Simulations of the Present and Late-Twenty-First-Century Western North Pacific Tropical Cyclone Activity Using a Regional Model

LIANG WU,* CHIA CHOU,†,‡, CHENG-TA CHEN,‡, RONGHUI HUANG,* THOMAS R. KNUTSON, & JOSEPH J. SIRUTIS,* STEPHEN T. GARNER,* CHRISTOPHER KERR,* CHIA-JUNG LEE,* & YA-CHIEN FENG†

* Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China
† Research Center for Environmental Changes, Academia Sinica, Taipei, Taiwan
‡ Department of Atmospheric Sciences, National Taiwan University, Taipei, Taiwan
‡ Department of Earth Sciences, National Taiwan Normal University, Taipei, Taiwan
& NOAA/Geophysical Fluid Dynamics Laboratory, Princeton, New Jersey

(Manuscript received 26 November 2012, in final form 14 January 2014)

ABSTRACT

A high-resolution regional atmospheric model is used to simulate present-day western North Pacific (WNP) tropical cyclone (TC) activity and to investigate the projected changes for the late twenty-first century. Compared to observations, the model can realistically simulate many basic features of the WNP TC activity climatology, such as the TC genesis location, track, and lifetime. A number of spatial and temporal features of observed TC interannual variability are captured, although observed variations in basinwide TC number are not. A relatively well-simulated feature is the contrast of years when the Asian summer monsoon trough extends eastward (retraces westward), more (fewer) TCs form within the southeastern quadrant of the WNP, and the corresponding TC activity is above (below) normal over most parts of the WNP east of 125°E. Future projections with the Coupled Model Intercomparison Project phase 3 (CMIP3) A1B scenario show a weak tendency for decreases in the number of WNP TCs, and for increases in the more intense TCs; these simulated changes are significant at the 80% level. The present-day simulation of intensity is limited to storms of intensity less than about 55 m s\(^{-1}\). There is also a weak (80% significance level) tendency for projected WNP TC activity to shift poleward under global warming. A regional-scale feature is a projected increase of the TC activity north of Taiwan, which would imply an increase in TCs making landfall in north China, the Korean Peninsula, and parts of Japan. However, given the weak statistical significance found for the simulated changes, an assessment of the robustness of such regional-scale projections will require further study.

1. Introduction

Tropical cyclones (TCs) are among the world’s most extreme natural disasters and the western North Pacific (WNP) is by far the most active TC basin. Roughly one-third of the global TC numbers and 39% of the TC energy (Maue 2011) occur there. Therefore, it is important to investigate TC activity in the WNP both in the present and in projections under global warming.

Previous observational studies have shown that the interannual variations of TC activity in a number of basins are closely related to large-scale phenomena, such as El Niño–Southern Oscillation (ENSO) (Camargo et al. 2007). In particular, observational studies have noted that in the WNP, a longitudinal shift in the location of TC formation and activity occurs between El Niño and La Niña years, although there is no significant interannual correlation between ENSO and storm counts over the entire WNP (Camargo and Sobel 2005; Camargo et al. 2007; Chan 1985, 2000; Wang and Chan 2002; Wu et al. 2012). Wu et al. (2012) proposed that the monsoon trough (MT) is a key factor that connects ENSO and TC activity over the WNP. Some models have successfully simulated aspects of observed interannual variability of TC activity over the WNP. For example, Camargo et al. (2005) and Zhao et al. (2009) demonstrated some simulation skill for year-to-year variation in the total WNP TC number using different general circulation models (GCMs); and Emanuel et al. (2008) used a statistical/dynamical downscaling

Corresponding author address: Prof. Cheng-Ta Chen, National Taiwan Normal University, Department of Earth Sciences, 88, Section 4, Ting-Chou Road, Taipei 116, Taiwan. E-mail: chen@rain.geos.ntnu.edu.tw

DOI: 10.1175/JCLI-D-12-00830.1

© 2014 American Meteorological Society
model and reproduced the variability of annual power dissipation of observed TCs over the basin. The 50-km grid atmospheric GCM of Zhao et al. (2009) has lower skill in capturing the interannual variability of TC number over the WNP than over the Atlantic, when driven with observed sea surface temperatures (SSTs). Camargo et al. (2005) demonstrated that this is largely a result of a strong relationship between ENSO and TC number of the Atlantic. Iizuka and Matsuura (2008) used a coupled ocean–atmosphere general circulation model (CGCM) to examine the relationship between WNP TC activity and ENSO. Their results show that the changes of model TC activity are related to their model’s ENSO. These modeling studies have shown potential skill in reproducing the interannual variability of TC activity. However, they did not fully investigate the details of interannual variability and the linkage between the WNP TC activity and the MT. A detailed understanding of TC climatology and interannual variability in present-day simulations could lead to greater confidence in future projections of TC activity, since a realistic simulation of the response of model-simulated TC activity to the large-scale climate forcing is a necessary (though not sufficient) condition for confidence in long-term TC climate change projections made with such a model.

