCs landfalling in Japan and suggested that their asymmetry and radii of maximum winds may be related to the environmental baroclinic-

ity. Also the rainfall distribution associated with a TC translating near Japan often becomes asymmetric (*e.g.*, Shimazu, 1998), which is considered to occur principally because of warm frontogenesis in the northeast quadrant of a TC (*e.g.*, Klein *et al.*, 2000). Sakakibara *et al.* (1985) observed a wide rainband in the northeast quadrant (*i.e.*, in front) of a northeastward-moving TC at the ET stage by using Doppler radars and found that convective-scale vertical updrafts were embedded below a mesoscale slantwise updraft. Such a wide rainband is not typical of a mature TC, but it often produces a large amount of precipitation across a wide region of Japan; thus, it is necessary to elucidate the small-scale structure and mechanism of rainbands associated with a TC at the ET stage. These features are different from those of a typical mature TC in the tropical environment and are considered to depend on the midlatitude environment. Therefore the clarification of their relationship to the environment is important for the prediction of the significant features associated with a TC and their resulting disastrous effects. Atallah *et al.* (2007) examined Atlantic TCs landfalling in the United States and concluded that the precipitation may shift to the right or left side of the storm track in response to synoptic-scale forcing.

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As factors affecting the precipitation distribution, they identified (1) a midlatitude trough approaching the TC from the northwest and (2) enhanced ridging downstream of the TC as a result of potential vorticity (PV) redistribution through diabatic heating.

Typhoon (TY) Tokage (0423) made landfall in western Japan in late October 2004 and translated slowly along the Japan Islands. It was associated with a large area of storm-force winds that caused high waves as well as heavy rainfall and the resultant floods. The death toll rose to 98, which was the worst among the 10 landfalling TCs in Japan in 2004. Most of the death toll was due to floods and landslides caused by torrential rainfall. The track of TY Tokage according to the best track data of the Japan Meteorological Agency (JMA) is plotted along with the sea surface temperature (SST) in Fig. 1 (see also Bessho and Kitabatake, 2006). Its lowest central pressure was 940 hPa from 1200 UTC 16 October through 1800 UTC 17 October, when Tokage moved north-westward over the sea south of Japan. After a slight increase in the central pressure and a recurvature near Okinawa, Tokage moved northeastward and made landfall in western Japan at about 0400 UTC 20 October, with a central pressure of 955 hPa and the asymmetric structure of a TC undergoing ET. A maximum surface 10-min-averaged wind speed of 45 m s^{-1} was observed at Murotomisaki, just east of Tokage's center, at 0550 UTC. During its passage along the Japan Islands, Tokage moved generally eastward and weakened rapidly. JMA reported that Tokage completed ET over eastern Japan at 1800 UTC 20 October, which is consistent with the

cyclone phase evolution described by Kitabatake and Bessho (2008). Tokage translated ENE along a region of large SST gradients ($\text{SST} = 18\text{--}24^\circ\text{C}$) after the completion of ET. TY Tokage was quite different from TY Songda, which made landfall in western Japan with a rather symmetric structure, subsequently translated into a higher latitude in the northern Sea of Japan, and then reintensified (Kitabatake *et al.*, 2007; hereafter K07).

Figure 2 shows the precipitation distribution when TY Tokage made landfall in western Japan. Particularly large amounts of precipitation were observed along the southern coast to the east of the cyclone center (*i.e.*, on the right-front of the northeastward-moving storm) at 0600 UTC 20 October (Fig. 2a), which is the typical precipitation distribution associated with a TC and is presumed to be induced by an ascent along the windward side of mountains (*e.g.*, Muramatsu, 1985), although no sign of the eyewall remnants was found near the cyclone center. Subsequently, rather large amounts of precipitation were also observed northeast and northwest of the center of Tokage, that is, on the left side of the eastward-moving storm, at 0900 UTC and 1200 UTC (Figs. 2b and c). Thus, the rainfall continued for almost one whole day and resulted in floods in several regions inland and along the northern coast of western Japan, where the heavy rainfall associated with a mature TC does not usually occur. Therefore, the rainfall associated with a TC in this region should be elucidated. The rainfall associated with Tokage was successfully simulated by Saito *et al.* (2006) and Tsuboki and Sakakibara (2006), but they did not discuss the relationship of the rainfall with the ET of the storm. In addition, the surface wind distribution of Tokage was different from that of many other TCs landfalling in Japan. Although typhoon-force or storm-force winds are often observed primarily on the right side of the track of a TC (*e.g.*, Fujibe *et al.*, 2006), strong northerly winds were observed on the left side of the track of Tokage (*e.g.*, Bessho and Kitabatake, 2006). Fujibe and Kitabatake (2007) suggested that this is characteristic of a TC at a late stage of ET.

The main purpose of this study is to clarify the features associated with the ET of TY Tokage, particularly the heavy rainfall along the northern coast of western Japan, which was on the left side of the storm track. The data are presented in section 2, and the results of the diagnoses in section 3. In subsections 3.1–3.3, the synoptic-scale diagnoses are shown in a manner similar to that in K07, including with potential vorticity (PV) thinking (Hoskins *et al.*, 1985) and a quasi-geostrophic perspective, and TY Tokage is compared with TY Songda, which, as discussed by K07, was characterized by strong winds and reintensification rather than heavy rainfall during the ET stage. Subsections 3.4 and 3.5 describe relatively small-scale diagnoses, including the lower-tropospheric front and vertical/slantwise instability, which likely contributed to the heavy rainfall. The stability in the lower troposphere related to the air–sea temperature difference is

Fig. 1 Track of Typhoon Tokage (0423) during 19–21 October 2004. Open circles (labeled with date/central pressure (hPa)) and black dots (labeled with central pressure) denote the positions at 0000 UTC and at 6-hour intervals (0600 UTC, 1200 UTC, and 1800 UTC), respectively, according to the best track data of the RMSC Tokyo-Typhoon Center. The contour lines denote the sea surface temperature (SST, $^\circ\text{C}$) on 20 October 2004 (see section 2). W, Y, and S are Wajima, Yonago, and Shionomisaki stations, respectively.

discussed in subsection 3.6. Section 4 summarizes these results.

