Tropical Cyclone Changes in the Western North Pacific in a Global Warming Scenario

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ABSTRACT

The influence of global warming on the climatology of tropical cyclones in the western North Pacific basin is examined using the high-resolution International Pacific Research Center (IPRC) regional climate model forced by ocean temperatures and horizontal boundary fields taken from the NCAR Community Climate System Model version 2 (CCSM2) coupled global climate model. The regional model is first tested in 10 yr of simulation with boundary forcing taken from observations and is shown to produce a reasonably good representation of the observed statistics of tropical cyclone numbers and locations. The model was then run for 10 yr with forcing from a present-day control run of the CCSM2 and then for 10 yr with forcing fields taken from the end of a long run with 6 times the present-day atmospheric CO₂ concentration. The global-mean surface air temperature warming in the perturbed run is 4.5 K, while the surface warming in the tropical western North Pacific is about 3 K. The results of these experiments reveal no statistically significant change in basinwide tropical cyclone numbers in the peak season from July to October in response to the CO₂ increase. However, a pronounced and statistically significant increase in tropical cyclone occurrence in the South China Sea is found. While the basinwide total number of storms remains nearly unchanged in the warm climate, there is a statistically significant increase in the average strength of the cyclones and in the number of the storms in the strongest wind categories.

1. Introduction

There is great interest in the question of how the statistics of tropical cyclone (TC) number, intensity, and location may change in response to large-scale climate forcing such as increases in greenhouse gas concentrations. Given the considerable impact of tropical cyclones on society, a reliable forecast of the response of the tropical cyclone climatology to a prescribed global forcing scenario would be of great practical value. Another motivation for studying the effect of large-scale climate change on tropical cyclones is provided by the work of Emanuel (2001). He proposed that warming tropical climate should lead to stronger tropical cyclones, and that this should enhance the wind-induced oceanic mixing. In turn this should drive an enhanced oceanic heat transport out of the Tropics, providing a potentially significant negative feedback helping to limit climate variations in the Tropics. The importance of this mechanism will depend, however, on how strongly the integrated global strength of the cyclones scales with tropical surface temperatures. There are theoretical reasons for believing that tropical cyclones should have higher potential intensities in a warmer mean climate (Emanuel 1987; Holland 1997), but determining the dependence of the actual realized tropical cyclone intensities (as well as numbers and locations) on the global climate state must rely on more detailed modeling or observational approaches.

Observational studies have compared the observed trends in tropical cyclone statistics to the trends in tropical surface temperature over the past several decades. Emanuel (2005) used an index of potential destructiveness that is based on the total dissipation of power, integrated over the lifetimes and storm intensities. His study concluded that this index is highly correlated with tropical sea surface temperature. Webster et al. (2005) examined the number of tropical cyclones and cyclone days as well as storm intensity over the past 35 yr. They saw a large increase in the number and proportion of hurricanes reaching categories 4 and 5. The largest increase occurred in the North Pacific, Indian, and southwest Pacific Oceans, and the smallest
percentage increase occurred in the North Atlantic Ocean. These conclusions are still somewhat controversial as it may be difficult to separate sustained, long-term trends in tropical cyclone activity and natural low-frequency variability in this activity. For example, Chan and Liu (2004) report finding no significant trend in tropical cyclone activity in the western North Pacific (WNP) in the last 40 yr, despite the surface warming that has been observed.

In principle, a forecast of the response of tropical cyclone statistics to climate forcing could be obtained through analysis of the results of a comprehensive ocean–atmosphere global climate model (GCM) as it is integrated through a particular climate scenario. There have been a number of published studies of this sort all focused on the tropical cyclone response to a doubling of CO$_2$ concentration (Broccoli and Manabe 1990; Bengtsson et al. 1996; Sugi et al. 2002; Tsutsui 2002). In each case the surface temperature warms significantly in the doubled-CO$_2$ integration, but the various models display somewhat different responses in the tropical cyclone climatology. One problem with such coupled GCM studies is the practical limitation to fairly coarse numerical resolution for the long integrations required. Broccoli and Manabe (1990) and Tsutsui (2002) used models with atmospheric components with effective horizontal grid spacing of about 300 km. Bengtsson et al. (1996) and Sugi et al. (2002) employed GCMs with atmospheric components with effective horizontal grid spacing of about 100 km. Experience has shown that while such coarse models may be able to reproduce some aspects of the observed tropical cyclone climatology, they are unable to simulate the most intense storms observed in the real atmosphere. Another approach that can be used to study storms in a particular region is to embed a limited-area regional climate model in a coarser-resolution GCM. This approach was adopted by Nguyen and Walsh (2001) to study effects of global warming on tropical cyclones in the Australian sector of the South Pacific and Indian Oceans. However, the regional climate model employed in their study was itself only of moderate resolution, with about 125-km horizontal grid spacing and only nine levels in the vertical.

