Spatially Inhomogeneous Trends of Tropical Cyclone Intensity over the Western North Pacific for 1977–2010

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ABSTRACT

The spatial distribution of trends in tropical cyclone (TC) intensity over the western North Pacific Ocean (WNP) during the period 1977–2010 was examined using five TC datasets. The spatial distribution of the TC intensity was expressed by seasonally averaged maximum wind speeds in 5° × 5° horizontal grids. The trends showed a spatial inhomogeneity, with a weakening in the tropical Philippine Sea (TP) and a strengthening in southern Japan and its southeastern ocean (SJ). This distribution could be described by TC intensification rate and genesis frequency, with the aid of the climatological direction of TC movement. The increasing intensification rate around the center of the WNP could mostly account for the increasing intensity over the SJ region, while the influence of both intensification rate and local genesis frequency mattered in the TP region because of the effect of the newly generated and less-developed weak TCs on the TC intensity. Thermodynamic variables (e.g., sea surface temperature, potential intensity, and 26°C isotherm depth) showed almost homogeneous changes in space, possibly favoring intensification rate and genesis frequency over the entire WNP. However, the decreasing intensification rate and genesis frequency in some tropical regions conflicted with the impact of thermodynamic variables; rather, they were in accord with the impact of dynamic variables (i.e., vorticity and wind shear). In conclusion, the spatially inhomogeneous trends in TC intensity could be explained by considering the thermodynamic and dynamic aspects in combination through intensification rate and genesis frequency.

1. Introduction

The way in which tropical cyclone (TC) intensity may change under a changing climate is a hotly debated research topic in the TC research community (Knutson et al. 2010). Many observational and modeling studies have examined possible changes in the strength of TCs over various ocean basins associated with climate change (e.g., Emanuel 2005; Webster et al. 2005; Elsner et al. 2008; Bender et al. 2010; Yu et al. 2010). For the western North Pacific Ocean (WNP) basin, studies based on high-resolution climate model simulations suggested that the relative number of intense TCs, maximum wind speed, and TC-induced rainfall may increase or remain unchanged in a warmer climate, indicating that the results are highly dependent on the model used (Oouchi et al. 2006; Yoshimura et al. 2006; Gualdi et al. 2008; Murakami et al. 2011).

Analysis of observational records also produced conflicting interpretations of historical trends of TC intensity for the WNP basin (Grossmann and Morgan 2011). Some studies argued that the overall TC intensity and the number of intense TCs show significant increasing trends...
with tropical ocean warming (Emanuel 2005; Webster et al. 2005), whereas others raised a counterargument that decadal variations are more dominant than specific trends (Chan 2006, 2008). These incompatible results mainly arise from uncertainty in the available TC intensity records. There is an obvious inconsistency in terms of wind speed among various TC best-track datasets over the WNP (Wu et al. 2006; Kamahori et al. 2006; Song et al. 2010). To enhance the reliability of trends in the TC intensity, efforts have been made to reanalyze or reproduce the TC intensity records in a homogeneous way (Kossin et al. 2007; Wu and Zhao 2012).

According to the previous literature, it is difficult to find consistent long-term changes in intensity-related TC parameters, particularly for basinwide total (or average) statistics, over the WNP. In this study, we adopt an alternative approach to assessing changes in TC intensity; that is, we focused on horizontal distribution, thereby illuminating subregions where the intensity changes are statistically significant and consistent in recent decades. This idea was inspired by Park et al. (2011), who found a consistent increasing trend of TC intensity over mid-latitude East Asian regions (i.e., Korean peninsula and Japanese islands) for the period 1977–2008 with the four available TC best-track datasets over the WNP. This finding was attributed to relevant changes in large-scale environments over the midlatitudes [i.e., sea surface temperature (SST) warming, tropospheric moistening, weakening of vertical wind shear, and strengthening of upward motion]. This study leads us to infer the existence of consistent weakening of TCs somewhere in the WNP under the condition that long-term changes in the intensity-related parameters for the whole basin do not show a significant and appreciable trend during recent decades (Wu et al. 2006; Kossin et al. 2007).

Consistent with our inference, Wu and Zhao (2012) recently demonstrated a spatially inhomogeneous distribution of TC intensity change (see their Fig. 6). However, they did not conduct an in-depth analysis of the horizontal distribution, but related it to the basinwide change of TC intensity. Thus, it is worth further concentrating our attention on the spatial pattern of the intensity changes. The spatial inhomogeneity of the intensity trend naturally implies that dynamic and thermodynamic environmental fields would vary considerably, depending on subregions, over the WNP. Actually, some previous studies may support this expectation (Archer and Caldeira 2008; Sohn and Park 2010), and this paper attempts a more thorough analysis by focusing on TC-related large-scale parameters.

