Response of Tropical Cyclone Tracks to Sea Surface Temperature in the Western North Pacific

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ABSTRACT

A set of short-term experiments using a regional atmospheric model (RAM) were carried out to investigate the response of tropical cyclone (TC) tracks to sea surface temperature (SST) in the western North Pacific. For 10 selected TC cases occurring during 2002–07, a warm and a cold run are performed with 2 and −2 K added to the SSTs uniformly over the model domain, respectively. The cases can be classified into three groups in terms of recurvature: recurved tracks in the warm and cold runs, a recurved track in the warm run and a nonrecurved track in the cold run, and nonrecurved tracks in both runs. Commonly the warm run produced northward movement of the TC faster than the cold run. The rapid northward migration can be mainly explained by the result that cyclonic circulation to the west of the TC is found in the steering flow in the warm run and it is not in the cold run. The beta effect is also activated under the warm SST environment. For the typical TC cases, a linear baroclinic model experiment is performed to examine how the cyclonic circulation is intensified in the warm run. The stationary linear response to diabatic heating obtained from the RAM experiment reveals that the intensified TC by the warm SST excites the cyclonic circulation in the lower troposphere to the west of the forcing position. The vorticity and thermodynamic equation analysis shows the detailed mechanism. The time scale of the linear response and the teleconnection are also discussed.

1. Introduction

The western North Pacific sees tropical cyclones (TCs) more frequently than any other ocean basin (Yumoto and Matsuura 2001). The TCs there are generated inside of the Asian monsoon trough from boreal summer to autumn and move northwestward along the western fringe of the Bonin high. The TC is primarily advected by background flow at a steering level ranging from 500 to 700 hPa (Holland 1983; Wu et al. 2005) or at the level where potential vorticity is at maximum (Hoover and Morgan 2006). When the Bonin high retreats eastward from its climatological location, TCs tend to undergo recurvature and move northeastward (Harr and Elsberry 1995; Nakazawa and Rajendran 2007). In other cases, the TC moves mostly toward the Chinese mainland directly. Hence, TC tracks in the western North Pacific are roughly classified into recurved and nonrecurved paths (e.g., Hirata and Kawamura 2014): nonrecurved TCs pose a risk to the Chinese coast, while recurved TCs often hit the Japanese islands. This classification is strongly linked with El Niño–Southern Oscillation (ENSO; Wang and Chan 2002).

The effect of global warming on TC tracks in the western North Pacific has been examined mostly by simulation with a general circulation model (GCM). Because TC tracks are attributable to steering flow, global warming may affect them through its effect on background flow. Though beta drift is typically secondary in environments where the steering flow is stronger than 5 m s⁻¹, global warming possibly changes beta drift because it can be affected by the strength of TCs and the environmental flow (Wu and Wang 2004). Changes in the tropical environment may also shift where TCs are.
generated. Wu and Wang (2004) indicated that TCs are expected to be recurved by changes in steering flow. Murakami et al. (2012) emphasized a change in genesis region with a single model analysis. In contrast, Yokoi et al. (2013) concluded that TC tracks will tend to recurve due to a southward shift in the jet stream, comparing several GCM projections to show this. Colbert et al. (2015) offered support for the result of Yokoi et al. by employing a Lagrangian tracking model that considers advection due to steering flow and beta drift. Although the reasons given differ, previous publications consistently conclude that global warming increases the chance of recurved tracks in the western North Pacific, where TCs have been observed to recurve due to a southward shift in the jet stream, comparing several GCM projections to show this. Colbert et al. (2015) offered support for the result of Yokoi et al. by employing a Lagrangian tracking model that considers advection due to steering flow and beta drift.

Recently, some researchers have explored whether TCs play an active role in background flow. Kawamura and Ogasawara (2006) and Yamada and Kawamura (2007) suggested that the diabatic heating generated by TCs serves as a Rossby wave source and intensifies the Pacific–Japan pattern (Nitta 1987; Kosaka and Nakamura 2006). Recently, Hirata and Kawamura (2014) showed that an anomalous anticyclone induced by a typhoon tends to shift the typhoon itself westward or northward (see Shibata et al. 2010). Kudo et al. (2014) moreover demonstrated the importance of typhoon-induced flow for moisture supply in the rainy season. Considering the importance for TC intensity, it is tempting to hypothesize that the TC track is attributable to steering flow not only in direct response to warmer SSTs but also in response to diabatic heating in TCs, with intensity perhaps increasing due to the warmer SSTs. However, it is quite difficult to distinguish between these by data analysis or GCM experiments, and so the possibility that the TC track is shifted due to the TC itself via its effect on steering flow in a warm climate has not been resolved yet.

The purpose of this study is to examine the hypothesis that stronger TCs in a warmer SST environment will change the steering flow such that TC tracks will tend to either shift eastward or recurve with a higher chance. Short-term experiments by Bond et al. (2010) and Ren et al. (2014) using a regional atmospheric model (RAM) hint that a direct response by the steering flow to the warmer SSTs should be ruled out because the lateral boundary provides the position of the midlatitude jet stream. Though Ren et al. (2014) also showed an eastern shift of the track in a single case for the western North Pacific typhoon, the lateral boundary is too constrained because the model domain is narrow compared with the size of typhoons. Bond et al. (2010) provided an interesting example, an eastward shift of a TC when the model settings used a sufficiently large domain, but they did not focus closely on the problem that we consider in this paper. For our purposes, we use a large domain in an RAM to mitigate the effect of lateral boundaries and choose five recurved cases and five nonrecurved cases to avoid subjective sampling as much as possible. Moreover, a linear baroclinic model (LBM) is used to support our hypothesis together with hindcast RAM simulations. The stationary linear response to diabatic heating produced by a TC can discriminate the effect of TC heating from the direct effect of SST.