2. Data

The parametric information on TY Tokage used in this study is based on the best track data basically at 6-hour intervals created by the Regional Specialized Meteorological Center (RSMC) Tokyo-Typhoon Center, including center positions, maximum wind speeds, translation directions, and speeds of the tropical cyclone.

The various gridded datasets are obtained from the JMA. The structure of TY Tokage and the baroclinic environment are examined by using the gridded regional analysis (RANAL) dataset with a horizontal resolution of 20 km, which was also used for the diagnosis of TY Songda by K07. The precipitation dataset over the Japan Islands is calculated by the JMA from data of the operational network of radars and raingauges over land and over sea within approximately 40 km of the coast. The horizontal resolution is approximately 5 km. The daily SST data at 0.25° horizontal resolution are obtained by interpolating the SST measured by altimeters on board satellites and in situ observations.

3. Results

3.1 Synoptic evolution

In this subsection, some basic analyses of standard pressure levels derived from the RANAL data set and cloud features in geostationary satellite imagery are used to examine the evolution of TY Tokage undergoing ET during 19–20 October 2004.

Figure 3 presents the sea-level pressure and lower-tropospheric temperature. TY Tokage is moving northward near Okinawa at 1200 UTC 19 October (Fig. 3a), when a frontal zone exists along the southern coast of the Japan Islands, extending southwestward into the East China Sea. Tokage approaches the Japan Islands and is located at the southern periphery of the preexisting frontal zone at 0000 UTC 20 October (Fig. 3b; see also Fig. 3d in Kitabatake and Bessho, 2008). The central pressure increases rapidly after its landfall in western Japan, and the isobars encircling the cyclone are deformed, becoming elliptic (Figs. 3c and d). The horizontal temperature gradients suggest that Tokage is absorbed into the preexisting frontal zone and becomes a frontal wave at 0000 UTC 21 October.

Figure 4 shows infrared (IR) imagery of Tokage approaching the Japan Islands. At 1133 UTC 19 October (Fig. 4a) Tokage has a loosely organized eye near Okinawa, accompanied by annular convection around it. At 2333 UTC 19 October (Fig. 4b), when Tokage is approaching western Japan, deep convection is limited to the east of the cyclone center, and thus the cloud pattern appears asymmetric, although the storm still appears symmetric with regard to the

Fig. 2 One-hour precipitation estimated on the basis of radar reflectivity and rain-gauge measurements. (a) 0500–0600 UTC, (b) 0800–0900 UTC, and (c) 1100–1200 UTC 20 Oct. Barbs are observed surface wind (10-min average) at the stations shown in Fig. 1.

Fig. 3 Sea-level pressure (bold contours, hPa), 900-hPa temperature (thin contours, °C), and its horizontal gradient (shading, $> 2^{\circ}\text{C} (100 \text{ km})^{-1}$) at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct, and (d) 0000 UTC 21 Oct. “X” in panel (d) denotes the center of ex-TY Tokage according to the best track data.

sea level pressure distribution (Fig. 3b). At 1133 UTC 20 October (Fig. 4c) deep convection is still present to the east of the cyclone center, and cirrus clouds north of the cyclone cover almost all of the Sea of Japan, while low-level clouds west of the cyclone center cover western Japan. The low-level clouds west of Tokage are located on the colder side of the cold front extending southwestward from the cyclone center (Fig. 3c). These cloud features, shown in Figs. 4a–c, primarily correspond to Steps 1, 2, and 3, respectively, of the conceptual model proposed by Klein *et al.* (2000). Finally, deep convection near the cyclone center dissipates at 2333 UTC 20 October (Fig. 4d), after the completion of the ET. A band of deep convection is present several hundred kilometers east of the cyclone (ex-Tokage), oriented NE–SW (Fig. 4d). This cloud band does not correspond to strong temperature gradients (Fig. 3d), which suggests that it may be associated with an upper cold front rather than with one in the lower troposphere.

Figure 5 presents the 500-hPa geopotential height and 1000–500 hPa thickness, along with the thickness advection by the 500-hPa geostrophic wind, which show the characteristics in the lower half of the troposphere. Although the

frontal zones at the 900 hPa level (Fig. 3) are intensified along the coast line by the air–sea contrast, broad zonal baroclinic zones appear in the thickness field in the midlatitudes (Fig. 5). At 1200 UTC 19 October (Fig. 5a), Tokage is near Okinawa, as shown in Fig. 3a, while a weak 500-hPa trough exists southwest of the Korean Peninsula, accompanied by a thermal trough along the eastern coast of the Chinese mainland. Tokage is located just downstream of the trough, associated with warm/cold advection to its northeast/southwest. At 0000 UTC 20 October (Fig. 5b) the warm/cold advection associated with Tokage increases in area and magnitude. The trough just absorbing Tokage is located downstream of the thermal trough, which implies the baroclinic development of the midlatitude trough (*e.g.*, Carlson, 1991, Chap. 7). At 1200 UTC 20 October (Fig. 5c), the trough absorbing Tokage is in phase with the thermal trough, which suggests that the possible baroclinic development is completed. At 0000 UTC 21 October (Fig. 5d), the area of warm advection advances to the east of the Japan Islands, while the area of cold advection is retarded over the southern coast of the Japan Islands. Note that the 500-hPa trough is relatively weak in this case compared to other ET cases (*e.g.*, TY Songda), although Atallah

Fig. 4 GOES-9 IR imagery at (a) 1133 UTC 19 Oct, (b) 2333 UTC 19 Oct, (c) 1133 UTC 20 Oct, and (d) 2333 UTC 20 Oct. "X" denotes the cyclone center according to the best track data.

et al. (2007) states that the interaction of TCs with an intense trough results in a left side of track (LOC) precipitation distribution.

To the north of Tokage, another 500-hPa trough is translating eastward along 45–50°N and reaches the Sea of Okhotsk at 1200 UTC 20 October (Fig. 5c), accompanied by a weak surface low (Fig. 3). The location of this trough relative to Tokage is the northeastern pattern defined by Harr *et al.* (2000), where a strong zonal flow prevents direct interaction between a decaying TC and the midlatitude circulation. This pattern is in contrast to that of the ET of TY Songda, which according to K07 showed the northwestern pattern of Harr *et al.* (2000), where a strong southerly flow downstream of a trough interacts with a TC.