The rest of this study is organized as follows. Section 2 illustrates the data and methods of analysis used. Section 3 shows spatial distributions of changes in TC intensity and other activity parameters (e.g., intensification rate, genesis, and passage) and then discusses the relevant changes in atmospheric and oceanic variables. Finally, concluding remarks are given in section 4.

### 2. Data and method

Four TC best-track datasets were used to examine the consistency of changes in TC intensity over the WNP. The best-track datasets include those of the Regional Specialized Meteorological Center (RSMC) Tokyo, Joint Typhoon Warning Center (JTWC), Hong Kong Observatory (HKO), and Shanghai Typhoon Institute (STI). Notable distinctions in TC intensities were reported among the four datasets (Knapp and Kruk 2010). Prior to the main analyses, the units and averaged time intervals for calculating maximum sustained wind speed were unified into meters per second and 10 min, respectively, using multiplicative factors [see Table 1 in Park et al. (2011)]. This does not change the properties of individual best-track data, but it enables us to compare intensity records of the four best-track data on approximately equal terms. The analyses of TC records were confined to the period 1977–2010 because records for maximum sustained winds are only available after 1977 in RSMC (Kamahori et al. 2006; Park et al. 2011).

To evaluate the trends found in the best-track data, the TC dataset of University of Wisconsin-Madison/National Climatic Data Center (UW/NCDC), introduced by Kossin et al. (2007), was also utilized. The data are reproduced by a simple multivariate log-linear regression model based on the new globally consistent satellite data (Knapp and Kossin 2007), which covers the period 1982–2006. Because the UW/NCDC data were trained for the North Atlantic, they may contain a substantial bias when applied in intensity analyses for other basins (e.g., in Fig. 2a, described in greater detail below, absolute values of the UW/NCDC data are considerably lower than those of the best-track data.) Thus, we excluded the UW/NCDC data when calculating averages of all datasets. Nevertheless, the UW/NCDC dataset is a good estimator for evaluating whether the trends based on the best-track data are reliable, because it is regarded as containing the most homogeneous data at present (Kossin et al. 2007). Although the intensity estimates of the UW/NCDC data for the WNP TCs are not complete, the consistent trends between the best-track and UW/NCDC data indicate that the trends are less likely to be artificial because of the heterogeneity of the best-track records (J. P. Kossin 2012, personal communication). The TC season was defined as 5 months, from July to November, when the formation of intense TCs (i.e., Saffir–Simpson categories 4 and 5) is more frequent in the WNP.
In addition, only the records with maximum sustained wind speeds exceeding 17 m s\(^{-1}\) were used in the analysis.

To examine the horizontal distribution of changes in TC activity parameters (e.g., intensity, intensification rate, genesis frequency, passage frequency, and translation speed), we adopted an overlapping latitude–longitude gridding method (Kim et al. 2010). Based on the center of any grid point, a \(5^\circ \times 5^\circ\) square window was set up, in which annual TC activity parameters were calculated by averaging all 6-hourly records observed during the TC season for the year. By shifting the window at a \(1^\circ\) interval in latitude and longitude (far less than the size of the window), gridded data with a \(1^\circ \times 1^\circ\) horizontal resolution were constructed, by which the distribution of annual TC activity parameters was spatially smoothed without modifying the original data properties. As no averaged records were obtained in grid points where no TCs passed during a year, these grid points were given a missing value and were not used for the calculation of trends in TC intensity and intensification rate. However, as the non-recorded points are meaningful for TC genesis and passage frequency, the zero values were included in the calculation of trends for these two TC activity parameters. The linear trends in TC intensity and intensification rate (genesis and passage frequency) were not calculated for the grid points where missing (zero value) years were detected for more than 14 (25) years. The selected threshold years are sufficient to show the major regions with significant trends. It is noted that replacements in the threshold years just alter the extension of the area without changing the calculated trends.

Various datasets were obtained to analyze large-scale environments related to TC intensity. The datasets include atmospheric reanalysis data from the European Centre for Medium-Range Weather Forecasts [ECMWF; i.e., the ECMWF Interim Re-Analysis (ERA-Interim); Dee et al. 2011], SST data from the National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation SST, version 2 (OISST2; Reynolds et al. 2002), and oceanic analysis data from the National Centers for Environmental Prediction (NCEP) Global Ocean Data Assimilation System (GODAS; Behringer et al. 1998). ERA-Interim provides the atmospheric state variables for the troposphere, such as temperature, relative vorticity, horizontal/vertical winds, specific/relative humidity, and sea level pressure at a \(1.5^\circ\) horizontal resolution for the period 1979–2010, while NCEP GODAS provides the ocean subsurface temperatures at 40 geometric depth levels with a \(1/3^\circ\) horizontal resolution for the period 1980–2010. The NOAA OISST2 covers the shorter period starting from 1982, at a \(1^\circ\) horizontal resolution.