The remainder of this paper is organized as follows. Section 2 describes the data used in this study and the experimental settings used for simulation with the RAM and LBM. Section 3 shows the TC track response to SSTs in the RAM and describes the importance of steering flow in the warm SST environment. Section 4 discusses the response to the diabatic heating induced by the TC to explain how and why the steering flow is reinforced by the intensified TC. Section 5 summarizes the paper.

2. Method

a. Data

From the typhoon best-track data compiled by the Japan Meteorological Agency (JMA), which includes information on the position, size, and strength of typhoons from genesis to lysis over the western North Pacific at a primary sampling interval of 6 h, we selected 10 different typhoon cases occurring during 2002–07 (Table 1). Typhoons Etau, Dianmu, Songda, Nabi, and Man-yi are generated in the equatorial North Pacific, move northward, and cross the Japanese islands after recurvature; Typhoons Sinlaku, Imbudo, Haitang, Longwang, and Saomai are generated in the equatorial North Pacific as well, but move westward, and land at the Chinese mainland.

We used the 6-h Japanese 55-year Reanalysis Project (JRA-55; Kobayashi et al. 2015) with a grid interval of 1.25°. The data are input as the initial, lateral boundary, and bottom boundary conditions of the RAM simulation (section 2b) and as the background flow in the LBM (section 2c). The SST data are also provided from JRA-55 (see Fig. 1 for the SST at 0000 UTC 1 September 2004 for the control run of the Songda experiment).

b. RAM experiments

The RAM used in this study is the JMA/Meteorological Research Institute nonhydrostatic model. The prognostic variables of the RAM are three-dimensional wind vector; potential temperature; and density of dry air, water vapor, cloud water, cloud ice, snow, and graupel. The model also includes state-of-the-art physical parameterizations such as cloud microphysics (Ikawa and Saito 1991), atmospheric radiative transfer (Sugi et al. 1990), turbulent mixing (Kumagai and Saito 2004), boundary layer processes (Sun and Chang 1986), and surface flux estimations (Sommeria 1976; Louis et al. 1982). Convective precipitation is compensated by the Kain–Fritsch scheme (Kain and Fritsch 1993). See Saito et al. (2006) for the details of this model.

RAM integration is performed for every TC period (Table 1) with a horizontal mesh size of 15 km in the Lambert conformal projection and 40 vertical levels with terrain-following coordinates. The model domain roughly covers the area from 5° to 55°N and from 90° to 170°E, with a buffer zone of approximately 400 km (Fig. 1), which is applied to all typhoon cases (Table 1). The integration period is the lifetime of each typhoon from genesis to lysis, following the record of the observed best-track data (Table 1).

A set of three runs was carried out for each typhoon experiment: the control run (hindcast simulation), a warm run with 2-K uniformly added to SSTs over the model domain, and a cold run with 2-K uniformly subtracted.

c. LBM experiments

The stationary linear response to diabatic heating is solved by using the LBM (Watanabe and Kimoto 2000). This model is based on a global, primitive equation model linearized around the basic state. The basic state, which comprises zonal and meridional wind, temperature, and surface pressure, is defined as the 31-day time average around the date for which the forcing is given from the JRA-55 data. Spectral T42 resolution is used in the horizontal directions, and there are 20 vertical levels, using the sigma coordinate. The LBM includes Rayleigh damping and Newtonian cooling. Because the response mostly achieves a steady state by 8 days (see section 4b), the response at 10 days is regarded as the stationary response in this paper. The forcing is given as the diabatic heating rate simulated in the warm or cold run at the timing around when the TC center in the warm run diverges from that in the cold run. Diabatic heating is evaluated as the potential temperature tendency excluding radiation obtained from the RAM simulation, which is primarily condensation heat release.

For example, the warm run of the Songda experiment shows condensational heating in the midtroposphere at
0000 UTC 1 September 2004, when the typhoon center is located at 20.4°N, 146.4°E, between Guam and the Ogasawara Islands (Fig. 2a). A spiral feature around the typhoon can be identified, even in a 15-km-resolution simulation. We take the spatial average for diabatic heating rate as in Fig. 2a and provide it at only the typhoon central grid cell and eight surrounding grid cells of the LBM as the forcing. In this example, the diabatic heating rate rises to 900 K day⁻¹ in the wall-cloud region in the RAM, but its maximum is reduced to 100 K day⁻¹ in the LBM as a consequence of spatial smoothing. It is remarked that the estimate of the diabatic heating rate would be more accurate in a model with finer resolution because the 15-km-resolution RAM simulation misses some mesoscale features in observed typhoons, such as clear polygonal eyewalls with intense updraft cores (Gentry and Lackmann 2010), and the deepening is unclear (see the appendix). It is remarked that, in contrast with the warm run, the TC in the cold run is much weaker (Fig. 2c) and the prescribed diabatic heating is one order smaller (Fig. 2d).