Knutson et al. (1998), Walsh and Ryan (2000), and Knutson and Tuleya (2004) adopted somewhat different approaches in order to more adequately resolve model tropical cyclones and thus study the effect of global warming on the development of even quite intense hurricanes. In particular Knutson et al. (1998) and Knutson and Tuleya (2004) examined the simulated tropical cyclones in coarse-resolution GCMs. They traced each tropical cyclone identified back to its genesis and early development stage. A multiply nested limited-area storm model run was then initiated with a more robust vortex inserted at the location of the GCM tropical cyclone, and environmental conditions (and SSTs) taken from the GCM run. The individual limited-area model integrations continued for 5 days, and the maximum intensity of the simulated cyclone during that period was determined. Knutson et al. (1998) used a version of the limited-area storm model with 18-km horizontal grid spacing in the innermost mesh, while Knutson and Tuleya (2004) had 9-km resolution in the innermost mesh of their model. For each of the control and doubled-CO$_2$ cases, Knutson et al. (1998) considered 51 storms in the western North Pacific region. Knutson and Tuleya (2004) considered results starting from several different global GCMs and studied cases in the western and eastern North Pacific and in the Atlantic. Walsh and Ryan (2000) adapted this method but used different hurricane and large-scale models. In particular, they employed the simulation of the Nguyen and Walsh (2001) regional climate model for the Australian sector as their starting point.

In the present study we use a high-resolution regional climate model that is driven by the output of a coarse-resolution GCM to assess the influence of global warming on the frequency and intensity of tropical cyclones in the western North Pacific. The regional climate model used is run at 1/2° horizontal resolution and with 28 levels in the vertical. The regional model differencing scheme is fourth order in the horizontal, which should lead to a more accurate treatment of small scales than for the more commonly employed second-order differencing schemes in gridpoint models. We will show that, when our western North Pacific regional model is driven by observed horizontal boundary conditions and SSTs, it simulates a rather realistic climatology of the cyclone numbers, locations, and intensities, although the model does not develop the most extremely intense storms that are observed. We use the model to forecast the response to a sixfold increase in atmospheric CO$_2$ concentration. This large climate forcing was adopted in order to produce a strong climate change–related signal that may stand out among the unforced interannual variability in tropical cyclone statistics.

Our experiment is similar in conception to one recently published (Oouchi et al. 2006). They analyzed the climatology of tropical cyclones in a 10-yr control and a 10-yr time slice warming simulation performed with an atmospheric global model in which the SSTs were taken from comparable periods of a low-resolution coupled GCM global warming experiment. Their integrations employ a very high resolution (T959)
atmospheric model (e.g., Mizuta et al. 2005). It is interesting to note that, despite the fine resolution employed, their control run displays a serious underprediction of the frequency of tropical cyclones in the western North Pacific. Oouchi et al. (2006) do not provide an explanation for this deficiency. However, they do note that it is consistent with a bias of unrealistically low rainfall in the tropical and subtropical western North Pacific, a bias shared by many current GCMs. As we will show later, our somewhat coarser-resolution regional model is able to simulate a much more realistic present-day climatology of the tropical cyclone distribution in this region.

The remainder of this paper is organized as follows. In section 2 the models and experiments are described briefly. The results are given in section 3 followed by further discussion in section 4. Our main conclusions are summarized in the last section.

2. Models, experiments, and observational data

a. Models

The numerical study is carried out using the high-resolution regional climate model developed at the International Pacific Research Center (IPRC-RegCM). The IPRC-RegCM uses hydrostatic, primitive equations in spherical coordinates with sigma (pressure normalized by surface pressure) as the vertical coordinate (Wang et al. 2003). The model equations are solved with a fourth-order conservative horizontal finite-difference scheme on an unstaggered longitude–latitude grid system. The time integration is performed using a leapfrog scheme with intermittent application of an Euler backward scheme. The model physics includes the cloud microphysics scheme of Wang (2001); a mass flux scheme for subgrid shallow convection, midlevel convection, and deep convection developed by Tiedtke (1989) with some modifications outlined in Wang et al. (2003); the radiation package developed by Edwards (1989) with some modifications outlined in Wang et al. (2003, 2004). The model equations are solved using the high-resolution regional climate model developed at the International Pacific Research Center (IPRC-RegCM). The IPRC-RegCM uses hydrostatic, primitive equations in spherical coordinates with sigma (pressure normalized by surface pressure) as the vertical coordinate (Wang et al. 2003). The model equations are solved with a fourth-order conservative horizontal finite-difference scheme on an unstaggered longitude–latitude grid system. The time integration is performed using a leapfrog scheme with intermittent application of an Euler backward scheme. The model physics includes the cloud microphysics scheme of Wang (2001); a mass flux scheme for subgrid shallow convection, midlevel convection, and deep convection developed by Tiedtke (1989) with some modifications outlined in Wang et al. (2003); the radiation package developed by Edwards and Slingo (1996) and further improved by Sun and Rikus (1999); the Biosphere–Atmosphere–Transfer Scheme (BATS) developed by Dickinson et al. (1993) for land surface processes; a modified Monin–Obukhov similarity scheme for flux calculations at the ocean surface; and a nonlocal $E-e$ turbulence closure scheme for subgrid-scale vertical mixing (Langland and Liou 1996), which was modified to include the effect of cloud buoyancy production of turbulence kinetic energy (Wang 1999). A one-way nesting is used to update the model time integration in a buffer zone near the lateral boundaries within which the model prognostic variables are nudged to reanalysis data or the output of the global climate model with an exponential nudging coefficient proposed by Giorgi et al. (1993) and later modified by Liang et al. (2001). The buffer zone in this experimental setup is 5° in extent. More details of the model can be found in Wang et al. (2003, 2004).