3. Results

a. Inhomogeneous spatial distribution of trends in TC intensity

Figure 1a shows the spatial pattern of climatological TC intensity during the analysis periods (1977–2010).
The maximum of climatological TC intensity is located in the northeast of Taiwan. The region is one of the main paths for TCs recurving from the eastern tropics to the midlatitudes over the WNP (Ho et al. 2004; Wu et al. 2005). Because these types of the recurving TCs generally have enough time to develop over the warm ocean basin, they can become very powerful TCs, thereby accounting for the location of maximum intensity. The contour intervals are relatively broad around this maximum core region, while narrowing down toward the East Asian continent, the Philippines, and eastern tropics. This implies that TCs may rapidly develop (decay) over the eastern tropics (after landfall). In addition, the coastal regions facing the South China Sea generally suffer less from intense TCs than those bordering the WNP including the Philippines, Taiwan, eastern China, South Korea, and Japan.

The overview of spatial distribution of the linear trends in TC intensity can be characterized by opposite signs between the southern and northern WNP (Fig. 1b). The subregions were properly divided into two regions by considering the signs, significances, and localities: southern Japan and its southeastern ocean (SJ; 25°–36°N, 130°–148°E) and the tropical Philippine Sea (TP; 10°–18°N, 127°–144°E). In the SJ region, positive signs are predominant, whereas negative signs are prevalent in the TP region. In general, all five TC datasets show consistent signs of trends with the average of the trend of the five datasets in the grid points where the trends are discernible. The grid points with three or more datasets showing a statistically significant trend at the 90% confidence level mostly reside in the SJ and TP regions. The grid points in which all datasets show a significant trend are mainly distributed over the southeastern ocean in SJ and the central part of TP. In contrast, the imperceptible trends for all datasets are observed around the region where the climatological maximum of TC intensity appeared (Fig. 1a). This result is in line with Webster et al. (2005), who found that there is a negligible change in the time series of the annual maximum TC wind speeds over the WNP.

Figure 2 presents the time series of TC intensity for the individual datasets, averages, and their confidence intervals of the four best-track data (except the UW/NCDC data as we discussed in section 2) in the SJ and TP. The confidence interval ($p < 0.1$) was acquired from bootstrap resampling with 10,000 replicates based on the annually four values of best-track data (Hall 1988), assuming that the accuracy of each best-track data may be...
same. To test the significance of trend, both the two-tailed \( t \) test and nonparametric Mann–Kendall test were utilized together, because the linear trend is generally subject to some assumptions. The merits of the nonparametric Mann–Kendall test are that 1) it is unnecessary to hypothesize normality and 2) the test is less sensitive to outliers because it detects the monotonic trend, which is not essentially linear (see Table 1) (McLeod et al. 1990; Chu et al. 2012).

In the SJ region, all the time series show gradual increments (Fig. 2a). The trends range from \( +0.64 \) to \( +1.54 \) m s\(^{-1}\) decade\(^{-1}\), which are all positive, but only the UW/NCDC and JTWC data show statistically significant trends at the 90% confidence level (Table 1). In the case defining a smaller box over the southeastern ocean of Japan (25°–33°N, 140°–144°E), the increasing trends of all-time series became steeper and statistically significant at the 90% confidence level except the RSMC data (Table 1). The increasing intensity in the SJ region was also reported in previous studies (e.g., Park et al. 2011; Wu and Zhao 2012). Park et al. (2011) demonstrated that the TC intensity has increased in midlatitude East Asia since the late 1970s, which was caused by SST warming, tropospheric moistening, and weakening of vertical wind shear.

In contrast, the time series in the TP region show overall decrements (Fig. 2b). The negative trends range from \( -2.75 \) to \( -0.81 \) m s\(^{-1}\) decade\(^{-1}\), all of which are statistically significant at the 90% confidence level except the UW/NCDC data (Table 1). The decrements were more evident in the central oceanic area of TP (11°–17°N, 133°–144°E), in which all datasets presented robust and significant decreasing trends. The trend of the UW/NCDC data was not significant in the case applying the Mann–Kendall test. However, the decreasing intensity in the TP region is thought to be as evident as the increasing intensity in the SJ region, because the trend was significant when using the Student’s \( t \) test and its decreasing rates were generally larger than those in the SJ region.