d. Steering flow and Lagrangian tracking

The steering flow at a particular point is here defined as the horizontal wind vertically averaged from the surface to 200 hPa with mass weighting after taking the horizontal average inside of a circle with a 500-km radius from the point. The steering flow is, therefore, computed for any place and for any time (Ho et al. 2004; Yokoi and Takayabu 2013; Wu et al. 2005) and here it is also used to evaluate the environmental flow surrounding the TC. We additionally include Lagrangian tracking of advection by steering flow \( \mathbf{V}_s \) at the TC center and beta drift

\[
\mathbf{V}_b = \left| \mathbf{V}_b \right| \begin{pmatrix} \sin \theta_b \\ \cos \theta_b \end{pmatrix},
\]

from the date when the LBM forcing is assigned. Following Smith (1993), the effect of beta drift is calculated as

\[
\left| \mathbf{V}_b \right| = 0.9 r_m \sqrt{\frac{\beta V_m}{\left| \mathbf{V}_s \right|}} \quad \text{and} \quad \theta_b = 308° - 9.6° \log_{10} \left( \frac{\beta r_m^2}{V_m} \right),
\]

where \( \beta \) is the rate of change of the Coriolis parameter, \( V_m \) is the maximum value of tangential velocity in the TC, and \( r_m \) is the radius at which \( V_m \) occurs. The steering flow and the beta drift are provided with a 3-h time resolution, as they are recorded as an output of the RAM. Therefore, the Lagrangian tracking can predict the typhoon position \( \mathbf{X} \) as

\[
\mathbf{X}(t + \Delta t) = \mathbf{X}(t) + \left[ \mathbf{V}_s(t) + \mathbf{V}_b(t) \right] \Delta t,
\]

where \( \Delta t \) is 3 h.

3. Results

a. TC response to SSTs and its classification

As shown in the appendix, the control run mostly reproduces the observed typhoon track for the 10 cases...
that we selected. We then regard it as the reference for warm and cold runs. Whether the observed track was recurved or not, the result of a set of warm and cold runs can be logically classified into one of four types in terms of recurvature. It is quite interesting that no typhoon experiments here find a nonrecurved track in the warm run and a recurved track in the corresponding cold run. The three other groups are discussed below.

Dianmu, Nabi, and Man-yi experiments show a recurved track in both warm and cold runs. Typhoon Dianmu starts at 9.3°N, 136.4°E, west of Guam, and mostly moves northward (Fig. 3a). The warm run’s track is approximately 300 km east of the cold run’s track. Because the movement speed is higher in the warm run, the TC lands at the Japanese islands in 6 days for the warm run but in 8 days for the cold run. The northward migration speed in the control run is between these two times. At 5 days, sea level pressure (SLP) in the TC center reaches 967 hPa in the cold run and 940 hPa in the warm run. Typhoon Nabi is generated at 15.0°N, 152.3°E, east of Guam, and takes a typical recurved course (Fig. 3b). The warm run is recurved around the Ogasawara Islands and moves faster toward the north. In contrast, the cold run’s TC moves northwestward for 8 days and then lands on the Korea Peninsula. This case is now categorized into recurvature for both runs, although the recurvature in the cold run was somewhat weak. The warm run rapidly develops the TC in 4 days with 930 hPa at the TC center, while the cold run prevents the TC from deepening below 960 hPa. The control run shows a TC track positioned between warm and cold runs. Typhoon Man-yi is initiated at 10.3°N, 142.3°E, south of Guam (Fig. 3c). In this case, both runs take a similar course with recurvature, but the warm run reaches the Japanese mainland 24 h faster than the cold run does. In the warm run, growth is much more rapid, and the SLP in the TC center attains its minimum within 3 days. In this group all cases in the warm run show the TC moving northward faster and with more deepening, resulting in a track located more or less eastward, which is consistent with results from previous studies (Bond et al. 2010; Ren et al. 2014).
Sinlaku, Songda, Saomai, and Etau are categorized as having a recurved track in the warm run and a non-recurved track in the cold run. Sinlaku is initially at \( 18.7^\circ N, 155.1^\circ E \) and moves directly westward in the observation (not shown). The warm run stays near the initial position for 6 days with slow development and moves northward after maturing (Fig. 4a). The cold run moves slowly westward, which is different from the warm run. The cold run does not develop the TC in 6 days, and its integration is ended before landing occurs. In contrast with Sinlaku, Songda shows recurvature

**Fig. 3.** Cyclone tracks and SLP (hPa; inset) at the cyclone center over the typhoon period simulated by the RAM for Typhoons (a) Dianmu, (b) Nabi, and (c) Man-yi. The thick line with open circles, closed circles, and open squares plotted every 24 h denote the warm, cold, and CTR runs, respectively.

**Fig. 4.** As in Fig. 3, but for Typhoons (a) Sinlaku, (b) Songda, and (c) Saomai.
as in the observation (Figs. A1a and 4b). Starting at 11.3°N, 165.0°E, unfortunately in a model buffer zone, the warm run’s TC moves apart from the cold run’s TC at 3 days. The warm run reaches 937 hPa before landing on the Japanese main islands. Considerably different, the cold run’s TC moves straight westward at an almost constant pace. The Saomai experiment gives a more modest result. Starting at 11.7°N, 146.5°E, the TC moves northwestward with ample deepening in the warm run and moves westward with insufficient deepening in the cold run (Fig. 4c). Saomai is thus categorized into the same group as Sinlaku and Songda, but the recurvature in the warm run is not clear, despite the TC landing on the Japanese islands. Typhoon Etau, too, may be categorized into this group (not shown). However, it shows slow growth in all runs, and so we exclude the Etau experiments from further analysis. Despite differences among the cases, warm SSTs are apt to cause the TC to track northward and undergo faster development.