The version of the IPRC-RegCM used in this study has 28 vertical levels with high resolution in the planetary boundary layer. The lowest model level is roughly 25 m above the surface. The model domain extends from 15°S to 55°N, 100°E to 160°W with a grid spacing of 0.5°, in both zonal and meridional directions. This resolution is still too low to represent the structure of real tropical cyclones, so the model is unlikely to simulate all aspects of tropical cyclone structure. However, previous experience with global models suggests that models with the ~50-km grid spacing employed here can spontaneously simulate a reasonable climate of tropical cyclones, although misrepresenting to some extent the intensities of the strongest storms. In the multiyear control simulations using global models with 300-km grid spacing described by Broccoli and Manabe (1990) and Tsutsui (2002), the deepest central surface pressures in the tropical cyclones were about 980 hPa. In the control simulation using a global model with ~100-km effective grid spacing reported by Bengtsson et al. (1995), the most intense western North Pacific tropical cyclone appearing had a minimum central pressure of 953 hPa and peak surface winds of ~45 m s$^{-1}$. The peak surface winds of less than ~50 m s$^{-1}$ are also apparent in the 10-yr control run with a model with ~100-km effective grid spacing described by Sugi et al. (2002). Hamilton and Hemler (1997) described results from a single season of control integration with a global gridpoint atmospheric model with spacing of about 35 km. They reported one western North Pacific typhoon with minimum pressure of 906 hPa and peak winds in the lowest model level of about 70 m s$^{-1}$, comparable to the strongest North Pacific typhoon that might typically be observed in a given year, but still weaker than the strongest typhoon ever observed (Typhoon Tip in 1979, which had an estimated central pressure as low as 870 hPa).

The simulations are forced by the output of the National Center for Atmospheric Research (NCAR) Community Climate System Model version 2 (CCSM2). Detailed descriptions of CCSM2 are given in Boville and Gent (1998) and Kiehl and Gent (2004). A number of climate change integrations, involving a variety of scenarios, have been documented for versions of the NCAR CSM (Boville and Gent 1998; Meehl et al. 2000; Boville et al. 2001; Dai et al. 2001).

The CCSM integrations reported here were performed locally at IPRC and were begun from the standard initial conditions provided by NCAR from their
control runs. The simulations were performed at T31 resolution, with 26 levels in the vertical. A 10-yr control simulation was performed, followed by a 50-yr simulation in which the atmosphere CO$_2$ concentration was changed to 6 times the present-day value.

**b. Experiments**

This study analyzes results from three experimental setups that were conducted with the IPRC-RegCM. All runs were performed for the peak season of tropical cyclones in the western North Pacific, that is, from July through October (JASO).

1) The present climate (PCL) is represented by 10 JASO periods of integration for 1991–2000 with the IPRC-RegCM driven by reanalysis data. The 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) global data (Simmons and Gibson 2000) at 6-h intervals with a resolution of 2.5° × 2.5° in the horizontal and 15 pressure levels up to 10 hPa were used to define the driving fields, which provide both initial and lateral boundary conditions to the RegCM. SSTs were obtained from the Reynolds weekly SST data with horizontal resolution of 1° × 1° (Reynolds et al. 2002). These PCL runs were employed to test the model’s ability to simulate present-day tropical cyclone climatology.

2) For the control experiment (CTL) the driving fields are obtained from the 6-hourly output of the control run of the NCAR CCSM2. An ensemble of 10 JASO integrations was performed, each started from the 1 July results from one of 10 consecutive years of the CCSM2 control run.

3) The 6 CO$_2$ conditions of the climate change experiment are obtained from the 6-hourly output from the last 10 yr of the NCAR CCSM2 global warming run. The NCAR CCSM2 has a value of global-mean climate sensitivity among the lowest of the current generation of global coupled climate models (Stowasser et al. 2006). The mean SST increase in the last 10 yr of the 6 × CO$_2$ integration relative to the CTL experiment in the months JASO is shown in Fig. 1 for the domain of present interest. Over most of the tropical cyclone generation region the SST increase is around 3 K. The global-mean surface air temperature warming is 4.5 K.

c. Evaluation of basic model climatology

The contours in Fig. 2 show the July–September long-term mean sea level pressure from observations (ERA-40) compared to results from the PCL, CTL, and 6 × CO$_2$ experiments. The arrows show the 850-hPa wind field.

The observed large-scale flow is dominated by the
East Asia–western Pacific summer monsoon trough and the western Pacific subtropical high (Fig. 2a). In long-term averages of low-level wind flow and sea level pressure the monsoon trough of the western North Pacific (during Northern Hemisphere summer) extends eastward from the south Asian low pressure trough and is accompanied by low-level southwesterly winds to the south of the trough axis. The relative vorticity associated with the monsoon trough in the western North Pacific is a vital factor for tropical cyclone formation in that region (Holland 1995). The western Pacific subtropical high moves northward in June and is at its northernmost position near 40°N in August and September. The ERA-40 climatology of low-level winds and sea surface pressure is reproduced by the PCL experiment very well (see Fig. 2b). The CTL experiment (see Fig. 2c) also captures these two main features of the circulation, although the western Pacific subtropical high appears to be too strong and is accompanied by unrealistically strong easterly winds penetrating too far into the monsoon trough region. At the same time the low pressure trough is somewhat too weak in the CTL simulation. The 6 × CO₂ experiment is very similar to the results of the CTL experiment.