The time series show large interannual fluctuations of TC intensity (Fig. 2). Therefore, calculation of the trend was subject to their start and end years. To examine the sensitivity of the trend to a chosen period, sliding regressions using 20-yr moving windows (i.e., 1977–96, 1978–97, …, 1991–2010) were calculated for the time series of TC intensity estimated from the average of the four best-track and UW/NCDC data in the SJ (Fig. 3a) and TP (Fig. 3b) regions. For the best-track data, the subtrends for the mean intensity were only displayed.

![Fig. 3](image-url)
because the subtrends for individual best-track datasets show very similar variations, with the significant correlation coefficients exceeding 0.75.

In the SJ region, the positive subtrends were dominant for the average intensity of the four best-track data, with only two 20-yr windows (1981–2000 and 1990–2009) showing weak negative trends (Fig. 3a). The former weak negative trend during 1981–2000 and two near-zero trends surrounding this period (1980–99 and 1982–2001) can be attributed to both the positive peak near the start year (1982) and the negative peak near the end year (1999). An analogous explanation holds for the latter one (1990–2009) and two weak positive trends (1989–2008 and 1991–2010). On the other hand, the strongest upward trends were found for the 20-yr windows in the early (1977–96 and 1978–97) and middle periods (1985–2004 and 1986–2005), where the start year is near the lowest point and the end year is near the peak point (Fig. 2a). The sign of the trend became more sensitive to the start and end years when applying a shorter period. Nevertheless, it is unlikely to be a fallacy that TC intensity in the SJ region has increased since 1977, because the majority of subtrends (Fig. 3a) as well as all the trends for the entire period (Table 1) shows upward tendencies.

On the other hand, the TP region showed conspicuous negative trends in TC intensity for the four best-track data for all subperiods (Fig. 3b). The negative trends in this region were more prominent than the positive trends in the SJ region. The negative subtrends were statistically significant for the windows during the midperiods (1981–2000, 1982–2001, 1983–2002, and 1984–2003), which start near the peak of 1985 and end near the lowest year of 1999. The negative trends were relatively weaker without statistical significance for the windows in the early (1977–96, 1978–97, 1979–98, and 1980–99) and the middle-to-late periods (1985–2004, 1986–2005, 1987–2006, 1988–2007, 1990–2009, and 1991–2010). Several dips around 1980 and the strong peak in 2009 contributed to the weakening of the negative subtrends (Fig. 2b). The 20-yr sliding regressions using the UW/NCDC data also showed consistent signs of the trend with those of the best-track data for both the SJ and TP regions (open bars in Fig. 3). Although uncertainty from the data quality remains, these results suggest that the opposing trends found in the SJ and TP regions certainly exist during the analysis period (1977–2010).

In addition to the two main analysis regions, some regions exhibited marginal trends with consistent signs among datasets (i.e., dot points), although only one or two sets have statistical significance. For example, the southern coast of China and the Philippines were likely to experience weakly decreasing TC intensity, while in the intermediate region between SJ and TP both negative and positive signs were observed. In the following subsection, the changes in these regions will be shortly described after discussing the changes in the main regions.

b. Spatial distribution of trends: Intensification rate, genesis frequency, passage frequency, and translation speed

If the intensity trends were computed over the whole life of TCs, the basin-total trends would diverge in various TC best-track datasets as shown in Wu et al. (2006), reflecting the effect of reversal trends in the TP and SJ regions. That is, it is important to interpret the mechanism of the differing trend patterns observed in the two regions. Prior to investigating relevant large-scale environments, we examined the spatial distribution of trends in other TC activity parameters (e.g., TC intensification rate, genesis, passage frequency, and translation speed) that could substantially affect the TC intensity.

1) INTENSIFICATION RATE

Here, we present the TC intensification rate, which is defined as an average wind speed change in a unit hour in a grid (Fig. 4). Figure 4a demonstrates the spatial pattern of climatological TC intensification rate and translational direction for the analysis period (1977–2010). The negative values are distributed over the coastline of the East Asian landmass and north of 25°N, while the positive values are spread over most of the entire ocean basin in the tropics. This climatological pattern naturally accounts for the spatial distribution of climatological TC intensity because of the straightforward physical relationship between the TC intensification rate and the TC intensity (Fig. 1a). Considering the climatological translational direction, it is expected that the zero line of TC intensification rate in the WNP will nearly coincide with the ridges of TC intensity (crosses in Fig. 1a). In other words, TCs generally reach their strongest intensity where the TC intensification rate changes its sign from positive to negative. Thus, the spatial distribution of trends in TC intensification rate can account for the spatially inhomogeneous trends in TC intensity with the aid of the climatological translational direction.