It has long been recognized that there are potential impacts of global warming on TC activities, but it remains particularly uncertain how TC activity would change in the future climate over the WNP. Most previous studies have predicted a future decrease in total TC genesis frequency, but an increase in more intense TCs (e.g., Bengtsson et al. 2007a,b; Knutson et al. 1998, 2010; Murakami et al. 2011, 2012b; Oouchi et al. 2006; Zhao et al. 2009). Nevertheless, there are some studies projecting a decrease in TC intensity (Hasegawa and Emori 2005) or an increase in TC genesis frequency (Stowasser et al. 2007), although the Hasagawa and Emori model is a relatively low-resolution (120-km grid) model and does not simulate TC intensity realistically. A recent study by Emanuel (2013) found that the frequency of downscaled TCs increases during the late twenty-first century in most locations using the projected climate changes from several models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) under the representative concentration pathway 8.5 (RCP8.5) scenario. This finding contrasts with that of investigations of WNP TC activity simulated explicitly in global climate models (Camargo 2013) or downscaled dynamically over the Atlantic basin (Knutson et al. 2013). Overall, there is still a large intermodel range in projected late-twenty-first-century changes in the WNP basin-scale TC activity found in different studies (Knutson et al. 2010). The detailed spatial distribution of projected TC changes within the individual basin is even more uncertain, yet such regional detail is very important for climate impact studies. Although several models project that the TC tracks and activity will shift eastward in the WNP (Li et al. 2010; Murakami et al. 2011; Ying et al. 2012), the detailed spatial distribution of this change varies among different model projections. There is increasing evidence that the spatial distribution of future SST changes and associated tropical circulation changes are key ingredients in shaping the projected regional responses of TC activity in different models (Knutson et al. 2010, 2013; Murakami et al. 2012a; Vecchi and Soden 2007).

Fundamental tests of model simulations of present TC activity include assessments of how well models can capture the geographical distribution, seasonality, interannual variability, and trend of TC activity (Bengtsson et al. 2007a; Camargo and Zebiak 2002; Knutson et al. 2007). Although many modeling studies have performed reasonably well in simulating the climatological mean and seasonal cycle of WNP TC activity, few have demonstrated skill in simulating variability of TC activity on interannual or longer time scales. Biases in simulated TC activity may be related to unrealistic large-scale environments simulated by models with coarse resolution (Manganello et al. 2012; Strachan et al. 2013). Knutson et al. (2007) used a high-resolution regional model with observed SST and large-scale atmospheric conditions [nudged on large scales toward a smoothed component of the observed National Centers for Environmental Prediction (NCEP) atmospheric reanalysis] and successfully reproduced the interannual variations in basinwide Atlantic hurricane frequency (1980–2006). A subsequent (unpublished) update through 2012 shows a modest positive bias in the long-term trend, perhaps associated with temporal homogeneities in the reanalysis (Vecchi et al. 2013). Knutson et al. (2008, 2013) projected future changes of Atlantic hurricane activity using this model forced by changes in SSTs and atmospheric conditions as simulated in climate model experiments in phase 3 of the Coupled Model Intercomparison Project (CMIP3; Meehl et al. 2007) using the A1B emission scenario.

TC genesis and development over the WNP display different characteristics from those over the Atlantic in both large-scale environmental factors and precursor perturbations (Fu et al. 2012). To assess the skill of the model in reproducing the present WNP TC activity and investigate future projected WNP TC activity, in this study we apply the same 18-km-grid regional model (ZETAC) framework of Knutson et al. (2007, 2008) to the WNP basin.

The article is arranged into six sections. Section 2 describes the data, data processing methods, model, and
experimental design. In section 3, we present the climatology of TC activity over the WNP in the present-day simulation. Section 4 examines the interannual variability of WNP TC activity and its relationship with the monsoon trough as simulated in the model. The projected future changes in TC activity over the WNP are presented in section 5. A summary of results is given in section 6.

2. Data and methods

a. Data

To compare model results with WNP basin observations of TCs, the best-track dataset from the Joint Typhoon Warning Center (JTWC) for the period 1980–2008 is used to construct the accumulated cyclone energy (ACE; Bell et al. 2000), track and intensity function (TIF; Wu et al. 2011), and other indices of the TC activity. The JTWC data include 6-hourly (0000, 0600, 1200, and 1800 UTC) center locations (latitude and longitude) and intensities (maximum 1-min mean sustained surface wind speeds at 10-m height) of TCs. We categorized tropical cyclones into two groups, tropical storm (TS) and typhoon (TY), according to the lifetime maximum wind speed thresholds of 17 and 33 m s$^{-1}$, respectively. The time of the TS genesis ($>17$ m s$^{-1}$) defines the start of the TC lifetime, and the end is defined as the time when it disappears from the track records. The International Best Track Archive for Climate Stewardship (IBTrACS) World Meteorological Organization (WMO) version v03r03 (referred to hereinafter simply as IBTrACS; Knapp et al. 2010) is also used to permit comparison of TC activity archived by different operational centers. IBTrACS uses 10-min mean wind speed at 10-m height for the maximum sustained wind estimate. Both datasets use a surface wind speed of 17 m s$^{-1}$ for the TS threshold, but 29 m s$^{-1}$ in IBTrACS and 33 m s$^{-1}$ in JTWC are used for TY.

Specified observed SSTs and several meteorological variables from the NCEP–U.S. Department of Energy (DOE) reanalysis version 1 (hereinafter called NCEP1; Kalnay et al. 1996) during the period 1980–2008 are used as the initial and boundary inputs and for interior spectral nudging of the model. The Global Precipitation Climatology Project (GPCP) dataset (Adler et al. 2003) is used for the observed precipitation. Only the data from June to November, which is the main TC season in the WNP, are used.

b. Methods

Several indices have been employed to represent the TC activity in this study. The accumulated cyclone energy (Bell et al. 2000) is used to evaluate the aggregate activity of TCs. The tropical cyclone track and intensity function (TIF) is constructed as in Wu et al. (2012) as another index of TC activity. The TIF index combines the ACE index with a cyclone track density function (TDF; Anderson and Gyakum 1989) in a given region over a period of time. Essentially, TIF is defined as the space and time smoothed value obtained by dividing the ACE in each grid box by the total ACE of TCs. An empirical orthogonal function (EOF) analysis (Lorenz 1956) is employed to extract dominant interannual variations of the TIF. The obtained EOF modes could provide physical insight into the principal spatiotemporal modes of variability of TC activity over the WNP.