The three remaining cases, Imbudo, Haitang, and Longwang, show a nonrecurved track for both warm and cold runs. From the initial position of 8.3°N, 140.9°E, the Imbudo’s TC goes northwestward and lands at the Chinese mainland (Fig. 5a). The TC track in the warm run is several hundreds of kilometers north of the track in the cold run. At 5 days, the TC passes Taiwan with its central SLP of 944 hPa in the warm run, while it passes Luzon, Philippines, with a central SLP of 979 hPa in the cold run. The Haitang case shows a very similar track between the warm and cold runs (Fig. 5b). The only clear difference for that pair of runs is in deepening of the TC. Different from the above examples, in the Longwang experiment, the TC does not land at the Chinese mainland in either run (Fig. 5c). The warm run shows westward movement of the TC with relatively slow development, but its integration ends when the TC is located south of Hainan. The control run moves westward more slowly. The cold run is degraded much more than the warm run, especially given the looping track in the east of the Philippines. In spite of various features in this nonrecurved group, a common result is further deepening of the TC under the warmer SST condition. In two cases of three, the warm ocean encourages the TC track to be more northward.

b. Steering flow and beta drift changes

In this subsection, we will explain why the TCs tend to move northward in the warmer SST environment. Among 10 cases, we choose a single typical case from each group described in section 3a, in which the warm run has greater northward movement than the cold run: Nabi from both the recurved groups (Fig. 3b), Songda from the group of recurved in warm and nonrecurved in cold (Fig. 4b), and Imbudo from both the nonrecurved groups (Fig. 5a).

Figures 6a and 6b show the steering flow of the warm and cold runs for the Nabi experiments at 1200 UTC 3 September 2005, when the TC center is located near the Ogasawara Islands and the warm run’s TC begins to separate from the cold run’s TC. Commonly, cyclonic
circulation is dominant around the TC and the mid-latitude jet stream meanders in northeast Asia. A noticeable difference between the runs can be seen in Southeast Asia. The monsoonal trough is enhanced in the warm run, which has the effect of increasing TC advection toward the north. Lagrangian tracking from the TC position at that date, in fact, indicates a northeastward track caused by the steering flow in the warm run and a northwestward track caused by the steering flow in the cold run. The cold run undergoes recurvature after northwestward translation in the tracking. The results of Lagrangian tracking strongly emphasize the eastward movement of the warm run’s TC, but tracking also captures the feature that both runs experience recurvature and the warm run takes a more eastern track. The steering flow at the TC center and the beta drift are larger in the warm run, but the contribution of the steering flow to the northward migration of the TC is dominant after 4 September (Figs. 6c,d). This indicates that the steering flow causes a faster movement of TC toward the north.

The steering flow in the Songda experiment is displayed in Figs. 7a and 7b. The warm run’s TC begins to leave the cold run’s TC near Guam. A common feature in both runs is cyclonic circulation obviously due to the TC. The midlatitude jet stream over Japan is southwest–northeast oriented in the warm run, while it is west–east oriented in the cold run. The easterly steering flow is prevalent in the subtropics for the cold run, while the monsoonal trough is extended over the latitudinal belt at 15°N in the warm run. The Lagrangian tracking indicates primary advection by the steering flow, which demonstrates that the warm run’s TC moves northward as it is advected by southerly wind (Fig. 7c). The beta drift is actually activated by the warm SST but the contribution to the northward movement of the TC is secondary (Figs. 7c,d). In contrast, the cold run’s TC moves westward as it is advected by a subtropical easterly. The northward component of steering flow is nearly zero (Fig. 7d).

Figures 8a and 8b show the steering flow in the warm and cold runs for the Imbudo experiment at the date.
when the TC is located just east of the Philippines. The steering flow is commonly westerly in the latitudinal belt between 30° and 40°N for both runs. The cyclonic circulation due to the TC is also common to both. Additionally, another small-scale cyclone emerges in the warm run just west of the Philippines. This is another TC that is amplified under the warm SST condition. According to the Lagrangian tracking result (Figs. 8c,d), the steering flow in the warm run pushes the TC northward more strongly than in the cold run. However, the dominance of steering flow advection to beta drift is moderate compared with Nabi and Songda cases (Figs. 6 and 7). It is remarked that the difference in the TC tracks between the two runs can also be caused by the Fujiwara effect, in which two contiguous vortices are rotated counterclockwise about each other (Brand 1970). In the Imbudo case, two vortices are situated on either side of the Philippines in the warm run, while a single TC is seen in the cold run. The Fujiwara effect can explain the warm-run track to the north relative to the cold-run track, perhaps because the effect strongly works in the warm run. In contrast, the Longwang TC (Fig. 5c) has a counterpart TC in its east. This effect is enhanced under the warm condition (not shown). In this case, the Fujiwara effect tends to discourage the TC from moving northward. In fact, in the warm run, the track holds in a relatively southern position in comparison with other cases.

c. Dependency on the domain size and the start time

Now we pay attention to the caution that the RAM experiments are sensitive to the model domain and the start time. This problem is discussed in the context of the Songda experiment, where the warm and cold runs showed the greatest difference in the TC track.