Figure 6 shows the 250-hPa geopotential height and isotachs, which reveal the relationships between upper-tropospheric jet streaks and the TC. An intense jet streak (A) is present over northern Japan at 1200 UTC 19 October (Fig. 6a), when Tokage is near Okinawa. Tokage approaches the right entrance of the jet streak A at 0000 UTC 20 October (Fig. 6b), while another weak jet streak (B) is located over southern China, and a third jet streak (C) is moving eastward from Mongolia to the Maritime Province of Siberia. Jet streaks B

and C both weaken rapidly during the ET of Tokage (Figs. 6c and d), and a fourth jet streak (D), to the west of Tokage, intensifies. The weakening of the jet streak B to the southwest of Tokage, in particular, results in the absence of the double-jet pattern favorable to a reintensification of an ET cyclone (e.g., K07). This is consistent with the fact that Tokage weakened rapidly after its landfall. The trough moving eastward along 45–50°N (Fig. 5) is indistinct at 250 hPa (Fig. 6), although it may contribute to the strengthening of the jet entrance over the northern Sea of Japan (Fig. 6c).

The evolution of low-level fronts associated with Tokage (Fig. 3) appears to be consistent with an open-wave (OW) type of cyclone (Kitabatake, 2008), although the 925 hPa equivalent potential temperature (θ_e) criterion for the classification of an ET is not applicable to the present high-resolution analyses. The fact that Tokage moves northeastward and completes ET at 36°N, south of the upper-tropospheric jet stream (Figs. 6c and d) is also consistent with the characteristics of an OW cyclone. For comparison, TY Songda, which moved northward and completed ET north of 40°N between two jet streaks (*i.e.*, the double-jet pattern), is similar to a seclusion-occlusion (SO) cyclone (Kitabatake, 2008).

Fig. 5 500-hPa geopotential height (bold contours, m) and 1000–500 hPa thickness (thin contours, m) at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct, and (d) 0000 UTC 21 Oct. Thin solid/dotted lines with shading denote positive/negative thickness advection (± 0.002 and ± 0.004 m s⁻¹).

3.2 Dynamic tropopause – A PV thinking approach

A PV perspective is helpful when discussing the interaction between an upper-tropospheric disturbance and a surface cyclone. Thus, Fig. 7 presents the potential temperature (θ), pressure, and winds on the dynamic tropopause (DT), defined as a 2-PVU (potential vorticity unit; 1 PVU = 10^{-6} K m² s⁻¹ kg⁻¹) potential vorticity surface, which is overlaid on the 850-hPa PV distribution, to represent lower level disturbances, including Tokage. The jet streaks shown in Fig. 6 correspond to the large pressure (or θ) gradients shown in Fig. 7. The DT ascends/descends north/west of Tokage as it moves northeastward until 1200 UTC 20 October (Fig. 7c), which is a common occurrence for the DT near a TC undergoing ET (e.g., Atallah and Bosart, 2003; K07) and one of the factors characterizing an LOC storm (Atallah *et al.*, 2007). A region of large pressure (or θ) gradients, that is, a tropopause folding, downstream of the upper trough over the Chinese mainland extends west of the surface cyclone at 1200 UTC 20 October (Fig. 7c). Positive pressure advection and nega-

tive θ advection on the DT over the TC imply an interaction between the TC and the upper-tropospheric disturbances. At the same time, the 850-hPa positive PV anomaly associated with Tokage become deformed over central Japan, suggesting a change in the structure of the cyclone, and small portions of the positive PV anomaly align along the region of strong temperature gradients (Fig. 3c) from the center of Tokage to the northeast over the sea off eastern Japan; thus, these PV anomalies are attributed to latent heat release in the frontal zone. Subsequently, the low-level PV anomaly associated with Tokage dissipates at 0000 UTC 21 October (Fig. 7d), which corresponds to the dissipation of deep convections in satellite imagery (Fig. 4d).

Strong pressure (or θ) gradients (Fig. 7c) are associated with a significant jet entrance over the Maritime Province of Siberia, and the northern Sea of Japan at 1200 UTC 20 October (see also the jet streak A in Fig. 6c), and they rapidly weaken over northern Japan (Fig. 7d) at 0000 UTC October. This weakening of the pressure (or θ) gradients (Fig. 7d) also

Fig. 6 Wind speed (bold contours, beginning at 30 m s^{-1}) and geopotential height (thin contours, m) at 250 hPa at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct, and (d) 0000 UTC 21 Oct.

corresponds to the dissipation of the deep convections associated with Tokage, implying that the strong gradients (Figs. 7a–c) are intensified by the “tropopause lifting” caused by the deep convections associated with Tokage.

TY Songda interacted with a significant upper-tropospheric disturbance that was several hundred kilometers in diameter and associated with the DT descending below 400 hPa, inducing the (re-)intensification of Songda at midlatitudes (K07). In the case of Tokage, two significant disturbances associated with the DT descending below 400 hPa occurs (Fig. 7): one, over Sakhalin (Fig. 7c), corresponds to the shortwave trough over the Sea of Okhotsk shown in Fig. 5c; the other is over Northeast China (Fig. 7d) and corresponds to the trough over the Chinese continent shown in Fig. 5d. Both of these disturbances are located 2000 km from the center of Tokage and not likely to interact with it, which is consistent with the fact that Tokage did not reintensify during its passage near Japan.

3.3 Diagnosis of quasi-geostrophic omega forcing

In this subsection, a quasi-geostrophic perspective is used to diagnose the environmental synoptic-scale forcing that may modify the structure of the TC and the associated

precipitation distribution. The quasi-geostrophic omega equation proposed by Trenberth (1978) is used to discuss the synoptic-scale interaction between Tokage and the midlatitude environment in comparison with that of TY Songda (see subsection 4.4 of K07). The omega equation with Trenberth’s approximation is as follows:

$$\frac{R_d}{p} \bar{s} \nabla_p^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} \approx 2 f_0 \frac{\partial \mathbf{V}_g}{\partial p} \cdot \nabla_p (\zeta_g + f), \quad (1)$$

where R_d is the gas constant for dry air, $\bar{s}$ is the local horizontal average of static stability, $\mathbf{V}_g$ is the geostrophic wind, and ζ_g is the geostrophic vorticity. Thus, the forcing term on the right-hand side of Equation (1) represents horizontal advection of the geostrophic absolute vorticity by the thermal wind.