Figure 3 shows a similar comparison of observational and model climatology for the 500–700-hPa layer mean relative humidity.

In the summer months the dry subtropical ridge and the wet monsoon trough also control the moisture in the middle troposphere. The largest values (above 70%) can be found in the monsoon trough region extending from the South China Sea to about 170°E. The humidity pattern is very well represented in the PCL-control climate. Also the 10-yr mean relative humidity pattern of the CTL experiment is in good agreement with the ERA-40 reanalysis data in most of the domain. However, in contrast to the ERA-40 data, which show the tongue of high values only extending to 170°E, the

Fig. 2. Mean sea level pressure (hPa) and wind field (m s⁻¹) at 850-hPa level averaged over July–September for (a) ERA-40 data, (b) the PCL experiment, (c) the CTL experiment, and (d) the 6 × CO₂ experiment. The results are averaged over 10 yr (RegCM simulations) and 43 yr (ERA-40), respectively.
simulations show considerably larger values farther east. The $6 \times \text{CO}_2$ experiment shows an increase in the midtropospheric relative humidity in this band.

The humidity pattern is reflected in the observed precipitation climatology shown in Fig. 4a.

The rainfall data over land for this climatology is taken from the Legates and Willmott (1990) analysis of historical rain gauge measurements. The ocean precipitation estimates are from the Microwave Sounding Unit (Spencer 1993) for the years 1979 to 1992. The observations exhibit large values of rainfall in the western Pacific warm pool region influenced by the monsoon through and only little precipitation under the influence of the western Pacific subtropical high. In the present-day model simulations (Figs. 4b-c) the precipitation seems to be unrealistically confined to the high moisture band extending from the South China Sea to about 170°E.

Overall the regional climate model forced by either the ERA-40 reanalysis data or the CCSM2 model can capture reasonably well the basic features of the climate in the regional model domain.

d. **Criteria for identifying tropical cyclones**

The method for the detection and tracking of the model tropical cyclones is similar to that documented in Nguyen and Walsh (2001) and relies upon detection criteria that are based on observed tropical cyclone characteristics. The following criteria must be satisfied for a system that we identify and track as a tropical storm in the IPRC-RegCM simulations:

1) There must be a vorticity local maximum at 850 hPa exceeding $2.5 \times 10^{-5} \text{ s}^{-1}$.
2) There must be a local pressure minimum within a radius of 250 km of the vorticity maximum; this minimum pressure is taken as defining the center of the storm.
3) The azimuthal mean tangential wind speed at the 850-hPa level must be higher than at 300 hPa.
4) The total tropospheric temperature anomaly (defined as the azimuthal mean temperature minus the mean temperature within a 10° circle around the center of the tropical storm) between 300 and 850 hPa must be greater than zero.
5) The closest local maximum of averaged temperature between 500 and 200 hPa is distinguishable and is defined as the center of the warm core. The distance between the center of the warm core and the center of the storm must not exceed 2.5° latitude. From the center of the warm core, the temperature must decrease by at least 0.5°C in all directions within a distance of 7.5° latitude.

To be considered as a model tropical storm trajectory, a storm must last at least 2 days and have a maximum wind speed at the lowest model level larger than 17 m s⁻¹ during at least 2 days. Cases satisfying all these criteria are referred to as tropical storms in this study.

e. Observational tropical cyclone data

Our model simulations are compared with the "best-track" dataset produced by the Joint Typhoon Warning Center (JTWC). The JTWC dataset provides 6-hourly estimates for the central position and intensity (peak surface wind) of each tropical storm identified in the western North Pacific. There are almost no direct in situ observations available within these storms, and the JTWC data are produced with indirect methods—in recent decades based mainly on interpretation of satellite cloud imagery. The dataset is provided by JTWC back to 1945, but in our study we consider only the data after 1970. Tropical cyclone detection rates have been close to 100% globally since around 1970, when global satellite coverage became nearly complete (Holland 1981). Reliable estimation of storm intensity depends on in situ aircraft measurements and satellite-based techniques. Techniques for estimating tropical cyclone intensity from satellite imagery were developed during the 1970s (Dvorak 1975). These techniques constitute today the primary means of estimating peak surface winds in the North Pacific since the reconnaissance aircraft measurements were discontinued in the western North Pacific in 1987.

In the absence of detailed data the JTWC cannot use
dynamical and thermodynamic criteria for identifying tropical cyclones like those we use for our model simulations, and this has to be considered when comparing JTWC data with our model results. In particular, the criteria described in section 2c above tend to class storms as extratropical rather than tropical as they move out of the subtropics, while JTWC generally tracks tropical storms for a considerable distance out of the Tropics.