An additional experiment for the Songda case is tried with a narrower domain in which the western boundary is shifted to the east (Fig. 9a). This domain does not include the area covered by the cold TC track of the standard-setting experiment after landing at the Chinese mainland (Fig. 4b). The resultant track for the cold run actually wanders around the western boundary, while the track for the warm run is quite similar to that obtained in the standard-setting experiment. This implies that the RAM experiment should be performed with a sufficiently large domain to cover movement across the TC lifetime from genesis to lysis. Except for the starting point of Songda, our experiment has a sufficiently large domain for the TC track in both the warm and cold runs.

Since the initial TC is taken from the reanalysis data, the dependency on start time may be more
crucial. If the Songda experiment started from 0000 UTC 1 September 2004, four days later than the standard setting, then TC development would be delayed. The resultant track for the warm run (Fig. 9b) is quite different from the track that was shown in section 3a (Fig. 5b), while the cold run exhibits TC recurvature. Looking at the steering flow in the warm run at four integration days, when the TC was located at 23.3°N, 127.8°E (Fig. 9c), northward migration is not encouraged. With the standard settings for our experiments, integration is performed for the period of the typhoon on the basis of the best-track data. If the start time were set to later, as in this example, the result would be quite different, and would not be appropriate in most cases. Similarly, if the start time were earlier, the warm run might shift its TC track northward earlier than in the standard-setting experiment.

4. Linear response mechanism and teleconnection

a. Linear response and its mechanism

In section 3, the results from most of the RAM experiments support the hypothesis that the intensified TC in the warm SST is primarily associated with the intensified steering flow, causing the TC to tend to move northward. However, in contrast with an understandable matter that the TC size and strength affect the beta drift, the causality for the excitation of cyclonic circulation west of the TC by the intensified TC has not been established yet because the SST change could possibly excite the cyclonic circulation directly. The intensified TC offers the strong TC vortex circulation and great diabatic heating in the midtroposphere, and pressure deepening at the surface. These could be a point forcing agent in the global LBM, but hereafter we compute the linear response to diabatic heating obtained from the RAM output (section 2c). This is because, as will be discussed in section 4b, diabatic heating produces the cyclone vortex as the local response and vice versa. This means that either diabatic heating or vortex forcing is enough to think of a nonlocal response of the TC forcing.

The stationary linear response to the diabatic heating at the date when the warm run’s TC begins to separate from the cold run’s one is displayed in Fig. 10. The linear response for the Nabi’s warm run induces the cyclonic circulation that encompasses the East China Sea (Fig. 10a). An accompanying anticyclone is excited to the east of Japan. The cyclonic
pattern in the LBM response is obviously similar to the steering flow in the warm run (Fig. 6a). The anticyclonic pattern seems to play a role in expanding the jet stream to the south (Fig. 10a). In contrast, little signals in steering flow can be found for the cold run with much weaker diabatic heating (Fig. 10b), consistent with the RAM results (Fig. 6b). Although the forcing is imposed in an easterly background flow, the Songda case shows a similar response to the Nabi’s case (Figs. 10c,d). For the warm run, the LBM response also exhibits a feature of cyclonic circulation to the west of the TC, consistent with the steering flow in the RAM (Fig. 7a). In contrast, such a signal is not found in neither the LBM response (Fig. 10d) nor the RAM’s steering flow (Fig. 7b) for the cold run. These results clarify that the Nabi and Songda TCs under the warm SST run are pulled to the north by the steering flow response to greater diabatic heating by themselves.

We also show the stationary linear response for the Imbudo case, in which the TC center is located closer to the equator. The LBM response for the warm run exhibits a cyclonic circulation in terms of steering flow in the northwest of the diabatic heating, and the resultant flow is only easterly at the TC position. This certainly captures a large-scale feature of the steering flow created by two vortices in the RAM (Fig. 8). Different from the other cases, therefore, the large-scale response to diabatic heating as obtained from the warm run of the Imbudo experiment does not much contribute to differences between the warm and cold runs.

The mechanism of the stationary linear response is basically the subtropical thermal response proposed in Hoskins and Karoly (1981). Looking at the cross section at 20°N for the LBM response in the warm run of Songda case (Fig. 11a), the upward motion is found at the TC forcing place, and an anticyclone in the upper troposphere and a cyclone in the lower troposphere reside in the west of the forcing. Because the steering flow response is defined as the mass-weighted average of wind, it almost corresponds to the lower-tropospheric response. We then think of each term for the vertical-averaged thermodynamic equation linearized around the background flow for the lower troposphere, approximately given as

\[ \bar{u} \frac{\partial \bar{\theta}}{\partial x} + \bar{v} \frac{\partial \bar{\theta}}{\partial y} + \bar{\omega} \frac{\partial \bar{\theta}}{\partial p} = \bar{Q}, \]