As shown in Fig. 4b, the TC intensification rate has been slightly getting slower in the eastern part of TP while being significantly boosted over the vast tropical and subtropical areas. As the TCs in the tropics generally move westward or northwestward according to the climatological translational direction (Fig. 4a), this pattern of TC intensification rate indicates that the TCs
from the easternmost or eastern outer region of TP experienced the suppression of TC intensification and became weaker over the eastern part of TP. In addition, the increasing intensification rate with climatological west-northwestward TC translation over the center of TP accounts for the moderate decrease of TC intensity over the eastern offshore of the Philippines (Figs. 1b and 4b). Continuing to the west, the again weakened intensification rate just outside of the western boundary of the TP region explains the decreasing intensity trends over the Philippines (Figs. 1b and 4b). The intensification rate shows the largest upward trends over the intermediate region between the TP and SJ regions. In a climatological sense, TCs passing over this region move northward (Fig. 4a), so that the more rapid intensification rate there could greatly enhance the TC intensity to its north (i.e., the SJ region; Figs. 1b and 4b). This interpretation is reasonable because the trends in TC intensification rate are trivial in the SJ region, indicating no evidence of further local strengthening of TCs therein. Thus, the spatial distribution of the trend of TC intensification rate is an effective method to interpret the trend of TC intensity.

However, the spatial distribution of TC intensification rate is not always a perfect indicator for the TC intensity as indicated by the following: 1) The intensification rate has increased over the northern South China Sea, but the TC intensity has slightly weakened along the southern coast of China (Figs. 1b and 4b). 2) The positive trends of TC intensity southeast of Taiwan was not as large as those in the SJ region even though the TCs passing there were also very likely to experience the strong positive trends of TC intensification rate over the western TP region and its north. 3) The declining trends of TC intensification rate are not large enough to explain the remarkable downturn of TC intensity in the TP region. Accordingly, interpretation solely by the TC intensification rate with the climatological translation direction is imperfect for those regions. This naturally leads us to investigate other parameters, such as TC genesis frequency, passage frequency, and translation speed.

2) GENESIS FREQUENCY

The location of TC genesis is an influential factor for TC intensity because TCs can have more or less time to develop over a warm ocean basin depending on the genesis location (Camargo and Sobel 2005; Wu and Wang 2008). As seen in Fig. 5a, the frequency of TC genesis has decreased considerably over the large domain of the tropical eastern WNP basin (i.e., east of the TP region) and the west and east sides of the Philippines. Conversely, it has increased in the south of Hainan Island, northern South China Sea, north of the TP region near 130°E, and the heart of the TP region.

TCs at their genesis locations must have intensity slightly above 17 m s⁻¹ in the best-track data, because this is the minimum threshold of TC intensity in this study. Thus, it is natural to say that the newly generated TCs are weaker than the developed TCs. Considering...
this point, the average intensity in the grid point should be reduced if more newly generated TCs are included. Therefore, with the genesis frequency, we can improve the insufficient interpretation of the trends in TC intensity by the TC intensification rate in three regions: the southern coast of China, southeast of Taiwan, and the TP region.

First, more TC genesis in the northern South China Sea and less TC genesis to the west and east of the Philippines (Fig. 5a) may contribute to the slight weakening of landfalling TCs in the southern coast of China (Fig. 1b). That is, the trends of TC genesis counteracted the trends of TC intensification rate in the northern South China Sea. As the two responsible factors offset each other, the changes in these regions could not be prominent. Second, the increasing genesis frequency north of the TP region near 130°E may explain the weak increasing trends of TC intensity southeast of Taiwan. Third, the slight increase of new TCs forming in the TP region can have a role in enhancing the weakening trends of TC intensity there. In addition, the great decrease in TC genesis frequency over the large domain of the tropical eastern WNP basin can contribute to the weakening of TC intensity in the TP region together with the suppressed TC intensification rate (Figs. 4b and 5a). This is because fewer numbers of developed TCs reached the eastern TP region. Considering that TCs in the low latitudes, where most TCs form, will generally intensify and propagate westward, this seems to be a reasonable speculation (Fig. 4a).

The TC genesis frequency can account for large parts of the decreasing intensity trends over the tropical and subtropical WNP regions, in combination with the intensification rate. However, the trivial local genesis frequency in the SJ region hardly affected the TC intensity therein because the climatological genesis location lies south of 25°N (not shown). Accordingly, it is natural to say that the trends of TC intensity in the SJ region were determined by the intensification rate at the southern entrance region of the domain.