Figure 8 depicts the 500-hPa geostrophic vorticity advection by the 300–850 hPa thermal wind (*i.e.*, the right-hand side of Eq. 1). A dipole pattern of positive/negative vorticity advection characteristic of a baroclinic disturbance is indistinct until 0000 UTC 20 October (Figs. 8a and b) but appears to be associated with Tokage at 1200 UTC 20 October (Fig. 8c) when Tokage begins to be absorbed into the frontal zone (Figs. 3c and 5c). The dipole decouples at 0000 UTC 21 October (Figs. 8d and 5d). These suggest that Tokage is under-

Fig. 7 Potential temperature (color scale, K), pressure (thin contours, hPa), and wind (arrows, m s^{-1}) on the dynamic tropopause (defined as the 2-PVU potential vorticity surface; $1 \text{ PVU} = 10^{-6} \text{ K m}^2 \text{ s}^{-1} \text{ kg}^{-1}$) at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct, and (d) 0000 UTC 21 Oct. Potential vorticity at 850 hPa is also shown in each panel (bold contours; contour interval, 2 PVU).

going ET at 1200 UTC 20 October but is poorly organized as a baroclinic low by 0000 UTC 21 October.

The dipole pattern of vorticity advection by the thermal wind (Fig. 8c; 1200 UTC 20 October) corresponds to that of thermal advection (shown in Fig. 5c). At 1200 UTC 19 October and 0000 UTC 20 October, although the dipole pattern of thermal advection can be seen in Figs. 5a and b, the vorticity advection by the thermal wind does not appear as a dipole (Figs. 8a and b). The vorticity advection term in the classical omega equation (not shown) may explain why the dipole pattern in the thermal advection is almost canceled out until 0000 UTC 20 October. Tokage arrives at the southern periphery of the upper-tropospheric jet stream at 0000 UTC 20 October (Fig. 6b) and moves into it at 1200 UTC (Fig. 6c), implying that the upper-tropospheric vorticity advection near Tokage is likely at a maximum at 0000 UTC 20 October.

Some other features related to Tokage and midlatitude disturbances are also apparent in Fig. 8. The midlatitude trough over the East China Sea seen in Fig. 5a corresponds to a positive vorticity anomaly near Shanghai in Fig. 8a, associated with positive/negative advection downstream/upstream of the vorticity anomaly. However, the vorticity anomaly as-

sociated with the midlatitude trough is much weaker than that of Tokage itself; thus, the vorticity advection by the thermal wind (and the resultant vertical forcing) associated with the midlatitude trough seems to affect Tokage very little, which is in contrast to the ET of Songda and consistent with the fact that Tokage did not reintensify as it passed over the Japan Islands.

The vertical forcing suggested by the vorticity advection by the thermal wind (Fig. 8c) is also consistent with the fact that the deep convections associated with Tokage are limited to areas east of the cyclone center (see Fig. 4c). The dipole pattern of positive and negative vorticity advection by the thermal wind (Fig. 8c) is similar to the LOC storm reported by Atallah *et al.* (2007). Thus, the magnitude of the thermal wind (*i.e.*, vertical wind shear or horizontal thermal gradients), rather than the midlatitude trough, may play an important role in the redistribution of precipitation. However, the downward forcing in the western semicircle of the cyclone center (Fig. 8c) is inconsistent with the precipitation northwest of the cyclone center (Fig. 2). This suggests that the shallow clouds west of the cyclone center originate in the lower troposphere, which is discussed in the following sub-

Fig. 8 500-hPa geopotential height (bold contours; contour interval, 100 m), thermal wind for the 300–850 hPa layer (arrows, m s^{-1}), 500-hPa geostrophic absolute vorticity (thin contours; contour interval, 0.0001 s^{-1}), and its advection by the thermal wind (color scale, 10^{-8} s^{-2}) at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct and (d) 0000 UTC 21 Oct.

sections.

3.4 Deformation of lower-tropospheric flows and frontogenesis

Whereas upper-tropospheric disturbances, including jet streaks and short-wave troughs, primarily contributed to the ET of TY Songda (K07), they seem contribute less to the ET of TY Tokage. Since the above discussion suggests that the lower-tropospheric baroclinicity dominantly affects the ET of Tokage, frontogenesis in the lower troposphere is diagnosed (Fig. 9). Two-dimensional frontogenesis F_H is expressed as follows:

$$F_H = \frac{1}{2} |\nabla_p \theta| (D \cos 2\alpha - \text{Div}) \quad (2)$$

where D is the total deformation of the horizontal flow (represented as the length of the green line segments in Fig. 9), α is the angle between the isotherms and the axis of dilatation (the orientation of green line segments), and Div is the horizontal divergence (color shading) (e.g., Bluestein, 1993, p. 248–253).

Temperature gradients are large along the southern coast of the Japan Islands at 1200 UTC 19 October (Figs. 3a and 9a), when Tokage is located near Okinawa in the subtropical airmass. Although the deformation of the horizontal flow, which is represented by the length of the green line segment in Fig. 9, is relatively weak, the axes of dilatation make a small angle with the isotherms (i.e., $\alpha \approx 0^\circ$ in Eq. (2)) over western Japan, indicating that the total deformation due to the horizontal flow contributes to frontogenesis. Horizontal convergence along the southern coast of western Japan also contributes to frontogenesis. The 900-hPa convergence/divergence on the warm/cold side of the frontal zone corresponds to the thermally direct circulation associated with the frontogenesis.

The deformation over western Japan increases in magnitude at 0000 UTC 20 October (Fig. 9b); thus, the strong cyclonic circulation associated with Tokage induces frontogenesis and the associated low-level convergence as it approaches the preexisting frontal zone. The warm frontogenetic region with strong deformation and associated low-level convergence shifts to eastern Japan at 1200 UTC 20 October

Fig. 9 Temperature (bold contours), streamlines (thin contours), total deformation with the direction of the axis of dilatation of the horizontal wind (green line segments, 10^{-5} s^{-1}), and horizontal divergence (color scale, 10^{-5} s^{-1}) at 900 hPa at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct, and (d) 0000 UTC 21 Oct.