where the overbar and prime denote background flow and perturbation, respectively; \((x, y)\) is the horizontal coordinate; \(p\) is pressure; \((u, v, \omega)\) are horizontal and vertical velocity; \(\theta\) is potential temperature; and \(\bar{Q}\) is diabatic heating rate in terms of the thermodynamic equation. The first, second, and third terms on the left-hand side of Eq. (4) represent the thermal advection
by zonal background flow, thermal advection by perturbation southerly, and adiabatic heating by perturbed upward motion, respectively. Adiabatic heating term is dominantly balanced with the diabatic heating (Fig. 11b), which explains the upward motion at the center of forcing. Next think of the vertically averaged vorticity equation linearized around the background flow for the lower troposphere approximately given as

\[
\frac{\partial \zeta'}{\partial x} + \beta v' = f \frac{\partial \omega'}{\partial p} \tag{5}
\]

where \( \zeta' \) is relative vorticity and \( f \) is Coriolis parameter. The first and second terms denote vorticity advection by background zonal wind and beta effect, respectively; the right-hand side is vortex-tube stretching. The upward motion in the midtroposphere induces the positive vertical derivative of vertical wind, say vortex-tube.
stretching, in the lower troposphere. This balances both terms on the left-hand side (Fig. 11c). The beta effect term produces the southerly in the center of forcing and the zonal advection term produces the low pressure in the west of forcing under the easterly environment (Fig. 10c). Both terms then explain the lower-tropospheric response displayed in Fig. 11a. The Nabi case shows a similar term balance (not shown), except for
the dominance of the beta effect term in Eq. (5). On the other hand, the response in the Imbudo case is slightly different from that in the Songda case. Considering that the forcing is nearer to the equator, it is speculated that the response to diabatic heating might be not purely subtropical but rather equatorial (Gill 1980). This needs to be more analyzed for a complete picture for the response, so we cease to pursue this problem more in this paper because this is not substantial for this paper’s purpose.

b. Time scale for the linear response

The stationary linear response shown above takes several days to build the stationary response in our LBM integration. Here we will briefly discuss how many days it actually takes for the response to build. As an example, we investigated the stationary response to the forcing fixed at a particular date and found consistency between the RAM’s steering flow at the date and a linear response to the forcing fixed at the chosen date. However, the steering flow that advects the TC at this date must be formed by this date, being responsible for the moving diabatic heating over several days from the start date. To consider the response rigorously, the transient linear response to moving diabatic heating prior to the examined date should be solved. Here, because a nonautonomous linear response has not been implemented in the LBM system that we used, the transient response to the forcing fixed at the central date of the period from the start time is substituted for the transient response to the forcing as it progresses during the period.

We use the Songda case as an example again. Figure 12 shows the 4-, 8-, and 12-day responses to diabatic heating at 0000 UTC 30 August 2004 for Songda. The 4-day response (Fig. 12a) is much weaker than the 8- and 12-day responses (Figs. 12b,c). The 8- and 12-day responses are regarded as the stationary response, which is consistent with the RAM’s steering flow under the warm SST environment (as concluded in section 3). In contrast, the 4-day response is actually an approximate solution for the response to diabatic heating from 28 August to 1 September in the LBM system. However, different from nature, zero perturbation is given for all climatic variables as the initial conditions of the LBM system. It takes about 5 days for the TC vortex to build as a local response to the prescribed diabatic heating. This may be related to the discussion by Hoskins and Rodwell (1995), in which the linear primitive equation model needs 5 days of warm up for the hydrostatic adjustment to orographic lift. Linked with our LBM experiments, it takes 5 days to build the TC vortex and low pressure in the center of forcing as the initial local response to diabatic heating (not shown). After that, it takes a couple of days to excite a cyclonic circulation to the west of TC toward the stationary response (Fig. 12b). Therefore, assuming that the movement of the TC in a couple of days is smaller than the size of TC, the stationary response to fixed diabatic heating can be substituted for the transient response to moving diabatic heating.
c. Teleconnection

As mentioned in the introduction, Yamada and Kawamura (2007) stressed remote effects of the TC-induced wave. We here discuss a remote effect of the TC-induced waves beyond a local baroclinic response near the TC shown in section 4a. The teleconnection pattern is excited from the local response near the TC. For example, the response to diabatic heating for the typhoon Songda on 1 September 2004 includes a cyclone anomaly to the north of the TC prevailing over the Sea of Okhotsk and an anticyclone anomaly to its east (Fig. 13a). The refractive index $K_r = \cos \phi \sqrt{B/\bar{u}}$ (Hoskins and Ambrizzi 1993), which is prescribed for the LBM diagnosis, indicates that forcing is imposed in the northern edge of the region, where Rossby waves of any wavenumber are prohibited. In this, $\phi$ is latitude and

$$B = \beta a^2 - \frac{1}{a} \frac{\partial}{\partial \phi} \left[ \frac{1}{\cos \phi} \frac{\partial}{\partial \phi} (\pi \cos \phi) \right],$$

(6)

where $a$ is Earth’s radius. In contrast, the pattern from the Sea of Okhotsk to the Aleutian Islands exists in a waveguide where planetary-scale waves can propagate. The horizontal component of wave activity flux (Takaya and Nakamura 2001),

$$W = \frac{1}{2 \sqrt{\bar{u}^2 + \bar{v}^2}} \left( \bar{u} \left[ \frac{\partial \psi}{\partial x} \right]^2 - \psi \frac{\partial^2 \psi}{\partial x^2} \right) + \bar{v} \left[ \frac{\partial \psi}{\partial y} \frac{\partial \psi}{\partial x} - \psi \frac{\partial^2 \psi}{\partial x \partial y} \right],$$

(7)

demonstrates northeastward wave propagation from the locally induced anticyclone around the TC. Here $\psi$ is an anomalous streamfunction. In contrast with this, there are few waves excited in the extratropics in the Imbudo case, probably because the forcing is deeply inside of the wave-prohibited region (Fig. 13b). However, even in the Songda case, when forcing is imposed in the premature stage of the typhoon (on 28 August 2004), the response is limited to the area around the forcing (Fig. 13c). This is related to the basic state, which prevents the waves from propagating northward. Hence, the remote effect of the TC-induced waves depends on whether TC forcing is located near the area where the wave propagation is permitted. We do not discuss this further here because nonautonomous response to subtropical diabatic heating should be solved as part of a rigorous discussion.