3. Results

Previous studies (e.g., Bengtsson et al. 1995) have shown reasonable agreement between observed tropical storm characteristics and tropical storm–like vortices generated by a GCM. Although the models generally have the tendency to underestimate the storm intensity, there are qualitative similarities to their observed counterparts: convergence, high moisture content and strong upward motions at the lower levels, and anticyclonic vorticity and divergence in the upper troposphere.

Figure 5 presents results for a typical storm appearing in the present simulations, in this case from one October in the PCL simulation. In particular, Fig. 5 shows radial cross sections of the azimuthally averaged azimuthal wind, radial wind, p velocity, and temperature anomaly computed from a snapshot near the time of peak intensity for this storm.

Although the resolution of 0.5° is still too coarse to resolve all tropical cyclone features, the overall structure of the simulated storms in our study seems to be consistent with typical observations, for example, the composite of observed North Pacific typhoons of Frank (1977). The main deficiency in the model results is that the simulated storms are unrealistically large compared to observations. The radius of maximum winds is about 150 km for the storms in the simulation versus around 25–50 km in typical observed mature typhoons (e.g., Weatherford and Gray 1988). However, the simulated storms have the typical pattern of cyclonic inflow in the lower troposphere and anticyclonic outflow above 300 hPa. Strong upward vertical motion exists inside about a 2° radius. Further outside a weak subsidence below 400 hPa is simulated. The temperature anomaly reveals a pronounced warm core centered at around 350 hPa. Furthermore, the relative humidity is very high in the inner core of the storm with values up to 100% in the eyewall. All these features correspond well with the typical observed tropical cyclone presented in Frank (1977).

The results for the structure of a modeled tropical storm depend on both the resolution of the model and the model parameterizations employed. Knutson and Tuleya (2004) compared four convective parameterization schemes in the Geophysical Fluid Dynamics Laboratory (GFDL) hurricane model and showed that the convective parameterization can have an important impact on the structure and intensity of the simulated hurricanes. As a preliminary to the present study, we tested the original moisture–convergence closure scheme of Tiedtke (1989) and the modified convective available potential energy (CAPE) closure by Nordeng (1994). Specifically we repeated two seasons of the PCL experiment employing the different convective schemes but using the same initial and boundary conditions. We found that the CAPE closure usually produced model tropical cyclones weaker than the original moisture–convergence closure, but overall results remain almost the same. Our results discussed below are based on those from the moisture–convergence closure for the mass flux convective parameterization scheme of Tiedtke (1989).

Even in a perfect simulation, a model employing a 50-km resolution will not produce observed intensities, and the wind structures and the peak surface winds are expected to be lower than observed. Walsh et al. (2007) investigated how the intensities of tropical cyclones are degraded by using simulations with different resolutions. In their study they used objectively derived, resolution-dependent criteria that are defined for the detection of tropical cyclones in model simulations and observationally based analyses. These criteria are derived from the wind profiles of observed tropical cyclones, averaged at various resolutions. For example their analysis of the time evolution of the maximum wind speeds of Hurricane Andrew, which was an intense storm with a small eye and a small radius of gale force winds, showed that when degraded to a resolution of 300 km, the wind field only just reaches tropical storm strength. Hurricane strength is not exceeded until the grid spacing drops to 100 km. Thereafter, wind speed increases rapidly as resolution increases. At a resolution of 50 km, with which our experiments are conducted, maximum wind speeds reach 50 m s⁻¹ compared to 75 m s⁻¹ obtained with a 10-km resolution. This is in agreement with the fact that no storm in our control simulation reached more than 52 m s⁻¹.

a. Geographic storm distribution

The procedure for tracking model tropical storms is applied to the 10-member ensembles of all three experiments. Figure 6a displays the concentration of genesis positions of tropical storms in the JTWC best-track data (years 1971–2003). The units are numbers per 5° square box per 4-month
(JASO) period. Most of the systems form in a region between 10°–25°N and 130°–150°E and in the South China Sea. Many studies have identified the confluence region between the monsoonal southwesterlies and the trade easterlies as a region in the western North Pacific where tropical cyclone formation is common (e.g., Briegel and Frank 1997). Holland (1995) argued that the confluence region can trap tropical waves in the mid- to lower troposphere, which can maintain and modulate the moist convection in the region. Ritchie and Holland (1999) identified five characteristic, low-level circulation patterns in the WNP that are associated with TC formations: monsoon shear line, monsoon confluence, monsoon gyre, easterly waves, and Rossby energy dispersion. The first two patterns are associated with 71% of all TC formations. The PCL runs (Fig. 6b) show a

![Diagram](image-url)
good resemblance of the tropical cyclone genesis region. However, the maxima are somewhat stronger than in the best-track data.

While there is little dispute that climate models can generate tropical cyclones, earlier studies have used different threshold criteria for deciding the cutoff between systems of tropical storm and tropical depression strength, which for observed storms is a 10-m wind speed of 17.5 m s\(^{-1}\). Walsh et al. (2007) argue that for models of limited horizontal resolution, a wind speed threshold that is lower than the observed is appropriate, because the model will not be able to generate storms that are as strong as those seen in reality due to its limited horizontal resolution. For horizontal resolutions finer than about 10 km, the results of Walsh et al. (2007) show that the observed threshold criterion of 17.5 m s\(^{-1}\) for the 10-m wind speed is appropriate. However, for a T106 climate model with an effective resolution of about 125 km, a 10-m wind speed of roughly 14.5 m s\(^{-1}\) is more appropriate. For a 50-km resolution model, a wind speed threshold of about 16.8 m s\(^{-1}\) should be applied that is close to the assumption made in our study.