To examine and illustrate the influence of the monsoon trough on the WNP TC activity, a composite analysis based on the leading EOF time coefficients associated with the trough is performed. The statistical significance of the composite anomalies is estimated by the Student’s $t$ test. In addition, strong monsoon trough (S-MT) and weak monsoon trough (W-MT) years are defined according to the value of the monsoon trough index (Wu et al. 2012) averaged from June to November. The Niño-3.4 index is based on equatorial Pacific SST anomalies averaged over the TC season (June–November).

c. Models and simulation settings

Numerical experiments are conducted using the Geophysical Fluid Dynamics Laboratory (GFDL) Regional Atmospheric Model (ZETAC), the same as in Knutson et al. (2007). The dynamical core of the model is compressible and nonhydrostatic. Sea surface temperatures are specified from NCEP1 (Kalnay et al. 1996) and simulated soil temperature and moisture from a land dynamics model (Milly and Shmakin 2002) are used. The model domain covers the WNP extending from 5° S to 45° N and from 90° E to 155° W with a grid spacing of 1/6° (~18 km) in both zonal and meridional directions, and 45 unevenly spaced vertical levels. More detailed descriptions of the physical processes, boundary layer turbulence, and radiative transfer schemes are given in Knutson et al. (2007).

A “spectral nudging” method is used to keep the simulated state close to the NCEP reanalysis on the large scale. In the interior of the model domain, the wind, temperature, and humidity of the model domain’s zonal and meridional wavenumbers 0, 1, and 2 at all levels are nudged toward the NCEP reanalysis on a relatively slower time scale (12 h). Along the perimeter of the domain, those variables are nudged on a faster time scale (2 h) within a 5° band. No cumulus convection scheme is used. Through the nudging procedure, the model maintains a more realistic large-scale environment, but it remains relatively unconstrained to generate
smaller-scale disturbances within the large-scale state. The model performed well in the Atlantic basin (Knutson et al. 2007).

To study the potential impact of future global warming on TCs, we run the same model for each main typhoon season (June–November) from 1980 to 2006, but altering the June–November mean atmospheric state and sea surface temperatures according to the late-twenty-first-century changes from an ensemble mean of CMIP3 climate models under the A1B emissions scenario. Similar to Knutson et al. (2008), the influence of multidecadal internal variability of individual models is reduced by using the multimodel mean. The patterns of warmer tropical WNP SST, enhanced upper tropospheric warming relative to the surface, and changes in vertical wind shear and low- to mid-atmospheric relative humidity over the main TC genesis region will be discussed in section 5. We acknowledge that it is unrealistic to assume that the large-scale component of interannual and interdecadal climate variation remains unchanged in the future, but we consider the ability of climate models to reliably project such changes to be very limited. Also, the large-scale component of intraseasonal variability was kept the same in the warming run. Although the timing and location of TC genesis over the WNP is strongly affected by intraseasonal variability (Camargo et al. 2009), the current climate models still have difficulty simulating the intraseasonal variability over the tropical Pacific Ocean (Lin et al. 2006). In any case, we exclude the possible impact of changes in intraseasonal variability in view of limited model skill and the uncertainty of projected changes in intraseasonal variability.

In the present study, we used the 6-h data and the following definitions to identify and track TCs simulated in the model:

1) A local 850-hPa relative vorticity maximum exceeds $1.6 \times 10^{-4} \text{s}^{-1}$.
2) The closest local minimum in sea surface pressure, within a 2° radius from the vorticity maximum, is defined as the center of the TC. The difference in sea surface pressure between the TC center and the environment at 5° away from the center is greater than 4 hPa.
3) The maximum 200–500-hPa layer mean temperature is greater than the environmental temperature (at 5° away from the warm core) by at least 3°C and within a 2° radius from the TC center.
4) If there is a TC at the next time step (6 h) within a distance of 400 km, it is considered to belong to the same trajectory as the initial TC. If the storm lasts more than 2 days and has at least a maximum 10-m surface wind speed (interpolated from lowest model level at 22 m) greater than 17 m s$^{-1}$ on two consecutive days, the TC track is established. If the maximum intensity exceeds 33 m s$^{-1}$ at any point in the lifetime, it is classified as a TY; otherwise, it is classified as a TS.

### 3. Climatology

Table 1 presents the mean statistics, including TC genesis frequency, genesis location, intensity, lifetime, translation speed, maximum wind, and lifetime maximum wind from observations and simulations. We first compare climatological features of TCs between observations. The TC intensity (ACE and lifetime maximum wind) is greater in the JTWC than in the IBTrACS, largely due to the different averaged times (1 versus 10 min) in defining TC intensity and the initial and ending points of the TC track for computing lifetime ACE. Overall, the two observed datasets are in good agreement except for intense storms with winds exceeding 60 m s$^{-1}$, which are not simulated in the model in any case. To compare model results, both best-track datasets, the JTWC and IBTrACS, are used, but only the JTWC results are emphasized here.