(Fig. 9c). The axes of dilatation make a large angle with the isotherms (*i.e.*, $\alpha \approx 90^\circ$ in Eq. 2) over western Japan (Figs. 9c and d), implying frontolysis, where divergence is also dominant. This 900-hPa convergence/divergence east/west of the cyclone center is mostly consistent with the forcing of vertical motion, depicted in Fig. 8. However, the divergence west of the cyclone center and the associated descendant motion seem still to be inconsistent with the precipitation in the northwest (left-rear) quadrant of Tokage (Fig. 2c).

3.5 Three-dimensional structure of precipitation over western Japan

As mentioned in section 1, the heavy rainfall disaster occurred primarily on the left side (*i.e.*, to the north) of the storm track of Tokage during the passage over the Japan Islands. The discussion in subsections 3.3 and 3.4 suggests that Tokage is associated with ascent related to the warm frontogenetic circulation east (in front) of the eastward-moving cyclone during its passage over the Japan Islands. The principal heavy rainfall east of the TC center (Fig. 2) is consistent with the asymmetric distribution of vertical forcing (shown in Fig. 8c) and is considered to be associated with the warm frontogenesis. However, Fig. 8c also shows that the downward

forcing is dominant in the northwestern (*i.e.*, left-rear) quadrant of Tokage, so the precipitation along the northern coast of western Japan (Figs. 2b and c) needs further explanation.

Figure 10 depicts the moisture flux convergence associated with the circulation of Tokage. The large precipitation amount along the southern coast of western Japan and over inland western Japan at 0600 UTC (Fig. 2a) is consistent with the moisture convergence at the 925-hPa level (Fig. 10a), and the southerly flow east (*i.e.*, in front) of the cyclone center is the primary contributor to this convergence. It is apparent that the precipitation along the northern coast of western Japan (Fig. 2a) is generated in relation to northeasterly flow in a cool airmass at the 925-hPa level over the southern Sea of Japan, as well as to the moisture convergence along the frontal zone at the 700-hPa level (Figs. 10a and b). At 1200 UTC, moisture convergence along the northern coast of western Japan occurs in association with a northeasterly airflow at 925 hPa (Fig. 10c), rather than at 700 hPa (Fig. 10d). The temperature of this 925-hPa northeasterly airflow in the southern Sea of Japan is below 14°C , which is much cooler than the air associated with a TC, and the northeasterly airflow resembles a cold conveyor belt (*e.g.*, Schultz, 2001), as is often found on the poleward side of an extratropical

Fig. 10 Temperature (contours, °C), moisture flux (arrows, $\text{g m kg}^{-1} \text{s}^{-1}$) and its divergence (color scale, $\text{g kg}^{-1} \text{s}^{-1}$) for (a) 925 hPa at 0600 UTC, (b) 700 hPa at 0600 UTC, (c) 925 hPa at 1200 UTC, and (d) 700 hPa at 1200 UTC 20 Oct.

cyclone. Thus, the precipitation in the northwest (rear-left) quadrant of the cyclone center is considered to be generated primarily in relation to northeasterly flow in a cool airmass, which is unusual for precipitation associated with a TC.

Since the precipitation on the cooler side of a cyclone can be related to the frontal zone, it is assumed to be induced by the conditional symmetric instability that is equivalent to the negative saturated equivalent potential vorticity (hereafter SPV; $\text{SPV} = g\eta \cdot \nabla\theta_{es}$, where g is gravity, η is the 3-dimensional absolute vorticity vector, and θ_{es} is the saturated equivalent potential temperature). Figures 11a and 11b depict SPV at the 700-hPa level. The saturated equivalent potential temperature θ_{es} shown in these figures is consistent with the temperature (see Figs. 10b and d). SPV associated with Tokage has an annular shape at 0600 UTC (Fig. 11a) and transforms into an elongated shape along the frontal zone by 1200 UTC (Fig. 11b). Figures 11c and 11d are vertical cross sections along the bold black lines in Figs. 11a and 11b, respectively, over the region of heavy rainfall in western Japan. The 2-PVU contour in the upper troposphere (Fig. 11b) coincides with the 2-PVU surface in Fig. 7c. The absolute momentum $M = v + f\bar{x}$ is depicted in Figs. 11c and 11d, where f is the

Coriolis parameter, x is the distance along the horizontal axis in each figure, and v is the wind speed normal to the cross section. In Fig. 11c the M surface can be traced from the SPV maximum in the lower troposphere to another SPV maximum in the upper troposphere. This implies that the lower-tropospheric disturbance (the northeastern portion of Tokage) is interacting with the upper-tropospheric midlatitude jet and trough system, which is common for a TC at the ET stage (e.g., Atallah and Bosart, 2003). Negative SPVs are present below the 800-hPa level on both the southeastern and northwestern sides of the SPV maximum (Fig. 11c). These negative SPVs are related to the conditional instability ($\partial\theta_{es}/\partial z < 0$). Another negative SPV region in the 800–600 hPa layer over 134°E is conditionally stable and thus indicates the conditional symmetric instability (Schultz and Schumacher, 1999). The warmer side of the frontal zone based on θ_{es} is almost saturated (relative humidity, $> 80\%$) and is conditionally symmetrically stable ($\text{SPV} > 0$) in the middle troposphere. These are consistent with the implication that the precipitation east (in front) of the cyclone center at 0600 UTC is induced by the thermally direct circulation associated with the warm frontogenesis (subsection 3.4).

Fig. 11 (a) Saturated equivalent potential vorticity (color scale, PVU), saturated equivalent potential temperature (solid contours, every 10 K), relative humidity (dashed contours, 30% and 80%) and horizontal wind (arrows, m s^{-1}) at 700 hPa at 0600 UTC 20 Oct. (b) Same as (a) except at 1200 UTC. (c) Cross section at 0600 UTC 20 Oct along line A-A' in panel (a). Saturated equivalent potential vorticity (color scale, PVU), saturated equivalent potential temperature (bold contours; contour interval, 5 K), absolute momentum (thin contours; contour interval, 10 m s^{-1}), and relative humidity (dashed contours, 30% and 80%). (d) Same as (c) except at 1200 UTC along line B-B' in panel (b).