3) PASSAGE FREQUENCY

The TC passage frequency was defined as the seasonal number of TCs passing through each grid box, which is a function of genesis locations and tracks (Ho et al. 2004). The TC intensity in a certain grid also depends on the TC passage frequency of intense TCs (Wu and Wang 2008). Figure 5b shows that the TC passage frequency has decreased significantly over the two large tropical areas, the South China Sea and the eastern subtropical area of the WNP, whereas it has increased significantly around Taiwan and marginally near the east coast of Japan. Considering the distribution of genesis frequency trends (Fig. 5a) and climatological translation direction (Fig. 4a), it is natural to expect large changes in TC passage frequency to the west and north of the regions where large changes in TC genesis frequency were observed (Fig. 5b). Previous studies have already identified this pattern of change (e.g., Tu et al. 2009; Park et al. 2011). More frequent TC movements to Taiwan and Japan were related with the enhanced monsoon trough and the northeastward shift of the WNP subtropical high.

Although both the spatial patterns of trends in TC passage frequency and intensity have similar signs in the
tropics and midlatitudes (Figs. 5b and 1b), the significant regions are not exactly coincident with each other. For example, the major decreasing region of TC intensity (i.e., the TP region) flanks on the two large regions with declining TC passage frequency. The significant increments of TC intensity appear in the SJ region, in which no considerable changes in TC passage frequency are detected. Taiwan and its vicinity do not show an appreciable intensity change in spite of the significant increase in TC passage frequency.

In particular, the area around Taiwan is interesting to discuss in more detail. As a majority of TCs affecting Taiwan comes from the southeast (Chu et al. 2010), it is natural to expect that a robust upward trend in TC passage frequency (Fig. 5b), together with the intensification rate (Fig. 4b), would lead to an increase in TC intensity. However, the intensity trends are trivial or slightly negative in the vicinity of Taiwan (Fig. 1b), which leads us to interpret that the increasing genesis frequency between 120° and 130°E along 20°N (Fig. 5a) totally offsets the other two factors. To sum up, the changes in TC passage frequency have less to do with TC intensity, compared with the changes in TC intensification rate and genesis frequency.

4) TRANSLATION SPEED

The combination of TC intensification rate and genesis frequency successfully explains the spatial distribution of trends in TC intensity over the WNP. In addition, the possible contribution of TC translation speed needs a validation because it has been known to affect TC intensity through modulation of the effect of TC-induced SST cooling on the intensification rate (Bender and Ginis 2000). Its effect on the intensification rate can vary depending on various factors, such as the residence time, depth of the mixed layer, and TC-induced wind speed (Wada and Usui 2007; Lin et al. 2009).

To examine the effect of TC translation speed on the intensification rate, we examined the spatial distribution of the trends in TC translation speed (Fig. 6a). In contrast with the aforementioned studies, the translation speed seems not to be an effective factor on the intensification rate in a climatological sense because it has not been considerably changed over the regions where the intensification rate has changed significantly during the analysis period (Figs. 4b and 6a). The unchanged translation speed is inconsistent with Chu et al. (2012), who found a significant decrease in TC translation speed in the subtropics during the period of 1958–2010, which is attributed to the shorter analysis period in this study (i.e., 1977–2010) than that of Chu et al. (2012). Thus, not all changes in the intensification rate were influenced by the changes in the translation speed during the analysis period of this study.

To clarify this point, we calculated changes in TC intensification with consideration of the residence time of TCs in a grid box (i.e., intensification rate \times \text{residence time}). This modified intensification rate is more physically linked with the translation speed, as the residence time in a grid box itself implies the translation speed. The spatial pattern of trends in the modified intensification rate is almost similar to that of the original intensification rate (Figs. 4b and 6b). This indicates that the intensification rate is a major factor to explain the spatial distribution of the trends in TC intensity, rather than the translation speed. In the following subsection,
the intensification rate and genesis frequency will be mainly discussed with large-scale tropospheric and oceanic environmental changes.