<table>
<thead>
<tr>
<th>JTWC</th>
<th>IBTrACS</th>
<th>ZETAC</th>
<th>ZETAC-warm</th>
<th>Change (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Genesis frequency of TC (yr$^{-1}$)</td>
<td>22.3</td>
<td>22.2</td>
<td>22.2</td>
<td>20.7</td>
</tr>
<tr>
<td>Genesis frequency of TY (yr$^{-1}$)</td>
<td>14.7</td>
<td>13.7</td>
<td>18.4</td>
<td>18.4</td>
</tr>
<tr>
<td>Genesis longitude of TC (°E)</td>
<td>137.1</td>
<td>137.9</td>
<td>137.7</td>
<td>138.1</td>
</tr>
<tr>
<td>Genesis latitude of TC (°N)</td>
<td>16.7</td>
<td>17.0</td>
<td>13.7</td>
<td>14.4</td>
</tr>
<tr>
<td>ACE (10$^4$ m$^2$ s$^{-2}$)</td>
<td>71.7</td>
<td>(51.8)</td>
<td>65.0</td>
<td>64.7</td>
</tr>
<tr>
<td>Lifetime (day)</td>
<td>6.2 (5.5)</td>
<td>7.3 (5.5)</td>
<td>8.6 (7.4)</td>
<td>8.7 (7.5)</td>
</tr>
<tr>
<td>Velocity of TC motion (m s$^{-1}$)</td>
<td>6.0 (6.0)</td>
<td>7.1 (6.0)</td>
<td>6.1 (6.0)</td>
<td>6.3 (6.3)</td>
</tr>
<tr>
<td>Lifetime-mean max. wind (m s$^{-1}$)</td>
<td>27.2</td>
<td>(30.7)</td>
<td>27.8</td>
<td>28.2</td>
</tr>
<tr>
<td>Lifetime maximum wind (m s$^{-1}$)</td>
<td>43.3</td>
<td>35.9</td>
<td>39.0</td>
<td>40.0</td>
</tr>
</tbody>
</table>
The simulated TCs are generally similar to the observations, but with more TY genesis, longer TC lifetime, weak averaged TC intensity (ACE and lifetime maximum wind), and an equatorward bias in mean TC genesis location (Table 1). However, perhaps the most notable difference is the underestimation of the frequency of intense TCs (Fig. 1), which, along with an overestimation of moderate-intensity TC frequency, is consistent with the previous study in the Atlantic (Knutson et al. 2007). Maximum surface winds of about 51 m s$^{-1}$ are the most intense TC simulated in the model, compared to about 82 m s$^{-1}$ in the observations. Correspondingly, the minimum surface pressure at the center of TC in the model is only about 910 hPa, not as low as the observed central pressure of 879 hPa. TCs with wind speeds between 30 and 50 m s$^{-1}$ are much more frequent in the model than in the observations. This could induce some discrepancies between the simulated and the observed TC intensity-related measures, particularly for those indices that are strongly influenced by intense TYs, such as ACE and TIF.

Figure 2 shows the climatology of the spatial distribution for annual TC genesis and occurrence density, along with ACE and TIF, from the model and the observations for 1980–2008. Compared to the observations, the simulated TC genesis is fairly realistic, but is shifted slightly eastward and southward. The climatological mean TC genesis position in the model, near 13.7$^\circ$N, 137.7$^\circ$E, is about 3$^\circ$ in latitude away from the observed mean position (Table 1). The TC occurrence pattern is also captured by the model, but is somewhat higher in the southeastern part of the domain. While the magnitude of the simulated ACE and TIF is slightly lower than observed (although with compensating errors as discussed above), the spatial distributions of the ACE and TIF in the model are fairly realistic (Figs. 2c,d).

The simulated TC number is 22.2 per year, which is consistent with the observations (Table 1). The seasonal cycle of TC number is also captured by the model, with the peak in August (Figs. 3a,b). However, the model
overestimates the genesis frequency of the TYs at 125% of the observed. The seasonal cycle of the ACE is also roughly captured by the model, but is overestimated in magnitude in July–August and underestimated in September–November (Figs. 3c,d). Although the annual ACE is underestimated by the model, the percentage of the ACE for TYs relative to the total ACE of TCs is reasonable, compared to the observation. Figure 4 shows the monthly mean climatological TC occurrence and 850-hPa wind vectors during June–November for the simulation and observations, respectively. The simulated large-scale circulation over the WNP, which is very similar to the observation, features a strong seasonality in the axis of the monsoon trough from June to November that has also been found in observations (Chia and Ropelewski 2002; Atkinson 1971). It is noted that the simulated TC occurrence is closely related to the monsoon trough location. From June to September when the monsoon trough is extended northward, TC genesis and occurrence also tend to shift toward the north. In October and November, the southward migration of TC activity is coupled with the monsoon trough’s southward retreat. This meridional shift in the location of the simulated TC activity, associated with the shifts of the monsoon trough, agrees well with the observations.

4. Interannual variations

a. Interannual variations of TC activity

The interannual variability of the TC genesis number in the model simulation and the observation is shown as a time series in Fig. 5a. Even though the means and trends are similar, no statistically significant correlation is found between the model simulation and the observations. The lack of skill in reproducing the year-to-year variation of TC number in the WNP, compared to the Atlantic basin (Knutson et al. 2007), could be largely due to the weak relationship between ENSO and basinwide TC numbers in the WNP observations (their correlation coefficient $r$ is 0.17, which is not statistically significant). Figures 5b–d show the time series for the TC occurrence, ACE, and TIF. The simulated long-term

---

![Figure 3](http://journals.ametsoc.org/doi/pdf/10.1175/JCLI-D-12-00830.1)
trends are similar to the observation, although a clear positive bias is found in TC occurrence and a slightly negative bias is found in ACE and TIF. The indices of TC activity shown in Figs. 5b–d clearly show a much better agreement between the model simulation and the observation than the TC number shown in Fig. 5a, but the correlations are still not statistically significant, except the TIF. The model has much lower skill in reproducing the interannual variability of basinwide TC activity over the WNP than in the Atlantic (Knutson...
et al. 2007). However, we will show that the model has some skill at simulating the interannual variability of TC activity in terms of spatial shifts of activity between different subregions over the WNP.