Figure 11d is the cross section west (in the rear) of the center of Tokage at 1200 UTC. Dry air (relative humidity, $< 30\%$) is dominant on the warmer side of the frontal zone, consistent with the downward motion (suggested in Fig. 8c). The moist air ($> 80\%$) in the middle troposphere is limited to near 134°E on the cooler side of the frontal zone, which corresponds to the precipitation in the northwestern (left-rear) quadrant of the cyclone (Fig. 2c). SPV is negative in the 700–600 hPa layer over $134\text{--}135^\circ\text{E}$ in the moist region, suggesting that the mid-tropospheric moist air and the related precipitation in the northwestern (left-rear) quadrant of the cyclone is related to the conditional symmetric instability. It also suggests that the negative SPV on the cooler side of the frontal zone over 134°E at 0600 UTC (Fig. 11c) and the possible conditional symmetric instability may also contribute to the precipitation in the northeastern (left-front) quadrant of the cyclone.

The tilt of the M surfaces (Figs. 11c and 11d) is related

to the vertical shear of the southwesterly component in the northwestern (left-rear) quadrant of Tokage, that is, a large wind speed of the southwesterly component in the upper troposphere versus a small or negative wind speed in the lower troposphere. Thus, the increase in the tilt of the M surface in the northwestern (left-rear) quadrant of the TC is related to an increase in the *southwesterly* component of the jet stream in the upper troposphere, as well as to an increase in *northeasterly* winds in the lower troposphere. The latter corresponds to the increase in the cyclonic wind speed to the *left* of the cyclone center relative to the storm track at the ET stage (Fujibe and Kitabatake, 2007). The strong northeasterly wind in the lower troposphere is also likely to induce a windward ascent of moist unstable air along the northern coast of western Japan, which would result in convection in the conditionally unstable layer of the cool airmass.

To discuss the actual stability, vertical profiles of the operational radiosonde observations around Tokage at 0600

UTC and 1200 UTC 20 October are shown in Fig. 12. For simplicity, equivalent potential temperature is not shown in the figure, but the boundary layer at each station is almost saturated. It is apparent that the tropospheric airmass observed at Shionomisaki is subtropical, whereas a cooler airmass is observed at Wajima and Yonago in the lower troposphere. The lower-tropospheric cooler air is moist and conditionally unstable, that is, with an almost uniform potential temperature (θ) and negative vertical gradients of the saturated equivalent potential temperature ($\partial\theta_{es}/\partial z < 0$) in the boundary layer. The surface air temperature is 17–18°C at Wajima and Yonago, which is cooler than the SST (about 20–22°C) in the adjacent sea (Fig. 1). This suggests that the heat and moisture fluxes from the sea surface to the atmosphere contribute to the reduction in the stability in the lower troposphere, not only in the warm airmass along the southern coast but also in the cooler airmass along the northern coast. At 1200 UTC (Fig. 12b), the conditionally unstable layer ($\partial\theta_{es}/\partial z < 0$) in the lower troposphere increases in height at Wajima and Yonago, and the lower-tropospheric θ decreases. These changes are considered to be associated with cold advection in the lower troposphere.

A relatively stable layer (*i.e.*, a frontal layer) at Yonago is found at 679–598 hPa level at 0600 UTC (Fig. 12a) and rises to the 663–567 hPa level at 1200 UTC (Fig. 12b), which is consistent with the strong horizontal temperature gradients south of Yonago at 700 hPa (Figs. 10b and d). A frontal layer is indistinct in the vertical profile at Wajima, although a distinct frontal zone on the isobaric surface (*e.g.*, Figs. 10a–d) is present over central Japan. A relatively stable layer appears at the 790–727 hPa level at Wajima at 0600 UTC (Fig. 12a), associated with a less stable layer (727–635 hPa) above it; these are attributed to a leeward subsidence north of the mountains in central Japan rather than to a frontal layer. This indistinct frontal layer in the mid-troposphere northeast of the cyclone center contrasts with the distinct frontal layer in the lower level associated with TY Songda (see the profile for Wakkanai in Fig. 8a of K07), and the frontal layer may be weakened by the vertical mixing. These facts suggest that the convections associated with Tokage occur not only in the warm subtropical airmass but also in the cooler midlatitude airmass.

The equilibrium level for free convection of a parcel near the surface at Shionomisaki ($\theta_{es} \approx 365$ K) is estimated to be about 100 hPa (16 500 m) at 0600 UTC (Fig. 12a), which is consistent with the fact that the elevation of the echo top observed by the operational radar network (data not shown) is 14–16 km for the deep convection near the center of Tokage located just west of Shionomisaki. However, the equilibrium level for a parcel near the surface at Wajima ($\theta_{es} \approx 330$ K) is estimated to be about 450 hPa (6500 m) at 0600 UTC (Fig. 12a), while the observed elevation of the echo top is about 8 km near Wajima, suggesting that the observed echo top along the northern coast of Japan may be due to a slant-

Fig. 12 Vertical profiles of potential temperature (bold lines, K) and saturated equivalent potential temperature (thin lines, K) at Wajima, Yonago, and Shionomisaki stations at (a) 0600 UTC, and (b) 1200 UTC 20 Oct. Locations of the observation sites are shown in Fig. 1.

wise ascent in a warm airmass rather than vertical convection in a cool airmass.

The equilibrium level for a parcel near the surface at Yonago at 1200 UTC ($\theta_{es} \approx 330$ K in Fig. 12b) is estimated to be about 450 hPa (6500 m). This is consistent with the fact that the observed echo top is at about 6 km for the relatively shallow clouds near Yonago, and thus suggests that the clouds west (in the rear) of Tokage at 1200 UTC 20 October may be generated in the cool airmass as a result of conditional (vertical) instability below the 800-hPa level, in spite of the synoptic-scale downward forcing (subsection 3.3). The 800–600 hPa layer is conditionally neutral (Fig. 12b), but the

negative SPV near Yonago (Figs. 11b and d) suggests conditional symmetric (slantwise) instability. Studies on the possible coexistence of slantwise convection and vertical (gravitational) convection are reviewed by Schultz and Schumacher (1999, section 5). They emphasize that the conditional instability is a special case of the conditional symmetric instability.