The geographical distribution of the tropical cyclone genesis region is also simulated quite well in the CTL runs (see Fig. 6c). However, this control experiment shows also a maximum of genesis locations around the date line that has no counterpart in the JWTC best-track data. The tendency for unrealistically large numbers of storms generated east of 160°E is shared by a number of other models (e.g., Sugi et al. 2002; Tsutsumi 2002; Yoshimura and Sugi 2005). The present PCL and
CTL simulations both feature realistic amounts of tropical cyclone genesis in the South China Sea. This contrasts with many of the earlier model studies that appear to significantly underestimate the numbers of genesis events in the South China Sea (Bengtsson et al. 1996; Tsutsui 2002; Sugi et al. 2002; Yoshimura and Sugi 2005).

Figure 6d depicts the geographical distribution of genesis frequency for the $6 \times \text{CO}_2$ experiment. The average number of tropical storms generated in the JTWC data during 1971–2003 for JASO is 18.4, and for 1991–2000 it is 20.6. The comparable model result for PCL is 17.9, while for CTL it is 19.6, and for the $6 \times \text{CO}_2$ simulation it is 23.3. The overall increase in the number of tropical cyclones in the global warming experiment is not statistically significant. However, this increase does contrast sharply with the substantial reductions in western North Pacific tropical cyclones reported in some earlier climate warming studies conducted with global models. Bengtsson et al. (1996) report a drop in average numbers of western North Pacific tropical cyclones from 32.8 in their control run to 21.6 in their doubled-$\text{CO}_2$ run (note that these numbers are for the full year). Sugi et al. (2002) find about 11 western North Pacific tropical cyclones during JASO in their control run and only about 5 in their doubled-$\text{CO}_2$ run (see their Fig. 4c).

Figure 7a shows a measure of the occurrence of the tropical storms over their whole lifetime derived from the JTWC best-track data (years 1971–2003).

The quantity plotted is the number of storms that were present in the domain counting each 12-hourly time step. The units are numbers per 2.5° square box per JASO period. Most of the systems occur in a region

**Figure 7.** Same as in Fig. 6, but for occurrence of tropical storms. The units are numbers per 2.5° square box per 4-month (JASO) period. Shaded areas indicate significance of the differences at a 95% level.
between 10°–30°N and 110°–150°E with a maximum in the South China Sea.

In the PCL runs there is fairly close agreement in the tropical storm distribution (Fig. 7b) with the JTWC data. Both model PCL results and JTWC observations display a clear maxima in the South China Sea. However, the observations show a wider distribution in latitude with more observed tropical cyclones over Japan and Korea than in the model results. Figure 8 shows the actual tracks of all storms in the JTWC data for JASO during 1991–2000, while Fig. 9 shows the same quantity for the 10 yr of CTL simulation.

The general pattern of the paths simulated in the Tropics looks reasonable, but the JTWC tracks extend much more frequently into the extratropics, notably over Japan and east of Japan. The difference is likely at least partly due to the difference in the criteria used to identify the tropical cyclones in the model and in the JTWC data (see section 2e). Furthermore, the observations include more interannual variability in the mean basin circulation than is present in the model runs. This could also contribute to the underestimation of storms north of the monsoon trough in the model results.

Figure 7c shows the tropical cyclone occurrence measure for the CTL experiment. The result is in reasonable agreement with the PCL simulation. Figure 7d shows the same quantity for the 6 × CO₂ experiment. Shading indicates 2.5° × 2.5° grid boxes for which the global warming and CTL results are judged different at the 95% confidence level (using a t test; e.g., Wilks 1995). The warming leads to significantly more storms occupying the South China Sea, but rather similar numbers as in the CTL experiment elsewhere. These results contrast with those of most earlier studies performed with global models. We repeated the statistical t-test analysis on coarser grids to rule out the possibility that our results reflect geographical shifts in the cyclone activity rather than a real increase in tropical cyclone numbers. Both on a 5° × 5° and a 7.5° × 7.5° grid the t test yields a significant increase in the number of tropical cyclones in the South China Sea region (not shown).

b. Seasonal distribution

Figure 10 shows the average number of storms present in each calendar month during JASO in the entire region considered here.

Results for the three model experiments, PCL, CTL, and 6 × CO₂, are presented along with JTWC data for 1971–2003 and for 1991–2000. Also shown are estimated error bars assuming that the values are sampled from a Gaussian distribution. The model PCL run and the observations both show maximum numbers in August. The GCM-driven cases, CTL and 6 × CO₂, both show a tendency for some more storms to occur in September and October than in August. However, the differences among model experiments and between the
model and observations are within the error bar estimates.

c. **Strength**

The tropical cyclone frequency distribution as a function of maximum wind speed of each cyclone was calculated for the CTL and the $6 \times CO_2$ experiments. The frequency depicted in Fig. 11 is given in numbers of 12-h time steps that a tropical cyclone was present in the domain.