Figure 6 shows the interannual variability of the location of the mean TC genesis for the model versus observations. The year-to-year variability in the zonal variation of TC genesis is well reproduced by the model (Fig. 6a). The long-term trend of westward shifting is also well simulated. The correlation coefficient is 0.67, exceeding the 99% confidence level. The interannual variation of the displacement in latitude for the simulated TC genesis is also significantly correlated ($r = 0.59$) with that for the observation, although it has a clear negative bias, and the model trend of a gradual southward shift is not in agreement with the gradual northward shift in the observations (Fig. 6b). Overall, the model is fairly skillful in simulating the interannual variation of the spatial distribution of TC activity over the WNP.

Previous studies have noted that the life span, intensity, and genesis location of TCs are intimately interrelated, and the interannual variation of the spatial distribution of TC activity has been attributed to shifts in the location of the TC genesis (Camargo and Sobel 2005; Camargo et al. 2007; Wang and Chan 2002; Wu et al. 2012). In addition, the model shows a fairly realistic simulation of the spatial distributions of standard deviation for TC genesis, occurrence, ACE, and TIF (not shown). These results imply that the model has a better skill in simulating the interannual variation of the spatial distribution of TC activity over the WNP than in basinwide aggregate statistics such as TC counts.

To further examine the spatiotemporal characteristics of TC activity in the model, an EOF analysis and the TIF index are used to characterize the major modes of interannual variability of TC activity in the model and observations. The two leading EOF modes of the simulated TIF are shown in Figs. 7a and 7b. They explain 41.6% and 10.4% of the total variance, respectively. The spatial distribution and magnitude of the first EOF mode (EOF1) of TIF for the model (Fig. 7a) has strong positive weightings for TIF values in most of the basin and a pronounced maximum at $15^\circ$N and $135^\circ$E. These are well-represented features in comparison to the observations (Fig. 7c), except for the South China Sea (SCS). The correlation coefficient between the principal component of the first EOF mode (PC1) of the simulation and the observation is 0.37, which exceeds the 95% confidence level. The spatial pattern of simulated second EOF mode (EOF2; Fig. 7b) shows positive values in most parts of the WNP north of
20°N and negative values in eastern part of the WNP south of 20°N. Although the model does not reproduce the observed long-term trend of the principal component of the second EOF mode (PC2), the correlation coefficient, 0.32, still exceeds the 90% confidence level. This implies that the model can reproduce the observed spatiotemporal change of TC activity in the WNP to a considerable extent. An exception occurs in the northern SCS, where the negative values of EOF1 and EOF2 in the observation are not found in the model simulation.

To further assess the model skill for present-day TC activity, Fig. 8 shows the tracks of all TCs in 1997 (1999) when the corresponding PC1 value was extremely positive (negative) and in 2002 (1983) when the corresponding PC2 value was extremely positive (negative). Comparing 1997 versus 1999, we find that TCs formed deeper in the tropics and more eastward in 1997, leading to longer typhoon intensity tracks. A similar behavior is simulated in the model. Between 2002 and 1983, there were more observed TCs in 2002 and they tended to take
more recurving tracks. However, in the model these two years were more similar in their TC behavior than they were in observations. As observed by Wu et al. (2011), the first EOF mode is apparently related to ENSO, as well as to MT. To confirm this, we compared the simulated time series with the Niño-3.4 indices and MT indices, respectively. The corresponding correlation coefficients are 0.49 and 0.62, exceeding the 99% confidence level. This implied that the model may provide useful information for understanding the impact of the MT on TC activity in the WNP.

b. Monsoon trough and TC activity

As observed by Wu et al. (2012), the interannual variability of TC activity over the WNP is closely related to the monsoon trough variability. During years when the monsoon trough extends eastward (retreats westward), more (fewer) TCs form within the southeastern quadrant of the WNP, and there is more (less) TC activity east of the Philippines. Here we examine, through analyzing the simulated TCs, this linkage between the monsoon trough variability and TC genesis over the WNP during July–November for the period 1980–2008. Based on Wu et al. (2012)’s monsoon trough index, the strong monsoon trough is well synchronized with the eastward extension of the monsoon, while the weak monsoon trough corresponds to a westward retreat. The model S-MT years with the highest seven values of the monsoon trough index (MTI) include 1982, 1986, 1990, 1991, 1997, 2002, and 2004 and the model W-MT years.

Composites of the simulated 850-hPa winds for seven S-MT and seven W-MT years are displayed in Figs. 9a and 9b, respectively. The locations of simulated TC genesis for the corresponding years and the simulated precipitation are also superimposed. Some striking contrasts are simulated for 850-hPa winds and precipitation between S-MT and W-MT years, when the simulated monsoon trough extends eastward (Fig. 9a) and retreats westward (Fig. 9b). These variations of the simulated monsoon trough are consistent with those found in the observations (Figs. 9c, d). The locations of the simulated TC genesis, displaying notable contrasts between the S-MT and W-MT years, are generally similar to the observation. More TCs are generated east of 150°E in the S-MT years (Fig. 9a) than in the W-MT years (Fig. 9b).