c. Changes in thermodynamic and dynamic fields

To reveal possible mechanisms for the regionally varying trends in TC intensity via the combination of the trends in TC intensification rate and genesis frequency, we examined both thermodynamic- and dynamic-related variables for the TC development. The thermodynamic variables were represented by three factors: maximum potential intensity (MPI), SST, and 26°C isothermal ocean depth (Gray 1979; Emanuel 1988; Wada and Usui 2007; Wada and Chan 2008). The MPI is generally referred to as the theoretical maximum intensity of a TC (Emanuel 1991). The basic concept of the MPI came from the Carnot cycle, which was regarded as the energy cycle of a TC (Emanuel 1987). The MPI was computed by

\[
\text{MPI} = \mu \sqrt{(T_s/T_0)(CKCD)(\text{CAPE}_s - \text{CAPE})},
\]

where \(\mu\) is the factor to reduce gradient wind to a 10-m wind; \(T_s\) is the SST; \(T_0\) is the outflow layer temperature; CKCD is the ratio of the exchange coefficient for enthalpy to the drag coefficient; \(\text{CAPE}_s\) is the saturation convective available potential energy at the radius of maximum winds; and \(\text{CAPE}\) is the convective available potential energy at the radius of maximum winds. The values of \(\mu\) and CKCD were taken to be 0.8 and 0.9, respectively. The MPI comprises two independent variables, CAPE and thermal efficiency (Emanuel 1991). CAPE denotes the limit of potential TC energy under the given status of variables in the troposphere and surface, such as moisture, temperature, and pressure. If the troposphere is unstable, a high CAPE can be achieved. The thermal efficiency can be thought of as how many rates of the potential TC energy are released into the troposphere to generate the TC wind. Thus, high MPI can be regarded as a favorable thermodynamic condition for the TC development. As shown in Fig. 7a, the MPI has increased significantly over most areas of the WNP, which means that the thermodynamic condition over the WNP basin has become more favorable for the TC development.

In addition to the MPI, oceanic conditions such as the depth of the 26°C isotherm and SST were also investigated. The depth of the 26°C isotherm can determine how deep the warm mixed layer of the ocean extends below the ocean surface, which is an essential factor to calculate the ocean thermal energy (i.e., TC heat potential; Gray 1979). This parameter provides important oceanic information for the TC intensity and its intensification rate (Wada and Usui 2007) because a warm SST of more than 26.5°C is a necessary condition for the TC development (Gray 1979). It is noted that the SST can easily drop below 26.5°C in those regions where the depth of the thermocline is shallow, because TC-induced strong winds can cause mixing of upper and lower seawater, thereby effectively cooling the oceanic mixed layer by the upwelling process (e.g., Lin et al. 2005; Lin et al. 2008; Wu et al. 2008). As shown in Fig. 7b, the 26°C isotherm has deepened across the WNP basin, although the deepening trend is relatively weaker along the subtropical belt of the Philippine Sea, the reason for which is unclear and needs a further investigation. The deepening of 26°C isothermal depth would provide oceanic conditions favoring the TC development in the WNP.
The SST has also increased considerably over the entire WNP, as is well known (Fig. 7c) (Emanuel 2005; Webster et al. 2005). Thus, it is represented that the atmospheric and oceanic thermodynamic fields have been more favorable for TC development over almost the entire WNP, but these results cannot fully explain the spatially inhomogeneous distribution of the TC intensity trends that results from the combined changes in TC intensification rate and genesis frequency. For instance, the TC intensification rate and genesis frequency have decreased around the Philippines and the tropical eastern WNP. This may imply that the thermodynamically favorable conditions over these regions have been masked by unfavorable dynamic conditions.

The anomalous SST distribution is informative to understand the large-scale atmospheric circulations. As shown in Fig. 7c, the SST trends vary in signs from the west to the east over the tropical Pacific Ocean. This reflects the steeper zonal SST gradient (i.e., La Niña-like SST pattern) (Trenberth 1997), owing to stronger warming in the western Pacific warm pool region. This increasing SST gradient during last three decades can be dynamically linked with the enhancement of the Walker circulation through the Bjerknes feedback in the near-equatorial ocean–atmosphere coupled process (Sohn and Park 2010; An et al. 2012). Figure 8a shows in the cross section of the trends in the zonal and vertical winds averaged between the latitudes of $5^\circ$S and $10^\circ$N. The pattern clearly presents an increasing clockwise overturning (i.e., Walker circulation) over the equatorial western-to-central Pacific sector.

Associated with this enhanced Walker circulation, dynamic environments relevant to the TC activity were also examined, such as the horizontal winds and relative vorticity at 850 hPa, and the magnitude of the vertical wind shear between 200 and 850 hPa (Figs. 8b,c). The increasing low-level easterlies near the equator have formed the meridional wind shear, inducing the more anticyclonic low-level flows along the tropics including the eastern TP region (Fig. 8b). On the other hand, the strengthening of the Walker circulation also has increased the magnitude of the vertical wind shear in the tropical North Pacific, with the maximum increments near the date line (Fig. 8c). The eastern part of TP region is also the part of a region of increased vertical wind shear, although it is not significant. In the eastern part of TP area, both the low-level flows and vertical wind shear have been more unfavorable, so that it would be reasonable to regard them as the main reasons for the decrease in TC intensification rate and genesis frequency, thereby reducing TC intensity in the TP region. A similar interpretation can be made to explain the decreasing TC intensity around the Philippines (i.e., the western TP region). The anomalous anticyclonic flow over the Philippines may be a factor to suppress TC intensification rate and genesis frequency therein, resulting in the weaker TCs in the Philippines and the southern coast of China (Figs. 1b, 4b, 5a, and 8b).