The frequency distribution of simulated tropical cyclones exhibits a sharp peak at 20 m s$^{-1}$. The largest maximum wind speed of the simulated tropical cyclones in the CTL experiment is around 50 m s$^{-1}$. On the other hand, the frequency distribution of observed tropical cyclones has a broad peak at 20–25 m s$^{-1}$, and extends over a broad range of maximum wind speed up to 80–85 m s$^{-1}$ (Sugi et al. 2002). It is obvious that the model fails to simulate the observed very intense storms, presumably due to the horizontal resolution employed.

The results of the $6 \times CO_2$ runs show a distinct increase in the number of storms with wind speeds larger than 20 m s$^{-1}$. Additionally, the highest wind speeds found are around 60 m s$^{-1}$. The two binned datasets shown in Fig. 11 were tested to see if the distributions are significantly different using a chi-square test (e.g., Wilks 1995).

The chi-square statistic used is

$$\chi^2 = \sum \frac{(C_i/P_i - P/C_i)^2}{P_i + C_i},$$

where $C_i$ and $P_i$ are the numbers of events in bin $i$ for the CTL and $6 \times CO_2$ experiments, respectively, and $R$ and $P$ are the respective numbers of data points. The test results show that the null hypothesis that the two distributions are drawn from the same population distribution function can be rejected at a 99% significance level. Thus the shift to higher wind speeds in the $6 \times CO_2$ experiment seen in Fig. 11 is statistically significant.

To assess the tropical cyclone threat it is valuable to have a combined measure of tropical cyclone frequency and strength. Emanuel (2005) defined an index of the potential destructiveness of tropical cyclones based on the total dissipation of power, integrated over the lifetime of the cyclone. He showed that this index has increased markedly since the mid-1970s in the North Atlantic and Pacific. This power dissipation index (PDI) for a single storm is

$$PDI = \int_0^T V_{max}^2 \, dt,$$

where the integral is over the lifetime of the storm and $V_{max}$ is the lowest-level wind speed at each time. The PDI is then summed over all the storms in a given
The PDI computed for each year of each of the PCL, CTL, and $6 \times CO_2$ experiments is shown in Fig. 12.

The average PDI values are very similar between the CTL and the PCL experiments. However, considerably larger values are found in the $6 \times CO_2$ experiment. The mean values for the 10 yr are $1.8 \times 10^{11}$ m$^3$ s$^{-2}$ and $1.9 \times 10^{11}$ m$^3$ s$^{-2}$ for the CTL and PCL experiments, respectively. The PDI increases by 50% to $2.8 \times 10^{11}$ m$^3$ s$^{-2}$ for the $6 \times CO_2$ runs. This suggests that future warming will lead to an upward trend in topical cyclone frequency.

Fig. 10. Long-term (1971–2003 and 1991–2000) monthly mean number of tropical storms in the western North Pacific for JASO obtained from the Joint Typhoon Warning Center compared to results obtained by the RegCM.

Fig. 11. Tropical cyclone frequency distribution as a function of maximum wind speed of each cyclone for the CTL and $6 \times CO_2$ experiments. The frequency is given in numbers of 12-h time steps that a tropical cyclone was present in the domain.

Fig. 12. Accumulated PDI [$\times 10^{-11}$ (m$^3$ s$^{-2}$)] for the western North Pacific for results of the PCL, CTL, and $6 \times CO_2$ experiments.
destructive potential in the western North Pacific region. Note that the 50% increase in PDI is associated with a $\sim 3$-K ocean surface warming. By contrast Emanuel (2005) finds that the PDI computed for the western North Pacific rises by a factor of 2 over the last three decades, during which the ocean warmed by about 0.5 K.

The PDI values are influenced by the number, average lifetime, and strength of the simulated tropical storms. As we noted earlier, the number of tropical cyclones only slightly increases in the global warming experiment. The average lifetime of the storm systems in the CTL experiment is 7.0 days, which is shorter than 8.7 days, the average observed lifetime of tropical cyclones in the western North Pacific reported by Camargo and Sobel (2005). The lifetime of the storms increases slightly in the $6 \times CO_2$ experiment to 7.7 days. However, applying a $t$ test reveals that this 10% increase is not statistically significant. Thus, the increase in the PDI is dominated by the intensity changes of the storms, which were discussed above.

4. Discussion

We computed the parameters usually considered to be closely related to tropical cyclone formation for the ensemble mean of the CTL and $6 \times CO_2$ experiments. The SST fields are shown in Fig. 1. From a physical point of view, a sufficient warm ocean mixed layer provides the necessary thermal energy and surface heat flux for convective development (Gray 1998). Gray has proposed that the heat content of the top 60 m of ocean is more relevant than the SST itself. We examined the depth of the warming in the ocean component at the end of the $6 \times CO_2$ CCSM2 model run and found that in the tropical western North Pacific the warming was relatively uniform down to $\sim 100$ m.