To obtain a sense of the spatial distribution in the number of TC genesis, the aforementioned analysis domain is divided into five subregions: region SW (0°–15°N, 120°–150°E), region NW (15°–30°N, 120°–150°E), region SE (0°–15°N, 150°E–180°), region NE (15°–30°N, 150°E–180°), and region SCS (0°–25°N, 100°–120°E). These subregions are outlined in Fig. 9. The average number of simulated TCs in the S-MT and W-MT years in the five subregions is shown in Fig. 10. The frequency of formation of TCs in the S-MT years is enhanced in the SE region and suppressed in the NW and SCS regions (Figs. 10a, b), with more significant changes in the SE region. The TC genesis frequency in region SE during the S-MT years is more than 2 times greater than that during the W-MT years. In other words, the increase (decrease) of TC genesis in the S-MT (W-MT) years comes primarily from the change in the SE region (Figs. 9 and 10), which is mainly due to changes of intense TCs (TYs; Fig. 10). On the other hand, both the NW and SCS regions contribute to the increase (decrease) of TC genesis in the W-MT (S-MT) years. The change in weaker TCs and TSs (Fig. 10b) played a more important role in these two regions. Despite the model bias in producing the number of weaker TCs and TSs, these contrasts correspond reasonably well to those found in the observations (Fig. 10).

The observations shown in Wu et al. (2012) suggest that the spatial pattern of TC activity is closely related to the longitudinal migration of MT. Figure 11a shows the difference of the simulated TC occurrence between the S-MT and W-MT years. The clear east–west pattern, positive differences east of the Philippines and negative differences over the SCS, is in an agreement with observations (Fig. 11c). With the increase of the northwestward tracks (during the S-MT years), TCs start in the south-east quadrant of the WNP and tend to recurve from northwestward to northeastward around 25°N and 130°E. The reduced occurrence of TCs over the SCS during the S-MT years is consistent with a previous study (Wu et al. 2012), but is shifted more eastward and southward than the observation. To further examine the spatial distribution of the simulated TC activity, the difference in ACE between the S-MT and W-MT years is also shown in Fig. 11b. These differences are clearly similar to those exhibited in the simulated TC occurrence anomalies, consistent with observations (Fig. 11d).
The average ACE per year in the S-MT (W-MT) years tends to increase (decrease), and the differences between the S-MT and W-MT years are statistically significant for both observations and simulations.

The above analyses reveal that the model is able to reproduce the shift in the spatial distribution of observed TC activity between the S-MT and W-MT years. To this extent, it shows some skill in reproducing the interannual variability of the spatial distribution of TC activity over the WNP.

5. Projected changes in tropical cyclone activity

Figures 12–14 show simulated changes in the large-scale environment induced by global warming, which are derived (via interior spectral nudging in the

![Fig. 10. Mean number of simulated (a) TY, and (b) TS genesis per year over each subregion outlined in Fig. 9 for strong (S-MT, red), neutral (N-MT, white), and weak monsoon trough (W-MT, blue) years. (c),(d) As in (a),(b), but for the observations.](image)

![Fig. 11. Difference between anomalies in S-MT and W-MT years of the (a) total number of model TC occurrences and (b) ACE in each 2.5° × 2.5° grid. Unit for number of TCs is yr⁻¹ and for ACE is 10⁴ m² s⁻² yr⁻¹. (c),(d) As in (a),(b), but for the observations.](image)
atmosphere) from the CMIP3 ensembles for the A1B emission scenario (late twenty-first century) and correspond to those for the Atlantic basin discussed by Knutson et al. (2008) (see the detailed description in section 2c). Over the WNP, specified SST changes from the CMIP3 are warmer everywhere in the late twenty-first century than in the control (Fig. 12a). The vertical profiles of temperature and moisture changes averaged over three different boxes are shown in Figs. 12b and 12c. It is clear that the upper troposphere warms up faster than the lower troposphere, increasing the dry static stability, which is consistent with previous studies (Chou et al. 2013; Knutson and Tuleya 1999). The vertical wind shear (Fig. 13a) is slightly reduced to the east of the TC main development region (blue box) with virtually no change toward the warm pool and SCS (red box). The midtropospheric relative humidity increases slightly over the whole domain (Fig. 13b). While the upper-level divergence shifts upward over most of the tropical WNP (Fig. 13c), the low-level vorticity increases (decreases) over the eastern (western) part of domain (Fig. 13c). A small eastward shift of the Pacific subtropical high and positive vorticity anomalies near the western edge of the subtropical high in the warmer climate are also apparent, along with changes in the mid-level steering flow (Fig. 14a). These changes, derived from the CMIP3 climate models’ ensemble mean projections, determine the late-twenty-first-century large-scale environmental conditions for TC genesis and development in the regional model.

### a. Impact of global warming on TC activity

The projected changes of WNP TC activity are summarized in Table 1. The average annual (June–November) number of TC genesis events over the WNP is 20.7 in the global warming simulation, which represents a reduction of 7%, compared to the present-day simulation. Although TC numbers decrease, the corresponding ACE index shows no significant changes under global warming, implying that the individual TC events have longer lifetimes and/or higher intensity in the
projections (average ACE of individual events increases 5.6%). The model projects enhancements of the mean TC intensity for both lifetime-mean maximum wind speed (0.4 m s$^{-1}$, 1.4% increase) and lifetime-maximum wind speed (1 m s$^{-1}$, 2.6% increase). The average number of typhoons is 18.4 in the global warming simulation—about the same in the present-day simulation. This implies a 6.0% increase in the percentage of the TY number with respect to the total TC number in the future projection. Moreover, TYs with lifetime maximum wind speeds exceeding 45 m s$^{-1}$ occur significantly more frequently in the warm climate simulation (Fig. 1). Meanwhile, the storm lifetime and mean translation speed of TCs are respectively 8.7 days and 6.3 m s$^{-1}$ in the global warming simulation, representing slight increases in TC lifetimes (1.2%) and translation speed (3.3%) compared to the control. The mean TC genesis location shifts slightly eastward and northward in the WNP (Table 1). Furthermore, the seasonal variation in projected TC number and ACE is similar to that in the present-day simulation, indicating no apparent changes in the seasonal cycle (not shown). None of the changes we examined in TC activity averaged over the entire WNP is statistically significant at the 90% level (Table 1) given the strong subregional spatial variations of these changes, such as shown in Fig. 15.