5. Conclusions

We investigated the response of TC tracks to SSTs that are uniformly warmed (or cooled) over the western North Pacific. First, a set of RAM hindcast experiments with warm or cold SSTs imposed as the surface boundary condition were performed for five recurved and five nonrecurved typhoon cases occurring during 2002–07. In seven of these cases, the TC in the warm run tended to move northward rapidly, regardless of whether the recurvature occurred or not. A large difference was observed in two cases, where the TC underwent recurvature in the warm run but moved westward to the Chinese mainland in the cold run. We classified the results for all cases into three groups: a recurved track in both warm and cold runs, a recurved track in the warm run and a nonrecurved track in the cold run, and a nonrecurved track in both runs. Steering flow analysis together with Lagrangian tracking revealed that, for the first two groups, the TC under warmer SSTs is advected northward by an anomalous cyclone to the west of the TC and an anomalous anticyclone to the east. The midlatitude jet stream is moderately intensified east of Japan, which also contributes to the TC track being farther east in the warm run. The stationary linear response to the TC heating, as obtained in the RAM with the warm SSTs, also shows a southerly steering flow at the TC center, flanked by a cyclone in the west and an anticyclone in the northeast, which is consistent with the steering flow. This response is just explained by subtropical thermal response proposed in Hoskins and Karoly (1981), supported by vorticity and thermodynamic equation analysis. The teleconnection pattern is also found as a result of wave propagation from the TC center, when the forcing resides near the wave-permissive area. We summarize the process in the illustration (Fig. 14) as follows. The TC develops more rapidly when moving westward over warmer ocean. In the recurved case, stronger intensification of the TC produces more diabatic heating, which then acts as a forcing on the planetary-scale atmosphere. The TC tends to move northward rapidly after recurvature, being advected by the anomalous steering flow. This can be
FIG. 13. Geopotential height at 300 hPa (m; contours with the TC position marked by an empty circle, an interval of 20 m, negative contours dashed, and zero contours omitted) for the stationary linear response to diabatic heating for (a) Songda at 0000 UTC 1 Sep 2004, (b) Imbudo at 0000 UTC 20 Jul 2003, and (c) Songda at 0000 UTC 28 Aug 2004. The stationary wavenumber $K_s$ (shading) is calculated from the basic-state zonal wind at the model $\sigma$ level 0.295 (~300 hPa). White, light gray, and dark gray denote areas with $K_s < 3$, $3 \leq K_s < 10$, and $10 \leq K_s$, respectively. Black indicates areas where $K_s$ is an imaginary number. The wave activity flux (vectors) is given for resulting stationary waves, following Takaya and Nakamura (2001). The unit vector shows 100 m$^2$ s$^{-2}$; areas where the length of the vector is less than 10 m$^2$ s$^{-2}$ and areas within 1600 km of the TC (surrounded by the white dotted line) are omitted.
called TC intensity–track feedback. In contrast, in the nonrecurved case, the resultant flow blocks westward migration of the TC.

We now discuss the limitations of our study, especially those arising from the use of RAM with a rough horizontal resolution. As briefly mentioned in section 2c, the model resolution set in this study is insufficiently fine to resolve the inner structure of TCs (e.g., Gentry and Lackmann 2010). If a mesh size of a few kilometers were used in an RAM hindcast experiment with warmer SSTs, the TC should develop more rapidly and generate more diabatic heating. This increased diabatic heating would excite a stronger southerly steering flow around the TC center. This implies that the 15-km-resolution RAM biases the timing with which the TC tracks for the warm and cold runs to the north of Japan, and this difference is relatively independent of differences in TC track. For example, the trough ranging from the Far East to the Korea Peninsula is more strongly deepened in the Nabi experiment. The Songda experiment exhibits a pair comprising a cyclone and anticyclone to the north of Japan as part of the difference. This might be interpreted as the effect of land–sea differences or the effect of SSTs. The direct response to the warm SST could be more precisely diagnosed by use of an atmospheric GCM, but this is clearly beyond the scope of this paper.