3.6 Air–sea thermal contrast

The results presented in the previous subsection indicate that the precipitation on the left side of Tokage's track is dominated by both frontal circulation and instability in the middle and lower troposphere. More careful examination of the vertical profiles at Yonago and Wajima (Fig. 12) shows that the potential temperature (θ) and the mixing ratio (not shown in the figure) are almost uniform in the boundary layer, suggesting that the mixed layer is formed in the cool boundary layer over the relatively warm sea surface (e.g., Garratt, 1992). Thus, the air–sea thermal contrast is assumed

to contribute to the lower-tropospheric instability in the northern semicircle of Tokage and the associated heavy rainfall.

The air–sea thermal contrast (ASC) shown in Fig. 13 is defined here as the difference between the saturated equivalent potential temperature for the SST and θ_e at 925 hPa, which is almost equivalent to θ_{es} at the same level in the boundary layer over the sea surface, as mentioned in the previous subsection. A circle with a radius of 500 km from the center of Tokage (depicted in each panel of Fig. 13) indicates the region of the primary surface cyclonic circulation. At 1200 UTC 19 October, the ASC is positive all over the sea surface (Fig. 13a), where the ASC is less than 10 K near the cyclone center and along the surface frontal zone (Fig. 3a), suggesting that the vertical mixing of the air in the boundary layer reduces the ASC. (Note that the 925-hPa isobaric surface is lowered toward the sea surface near the center of the TC.) At 0000 UTC 20 October (Fig. 13b), the value of the ASC is still small near the cyclone center, while it increases

Fig. 13 Equivalent potential temperature (contours, K) and wind (arrows, m s^{-1}) at 925 hPa, and the difference between the saturated equivalent potential temperature at the SST and the 925 hPa equivalent potential temperature (color scale, K) at (a) 1200 UTC 19 Oct, (b) 0000 UTC 20 Oct, (c) 1200 UTC 20 Oct, and (d) 0000 UTC 21 Oct. The black circle in each panel is drawn with a radius 500 km around the surface cyclone center.

west of the cyclone. At 1200 UTC (Fig. 13c) when the cyclone center is located on the main island of Japan, small ASC values occur along the cold frontal zone extending to the south of eastern Japan, and large ASC values are spread near western Japan and the southern Sea of Japan. The increase in the ASC west of the cyclone corresponds to the cold advection shown in Fig. 5, and is also consistent with the lower-tropospheric conditional instability, shown in Fig. 12. Negative values of the ASC over the sea off the eastern coast of Japan are related to a relatively warm airflow over a cool sea surface (Fig. 1), where the cooling of lower-tropospheric air by the cold sea surface likely generates a statically stable boundary layer.

The increase in the ACS west of the cyclone is consistent with the description by Ritchie and Elsberry (2001) of surface sensible and latent fluxes from a relatively warmer SST with advection of cooler air to the west of a TC at the transformation stage of ET. Thorncroft and Jones (2000) investigated the role of surface fluxes on the ET of Atlantic hurricanes and showed that when a hurricane moves over cold water the downward surface fluxes toward the cold ocean act to produce a stable boundary layer that destroys the warm core of the hurricane at low levels. Thorncroft and Jones referred to the remnants of a TC above the stable boundary layer as thermally “decoupled” from the ocean. An example of a profile suggesting convection decoupled from the underlying sea surface by a cool boundary layer is found in the case of TY Songda (Fig. 8a of K07), where deep convection is likely to occur in the conditionally unstable warm airmass in the mid-troposphere above the stable layer in the lower troposphere, as well as in the case of TY Higos (0221; Kitabatake, 2008).

Figure 14 presents the time sequence of the ASC and the SST averaged within a 500-km radius from the center of Tokage (see the black circle in each panel of Fig. 13; the effect of the land surface is ignored). While a tropical SST of 26°C is considered to be a necessary condition for the genesis of a TC (Gray, 1975), each of a decrease in the SST and an increase in the SST gradients in the midlatitude is considered as one of the necessary conditions for the weakening of the TC and its structural change to an extratropical cyclone (e.g., Jones *et al.*, 2003). Tokage travels over the tropical sea surface ($> 26^{\circ}\text{C}$) at least until 1200 UTC 19 October, and subsequently moves over a cooler sea surface. The averaged SST is 23.4°C (Fig. 14) at 1800 UTC 20 October, when Tokage completed ET (Fig. 1). The cool SST underlying the TC seems unfavorable for sustaining the TC intensity, and it is inconsistent with the fact that Tokage made landfall in western Japan as an intense typhoon (Fig. 1).

Sensible and latent heat fluxes from the sea surface to the air depend on the air–sea thermal contrast rather than on the SST itself, and thus the ASC likely contributes to the intensity and structure of the TC. The ASC averaged within 500 km of the cyclone center is approximately constant (15 K)

Fig. 14 Time sequence of SST ($^{\circ}\text{C}$, solid line) and the difference between the saturated equivalent potential temperature at the SST and the 925-hPa equivalent potential temperature (K, dashed line) averaged within a radius of 500 km from the center of Tokage (see the black circles in Fig. 13).

while Tokage travels over the tropical sea surface; subsequently, it increases over the cooler sea surface until 1200 UTC 20 October, in association with a decrease in the 925-hPa θ_e north and west of the cyclone center (Figs. 13b and c). In spite of the cold advection in the lower troposphere and the decrease of SST, Tokage made landfall in western Japan as an intense typhoon (Fig. 1).

After 1200 UTC 20 October, the ASC rapidly decreases along with the decreasing SST (Fig. 14), which is related to the increase in the area of the negative ASC northeast of Tokage (Figs. 13c and d), where a high θ_e air overlies the cooler sea surface to the east, off eastern and northern Japan. The averaged ASC falls below 15 K at 0000 UTC 21 October and finally below 0 K at 1200 UTC (Fig. 14), after deep convective clouds near the cyclone have dissipated (Fig. 4d). This suggests that the heat transfer from the sea surface to the air decreases and finally becomes reversed; thus the circulation associated with Tokage becomes thermally decoupled from the sea surface after the completion of the ET.