Figure 15 shows the spatial distributions of projected changes (CMIP3, A1B scenario) in the TC genesis, TC occurrence, ACE, and TIF. In the future warmer climate, a decrease in the number of TC genesis events occurs between 140° and 150°E, while slightly more TCs formed to the east of this region, resulting in a slight (0.4° longitude) eastward shift of the mean genesis location (Fig. 15a; Table 1). However, the regional patterns of change of various TC metrics shown in Fig. 15 are rather noisy with regions of increase and decrease interspersed around the domain. Moreover, since the fraction of grid points with local significance at the 0.20 level (80% on the map) is only about 10%, the structure of changes shown here is not regarded as statistically robust. Therefore we do not focus any further on the regional details.

Another potentially interesting change is the projected poleward shift in TC activity (ACE) occurring north of Taiwan. Such a change would imply an increase in TCs making landfall in north China, the Korean Peninsula, and parts of Japan. The regional changes in the ACE are summarized for two subregions in Fig. 15e. These show a substantial change in regional variation.
between the areas, with a nearly 9% decrease over the region south of China (region S: 7.5°–17.5°N, 115°–160°E) and an increase of nearly 10% over the subregion north of 30°N (region N: 30°–40°N, 125°–170°E). These changes are only statistically significant at the 70% level, however. These results confirm that the TC activity (ACE) shows only a weak tendency for a poleward shift in our experiments. Thus, while there is a hint of a northward shift of TC activity reminiscent of the behavior seen in observations since 2000 (Tu et al. 2009; Wu et al. 2011), our model does not simulate a robust or statistically significant (90%) northward shift in activity for the late twenty-first century.

b. Reasons for future change in TC activity

Despite the finding that the projected changes in TC genesis in our climate change experiments are weak and only significant at the 80% level, it is of interest to further examine the simulated TC genesis and track changes using published statistical indices of TC genesis. The CMIP3 A1B scenarios show substantial impacts on large-scale environment conditions that are considered to influence the TC formation and development. To further probe these changes in large-scale environmental conditions thought to be most relevant to TC genesis, we used a variant of Emanuel and Nolan’s (2004) genesis potential index (GPI) as modified by Murakami and Wang (2010):

\[
\text{GPI} = 10^5 \eta^{3/2} \left( \frac{\text{RH}}{50} \right)^3 \left( \frac{\text{MPI}}{70} \right)^3 \times (1 + 0.1VWS)^{-2} \left( \frac{-\omega + 0.1}{0.1} \right),
\]

where \( \eta \) is the 850-hPa absolute vorticity, \( \text{RH} \) is the 700-hPa relative humidity, \( \text{MPI} \) is the maximum potential intensity (Bister and Emanuel 1998), \( VWS \) is the 200–850-hPa vertical wind shear, and \( \omega \) is the 500-hPa vertical wind velocity. Figure 16 shows the spatial distributions of future change in the total GPI, and area mean of GPI changes obtained by varying several factors in the equation individually. The results show a predominant projected increase in the GPI over the WNP, while the projected GPI weakens slightly in...
a relatively small region east of the Philippines. Notably, the total GPI changes are inconsistent with the model-projected mixture of increases and decreases in TC genesis around the basin (Fig. 15a). Discrepancies between the results of direct simulations of TC changes versus changes inferred indirectly from the GPI changes have also been pointed out by previous studies (e.g., Caron and Jones 2008; Emanuel 2013; Yokoi and Takayabu 2009). To further explore the GPI changes, the changes obtained by varying each individual GPI element are shown in Fig. 16b. Increased MPI, which is a theoretical expression for maximum potential TC intensity for a given set of thermodynamic conditions, appears to be the primary contributor to the projected GPI increase. Based on our dynamical downscaling results, the projected MPI increases do not lead to a predominantly positive change in genesis frequency across the basin, and thus the MPI contribution to GPI appears to be a major source of discrepancy between our dynamical projections and our GPI-based projections of TC genesis. On the other hand, the vertical motion ($\omega$) term contributes to a decrease in GPI over the WNP, suggesting that changes in mid-troposphere vertical ascent associated with convective may contribute toward the slightly reduced TC genesis activity over the WNP as a whole. While the relative humidity, vorticity, and vertical wind shear show only small contributions toward the projected general increase of GPI, the relative humidity and vorticity are the most important contributors to the interannual variability of the genesis location in the WNP (Camargo et al. 2007). Emanuel et al. (2008) and Emanuel (2010) also noticed that the relative humidity term in GPI tends to be of lesser importance to GPI changes for other climate states. Although they seem to be of secondary importance in terms of GPI, the changes in the dynamic factors (vorticity and VWS) seem to be related to the distributions of the future change in the TC genesis simulated by the model (Figs. 15a and 16b). East of 150$^\circ$E (west of 140$^\circ$E), weaker (stronger) vertical wind shear and positive (negative) low-level vorticity anomalies both favor (disfavor)}
enhanced TC activity (Figs. 13a,c). In addition, the decreased upper-level (200 hPa) divergence in the tropical WNP would tend to reduce the possibility of development of deep convection, and thus is an unfavorable factor for TC activity (Fig. 13c). Those results suggest that dynamic variables are relatively more important (compared to MPI) for the changes in formation of TCs in a warmer climate than is currently estimated by the GPI formula. However, such a conclusion would need to be verified with future work by comparing the relative importance of key dynamic and thermodynamic variables in determining the TC formation in a variety of climate settings.