We have limited our discussion to the TC response in the western North Pacific under the assumption of a uniform increase in SST, which would likely occur with global warming. As discussed briefly in the introduction, a change in North Pacific TCs in response to anthropogenic forcing can be accounted for by a change in various influences, such as tropical SST distribution, and the intensities of ENSO, summertime Asian monsoons, and the upper-level jet stream. It seems to be most important to examine whether the TCs in the western North Pacific are strongly affected by ENSO in terms of frequency, intensity, and tracks (Wang and Chan 2002; Yokoi and Takayabu 2009). However, a change in the zonal gradient of tropical SSTs in the Pacific is quite model dependent (Yeh et al. 2012) and, related to this, the changes in ENSO intensity are uncertain (Solomon and Newman 2011). It is, therefore, natural that TC estimation still depends on physical parameterizations in a GCM (Murakami et al. 2012). Additionally, simulation of summertime Asian monsoons has been recently improved, and results from CMIP5 models suggest that the monsoonal rainfall and circulation is likely to be intensified. Changes that would occur in the upper-level jet stream are also not yet certain, but the CMIP5 model ensemble shows the Northern Hemisphere jet shifting poleward by only 1° (Barnes and Polvani 2013), and the jet stream is expected to be slightly weakened from East Asia to the North Pacific during boreal summer (Lee et al. 2014). Summarizing the discussion here, the problem of changes in TC tracks in response to global warming can be divided into the response to a uniform increase in SSTs (clarified in this paper) and the response to changes in tropical SST distribution because the uncertainty of changes in Asian monsoons and the jet stream is not crucial to predicting future TC tracks in the North Pacific (see Wang and Wu 2012). Hence, the problem of TC tracks with global warming could be solved by clarifying future changes in the tropical SST distribution.

It is interesting to apply our finding on the TC intensity–track feedback to other ocean basins. For
example, a number of hurricanes in the North Atlantic tend to pass along the western fringe of the Azores high, whose climatology is essentially a consequence of the land–sea contrast and the topography (Miyasaka and Nakamura 2005) together with indirect support by tropical monsoonal heating (Rodwell and Hoskins 2001). Because the upper-level westerly flow meanders over North America in boreal summer, quite differently from the North Pacific, the response to TC-scale diabatic heating in the subtropics is expected to be quite different from what we obtained in this paper. We could apply an insightful discussion by Minobe et al. (2010), in which deep convection with an oceanic front scale along the Gulf Stream induces adiabatic ascent. Combining this discussion with the result on the atmospheric response to an Atlantic SST anomaly by Palmer and Sun (1985), a barotropic anticyclone anomaly may be stimulated just east of the anomalous diabatic heating accompanied by TCs. If TC intensity–track feedback were to operate, then a TC in a warmer environment would go farther northward. However, there are few publications on a nonlinear response to diabatic heating under a meandering jet stream, such as the stream over summertime North America. Further investigation is needed before our finding can be applied to other ocean basins.

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![FIG. A1. Cyclone tracks over the typhoon period for TCs (a) Songda and (b) Imbudo. The thick line with the open squares plotted every 24 h, denotes the CTR run; the thin line with crosses denotes the observed track based on the best-track data archived by the Japan Meteorological Agency. SLP (hPa) at the cyclone center for TCs (c) Songda and (d) Imbudo.](image1)

![FIG. A2. Error of the position of typhoon center for the CTR run of each RAM experiment at 5 days from the initial date. The horizontal and vertical axes indicate errors in the east and north directions, respectively.](image2)
Institute’s nonhydrostatic model and JRA-55 data were used with permission of the JMA and with special assistance by Dr. S. Hayashi. The model simulations were performed on the Hokkaido University High Performance Computing System. The LBM was used with the permission of Dr. M. Watanabe. Figures were drawn with Grid Analysis and Display System.

APPENDIX

Performance of the RAM

Here we check the performance of the RAM with a 15-km horizontal mesh size for recurved and non-recurved typhoons. The recurved typhoon, Typhoon Songda, is initially located at 11.3°N, 140.9°E at 0000 UTC 28 August 2004, and then moves almost westward slowly before recurving and crossing the Japan Sea accompanied by strong wind over Japan (Fig. A1a). The error in the position of the typhoon center is only about 300 km at 5 days after genesis, though it expands when Songda passes around the Ryukyu Islands (Fig. A2). The RAM control runs for other recurved cases, except for Etau, well reproduce the observed tracks. The control run for Etau does not show recurvature around the Ryukyu Islands, which may be attributable to its immature status there (not shown). Moreover, the RAM broadly fails to simulate the observed typhoon intensity (Table 1). Looking at the deepening of central SLP in the life cycle of Songda and its simulation by the RAM (Fig. A1c), Songda attained the stage with the strongest deepening pressure, of 925 hPa on 3–5 September 2004, just before recurvature, but the control run shows only a slow development of the TC, with a resultant minimum pressure of 951 hPa at the same date. Both in observation and simulation, the central pressure increases after recurvature, and the typhoon undergoes extratropical transition (Jones et al. 2003). We obtain a similar result from experiments for the other cases (not shown).

The RAM track almost follows the best track for the non-recurved typhoon, Typhoon Imbudo, which was initially located at 8.3°N, 140.9°E at 0600 UTC 17 July 2003, and then moved northwestward before reaching the southern coast of China (Fig. A1b). The error in the typhoon position is only about 400 km to the north for five integration days (Fig. A2). RAM’s control run for other non-recurved cases well reproduces the track, except for Sinlaku (Table 1). The Sinlaku experiment takes an unrealistic, abrupt turn toward the north, though the best track moved smoothly to the west. Though Imbudo and Saomai experiments reproduce the correct intensity (Fig. A1d), the RAM has a bias toward very slow development for the Longwang and Sinlaku cases and also has a bias toward weaker deepening for Haitan (not shown).

REFERENCES