In contrast, there is an anomalous large-scale cyclonic flow and weak vertical wind shear in the western part of TP and parts of the subtropics including the northern part of the South China Sea, the south of Taiwan, and
Japan (Figs. 8b,c). The anomalous cyclonic flow is thought to be the result of anomalous anticyclonic flows in the tropics, and the weak vertical wind shear is closely related to the recently weakened and northward-shifted jet stream (e.g., Kwon et al 2007; Archer and Caldeira 2008). Both of these changes could increase TC intensification rate and genesis frequency with the thermodynamic conditions over the same regions, thereby accounting for the significantly stronger TCs in the SJ region and also weaker TCs in the southern coast of China related to increasing TC genesis frequency over the northern South China Sea.

Consequently, it is essential to consider both thermodynamic and dynamic conditions when investigating the spatial distribution of linear trends in TC intensity through the intensification rate and genesis frequency. In addition, in the TP regions, dynamic variables seem to have a more decisive role for TC intensification rate and genesis frequency than thermodynamic variables.

4. Concluding remarks

This study examined the spatial distribution of trends in TC intensity for the period 1977–2010 using five TC datasets. While unclear and inconsistent long-term changes in TC intensity were observed with regard to the whole life cycle of TCs in the WNP (Wu et al. 2006), consistent trends for all TC datasets were newly identified in the TP and SJ regions (Fig. 1b). For all TC datasets, the TC intensity consistently strengthened in the SJ region, whereas it weakened in the TP region. These opposing trends in TC intensity were mainly affected by TC intensification rate and genesis frequency. In particular, the changes in TC intensification rate could mostly support spatial distribution of linear trends in TC intensity (Fig. 4b). The suppressed (increased) TC intensification rate in the eastern part of TP (the subtropics and the western part of TP) weakened (strengthened) TCs in the TP (SJ) region. However, in the tropics, the TC intensification rate could not sufficiently account for the spatial change in TC intensity, but rather the TC genesis frequency could explain it: a decrease (increase) in the eastern part of TP and the Philippines (the center of TP and northern part of the South China Sea) (Fig. 5a). Thus, it is suggested that both TC intensification rate and genesis frequency have been the decisive factors for regionally inhomogeneous intensity trends.

The increased TC intensification rate and genesis frequency in the subtropics and the western part of TP were driven by both thermodynamic and dynamic climate variables: the high MPI, deeper 26°C isotherm, warm SST, weak vertical wind shear, and anomalous cyclonic large-scale flow in the same region (Figs. 7 and 8). However, the reduced TC intensification rate and genesis frequency around the east of TP were attributable to unfavorable dynamic variables: the low-level anomalous anticyclonic flows and strong vertical wind shear therein (Fig. 8), which canceled out and overcame the effects of the favorable thermodynamic conditions (Fig. 7). These changes in dynamic variables were closely linked with the recently enhanced Walker circulation over the Pacific. In conclusion, it is suggested that the dynamic variables (e.g., relative vorticity and vertical wind shear) could have more conclusive roles for the changes in TC intensity in the eastern part of TP. This conclusion supports that of Chan (2009), who suggested that dynamic fields might be more dominant factors than thermodynamic fields in determining TC intensity over the WNP, as thermodynamic environments could not fully explain the variation in TC intensity.

However, it is still unknown what each factor’s exact role in contributing to the TC intensity may be. To clarify this, it is necessary to perform model experiments. In addition, it is also not clear that the recent spatial changes in TC intensity will continue to show the same distribution in potential climate change. In this study, it is shown that the dynamic fields, which are very important factors for TC intensity, are highly dependent on the horizontal distribution of SST over the Pacific—the predictions for which show considerable inconsistency between different climate models (Sugi et al. 2009). Thus, it is almost impossible to foresee the future spatial changes in TC intensity. Nevertheless, because it is possible to check whether the dynamic variables will remain decisive factors in the warm climate, it might be meaningful, in a future study, to investigate the spatial distribution of long-term changes in TC intensity using distinct SST gradients based on various model simulations.

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