However, it should be also noted that the low-level PV anomalies near 40°N, 145°E along the warm frontal zone east of Japan (subsection 3.2; Figs. 7c and d) are located along the 0–10 K ASC contours north of the negative ASC regions (Figs. 13c and d, respectively). This indicates that these low-level PV anomalies are diabatically generated in response to the interaction between the cool airmass on the poleward side of the frontal zone and the SST, which is much cooler than that of the subtropical ocean but warmer than the overlying air in the boundary layer. It also suggests that the precipitation on the left (*i.e.*, poleward) side of Tokage translating over the Japan Islands is produced in an environment similar to the frontal rainfall after the completion of ET rather than by a mature stage TC.

4. Summary and remarks

The evolution of TY Tokage (0423), which caused torrential rainfall and the resultant floods in western Japan in October 2004, is discussed in comparison with that of TY Songda (0418) by using the gridded datasets produced by the JMA. Tokage has an asymmetric structure of clouds and precipitation that is primarily distributed east and north of the cyclone center, that is, in front and on the left side of the storm, when Tokage makes landfall in western Japan as a strong typhoon. Tokage weakens after its landfall and underwent ET without reintensification, being apparently absorbed into a frontal zone preexisting over the Japan Islands. Tokage is located in the right entrance of a jet streak downstream of an upper trough. However, the upper trough west of Tokage is relatively weak; thus, the ET of Tokage appears to be the northeastern pattern defined by Harr *et al.* (2000) in contrast to the ET of TY Songda, which K07 indicated to have been the northwest pattern of Harr *et al.* (2000).

The interaction between Tokage and the upper-tropospheric disturbances are diagnosed with PV thinking within a quasi-geostrophic framework. The upper trough approaching Tokage from the west has a large wavelength compared with the horizontal scale of Tokage itself, as well as compared with the short-wave trough that interacted with TY Songda; thus, the interaction between Tokage and the upper disturbances is presumed to be smaller than in the case of TY Songda.

Tokage appeared to transform into a frontal cyclone in the baroclinic zone preexisting along the Japan Islands. The diagnosis of lower-tropospheric frontogenesis indicates that the flow over western Japan is frontogenetic before Tokage approaches, and the strong cyclonic flow associated with Tokage intensifies the frontogenesis in the northeast quadrant (*i.e.*, front) of the cyclone. The vertical circulation associated with the frontogenesis and the related low-level convergence along the front is considered to intensify the precipitation, although it appears to be inconsistent with the precipitation in the northwest (left-rear) quadrant of the eastward-moving TC.

Tokage is associated with much precipitation on the left side of the storm track, including along the northern coast of western Japan, whereas a TC is often associated with much precipitation on the right side of its track, along the southern coasts of the Japan Islands. The magnitude of the vertical shear that results from ridging downstream of the TC, rather than the midlatitude trough indicated by Atallah *et al.* (2007), may play an important role in the redistribution of the precipitation. The precipitation on the left side of Tokage is generated on the cool side of the low-level front. The southerly flow is the primary contributor to the moisture flux convergence east of the cyclone center, whereas the cool northeasterly flow contributes to the moisture convergence northwest of the cyclone center. The diagnosis on the basis

of the saturated equivalent potential vorticity and the absolute momentum indicated conditional symmetric instability. Thus, the precipitation associated with Tokage is considered to be generated in a less stable cool airmass over a warm sea surface in the southern Sea of Japan. However, precipitation on the left side of the storm track often occurs in association with TCs that make landfall in the North American Continent (*e.g.*, Atallah *et al.*, 2007), where the cool airmass on the left side of the storm track is not under influence of a warm sea surface. Therefore, the precipitation on the left side of the storm track is considered to be primarily related to the environmental flow in the midlatitude, including jet streaks and troughs, which may associated with the (conditional) symmetric instability.

The above discussion of the lower-tropospheric stability associated with the storm implies a contribution of heat transfer from the ocean to the air. Tokage approaches western Japan where the SST in the adjacent sea is cooler than that in the tropical ocean, while the temperature in the lower troposphere decreases, and consequently the thermal contrast near the TC increases at the ET stage, attaining its maximum when Tokage makes landfall in western Japan (Fig. 14) as an intense typhoon. This increase in thermal contrast and the related heat fluxes from the sea surface to the air may contribute to the intensity of Tokage at the time of the landfall. Subsequently, Tokage weakens rapidly and it appears to be decoupled from the SST after the completion of the ET. These seem to be consistent with those of Thorncroft and Jones (2000). However, Kitabatake (2008) showed that the warm core of TY Higos (0221) was thermally decoupled from the underlying sea surface by a cool, stable layer in the lower troposphere *before* the completion of the ET, although it moved into higher latitudes as an intense storm associated with the mid-tropospheric warm core. This is in contrast to TY Tokage, which remains thermally coupled with the underlying sea surface until the completion of ET. The relation between the air–sea thermal contrast and the intensity change of TCs at the ET stage is left for a future study.

Recently, the intensity of a TC over a tropical sea surface is often discussed in relation to the ocean heat content rather than the SST (*e.g.*, Wada and Usui, 2007). Although considerable efforts have been made to study the air–sea interaction of both tropical and extratropical cyclones, the implications for ET have received little attention (Jones *et al.*, 2003). Research is needed to clarify the impacts of the ocean, including the SST, the ocean heat content, and their horizontal gradients, on the intensity and the structure of a storm.

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台風 0423 号 (Tokage) の温帯低気圧化とそれに伴って進路左側に生じた西日本の大雨

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台風 0423 号 (アジア名 Tokage) は 2004 年 10 月に強い勢力で西日本に上陸し、東日本で温帯低気圧化した。この台風は進路左側を含む大雨と強風が特徴であった。この台風について気象庁領域客観解析 (RANAL) を用いて診断を行った。台風が西日本に上陸した際には台風は上層ジェットストリークの入り口及びトラフ下流に位置していたが、トラフは比較的弱いもので、温帯化後に再発達がなかったことに整合的であった。台風の強い下層低気圧性循環が西日本に以前から停滞していた下層前線を強化し、最終的に台風はその傾圧帯で前線性低気圧に変わった。台風が上部対流圏ジェット気流に接近したことによる条件付対称不安定と、前線帯の寒気側かつ日本海南部の比較的暖かい海面上の下層条件付不安定が、台風の進路左側である西日本日本海側の降水に寄与したと考えられる。台風の中緯度での勢力や構造に対する大気海洋相互作用の影響についてさらに研究する必要がある。