Another possibility is that the changes in TC genesis under climate change may be related to changes in synoptic-scale disturbances that can trigger individual cyclogenesis (Wu et al. 2012, 2013). The existence of a larger number of synoptic-scale disturbances is thought to be an important forcing mechanism for tropical cyclogenesis over the WNP. These synoptic-scale disturbances often associated in the WNP with westward-propagating wave disturbances in the easterlies, such as mixed Rossby–gravity (MRG) waves and tropical depression (TD)-type disturbances. Here, we examine the disturbance activity is analyzing the variance of 2–8-day filtered eddy kinetic energy (EKE) at 850 hPa, following Wu et al. (2012). Maximum EKE in the lower troposphere is observed around the monsoon trough region in the present climate (Fig. 14b). With the future eastward extension of the MT (not shown), the location of northward propagating disturbances tends to shift eastward along the monsoon trough along with the maximum EKE. Projected decreases of low-level synoptic-scale disturbances activity cover much of the western part of the tropical Pacific whereas relatively small changes are projected for the eastern tropical Pacific (Fig. 14b). In the same large-scale background conditions, the weak disturbance activity may suppress genesis by forcing precursor perturbations to descend. The reduced low-level synoptic-scale disturbances in the western part of the basin may thus be a factor contributing to the areas of reduced TC genesis in that region.

As shown in Fig. 14a, the western Pacific subtropical high is projected to retreat eastward slightly in the global warming simulation, with increased 500-hPa mean relative vorticity in the SCS and east of Taiwan. This change in 500-hPa steering flows would allow for more TCs to move northwestward in the western part of the basin and eventually recurve in higher latitudes. This may produce some of the increase in TC activity (ACE) in higher latitudes (30°–40°N) in the warm climate projections (Figs. 15c,e).

While the projected changes in TC activity (e.g., ACE in Figs. 15c,e) seem to be related to the projected circulation changes and associated shift in tracks, the regional details of the change are generally not statistically robust overall, indicating that the projected changes are relatively small compared with the internal variability within the model integrations.

6. Conclusions

A simulation of present-day (1980–2008) and projected (late twenty-first century; CMIP3 A1B scenario) WNP TC activity was conducted using an 18-km-grid regional atmospheric model. The simulated climatology and seasonal cycle of TC activity for the present climate show many characteristics similar to observations, such as the annual basinwide TC number, TC tracks (occurrence), and TC genesis location. The spatial distributions of the climatological mean observed TC tracks, occurrence, and ACE were fairly well captured in the present-day simulation although systematic biases were found in the genesis location and in the storm intensity distribution. The model simulates too few weak storms and does not simulate any TCs with winds exceeding 55 m s⁻¹. Thus intensity change projections from the model should be viewed with caution. The model shows some skill in simulating the meridional and zonal variations from year to year in TC genesis location, although it was not able to capture the observed interannual variation of total TC number. Regarding the spatial and
temporal variations of TIF, the model simulates similar patterns for the first and second EOFs, compared to observations. Furthermore, the relationship between the monsoon trough and the interannual variability of WNP TC activity in the model is in good agreement with observations, suggesting that the model has skill in simulating the influence of large-scale circulation changes on the spatial interannual variability of TC activity over the WNP.

In the late twenty-first century global warming projections, the large-scale environment over the WNP is projected to have significant changes, such as warmer SST, weaker vertical wind shear over most of the area, increased midlevel relative humidity, and changes to the steering flow. In response to the changes in the large-scale environment and synoptic-scale disturbances, the model’s TC activity showed relatively small changes: the basinwide TC genesis count decreased by 7%, but the average ACE of individual TC events increased by 5.6% and the average lifetime-maximum wind speed increases by 2.6%. The changes in genesis count and intensity were only marginally significant (80% level). There is also a hint of a poleward shift in WNP TC activity under global warming in association with changes in the large-scale steering flow, but again this is not statistically significant at the 90% level. A regional feature of the simulations is a projected increase in TC activity north of Taiwan, which would imply an increase in TCs making landfall in north China, the Korean Peninsula, and Japan. While this future projection is reminiscent of the systematic shifts in the prevailing TC tracks observed over the period 1965–2003 (Wu et al. 2005; Wu et al. 2011), since the percentage of areas with significant changes is not greater than expected by chance, we do not regard such a regional feature as statistically robust in our simulations. In summary, the projected TC changes are rather small compared with the model’s internal variability, although the intensity projections need to be revisited using a higher-resolution model owing to our regional model’s limitation at simulating intense TCs.

Acknowledgments. This work was supported by the Chinese Academy of Sciences Grant KZCX2-EW-QN204, the National Basic Research Program of China under Grant 2014CB953902, the National Natural Science Foundation Grants 41205052, 41230527, and 41375065, and the National Science Council Grants NSC98-2745-M-001-005-MY3, NSC 99-2111-M-003-002, and NSC100-2119-M-003-005-MY5. We thank the WCRP CMIP3 modeling groups for contributing their model runs, and PCMDI and the IPCC Data Archive at LLNL/DOE for providing access to model data; and NCEP, NCAR, and NOAA/ESRL for their efforts in developing and distributing the atmospheric reanalysis data used in the study. We are grateful to the National Center for High-Performance Computing for computer time and facilities.

REFERENCES


——, 2000: Tropical cyclone activity over the western North Pacific associated with El Niño and La Niña events. J. Climate,
1M AY 2014 W U E T A L. 3423