During middle summer, virtually the entire western North Pacific Tropics have a high enough SST ($>26^\circ$C) in the CTL run for tropical cyclones to develop. The theory by Emanuel (1988) indicates that the potential intensity of hurricanes depends mostly on the ocean and atmospheric temperatures. Our simulated increase in strong storms is consistent with the SST increase seen in the tropical cyclone genesis region in the global warming experiment. However, interaction of tropical cyclones with their atmospheric and oceanic environment limits the intensity of most storms to values well below the theoretical limit. In the following analysis the differences between the $6 \times CO_2$ and the CTL experiment in the large-scale environmental parameters that can affect the tropical cyclone formation are investigated (see Fig. 13).

Tropical cyclone formation usually occurs in a low vertical wind shear environment (Gray 1968, 1998). The vertical wind shear

$$S_z = \sqrt{(u_{850} - u_{200})^2 + (v_{850} - v_{200})^2}$$

was calculated for the CTL and $6 \times CO_2$ experiments, where $u$ and $v$ are the zonal and meridional wind components at 850 and 200 hPa, respectively. The ensemble mean of the 10-yr control runs shows large values of $S_z$ in the South China Sea region and smaller values farther to the east. The pattern of $S_z$ is in good agreement with the long-term mean of ERA-40 data (not shown). In the $6 \times CO_2$ experiment $S_z$ is reduced in the western
part of the basin (see Fig. 13a). This corresponds well with the region with the most pronounced increase in the tropical storm occurrence seen in Fig. 7. Farther east an increase in the vertical shear is simulated in the $6 \times CO_2$ experiment.

A high value of relative humidity in the middle troposphere is favorable for tropical cyclone formation (Cheung 2004). The 500–700-hPa layer mean relative humidity is used to represent the moisture in the middle troposphere. As shown in Fig. 3 the large-scale distribution of moisture is controlled in the summer months by the dry subtropical ridge and the wet monsoon trough. The maximum of tropical cyclone occurrence seen in Fig. 7 corresponds to the band of high relative humidity between $5^\circ$ and $20^\circ$N. As discussed in section 2c, in contrast to the ERA-40 data, which show the tongue of high values only extending to $170^\circ$E, the simulations show considerably larger values farther east. This may partly explain the second maximum of tropical cyclone occurrence, which was simulated by the RegCM around the date line at $15^\circ$N (see Fig. 7). In Fig. 13b the changes in the relative humidity between the 10-yr mean of the $6 \times CO_2$ and the CTL experiments are shown. The relative humidity increases slightly in the $6 \times CO_2$ case in most of the domain considered. A larger increase is found around the equator at $160^\circ$E, which is caused by an extension of the moist monsoon trough to the southeast in the $6 \times CO_2$ experiment. The slight increase in relative humidity found in the RegCM favors enhanced tropical cyclone activity.

The change in dry static stability is shown in Fig. 13c. Here the dry static stability is defined as the difference in potential temperature at 200 and 1000 hPa. This measure increases by around 18% in the $6 \times CO_2$ experiment, suppressing convective development, which is essential for tropical cyclone formation. In contrast to the rather uniform increase in the dry static stability, the change in convective instability as measured by CAPE shows a distinct regional distribution (see Fig. 13d). It decreases over most of the region with high tropical cyclone activity, although the surface moisture and the moist static energy of the surface air increases everywhere in the Tropics (not shown). Therefore the reduction in CAPE in this region is due to the increased dry static stability, which is larger than the increase in CAPE due to the increase in moist static energy. Around the equator east of $160^\circ$E the CAPE values increase in the $6 \times CO_2$ experiment. Although CAPE is considered of importance to deep convection in the Tropics, the consequence of deep convection however is to remove the environmental CAPE. In the life cycle of a tropical cyclone, some CAPE is needed in the formation stage before any mesoscale organization of a cyclonic core develops. After its formation, a tropical cyclone can intensify by extracting energy from the underlying ocean and then become self-sustained under favorable environmental conditions (Emanuel 1987, 1988). As a result, the change in CAPE between the CTL and $6 \times CO_2$ experiments does not reflect the actual change in tropical cyclone activity; rather, the reduced vertical shear is largely responsible for the significant increase in the tropical storm intensity (Fig. 13a).

Figure 14 shows the same quantities computed from the global CCSM2 simulations. Overall there is agreement between the results of the global and regional climate model regarding the changes in these large-scale environmental parameters associated with tropical cyclone activity. In the South China Sea the large-
The change in cyclogenesis frequency in the various parts of the domain in our experiments seems to be consistent with our understanding of the large-scale environmental factors that control tropical cyclone formation. Notably, the increase in cyclogenesis in the South China Sea in the global warming experiment accompanies increases in relative humidity in the midtroposphere and decreases in the vertical wind shear. These favorable changes in the large-scale environment are not found in most other regions of the western North Pacific, where lower relative humidity, higher stability, and stronger vertical shear appear in the warm climate simulation. The long-term mean values of these relevant environmental factors, and their changes in response to global warming, are similar in the nested regional atmospheric model to those in the coarse-resolution global coupled climate model.

We find that the average intensity (measured by the peak wind speed) and the number of intense storms over the entire western North Pacific region rises significantly in the global warming simulation. This result seems to be common to all the previous studies of global warming influence on tropical cyclones. We calculated the Power Dissipation Index defined by Emanuel (2005) for all the storms in our present-day and global warming climate simulations. We find that the PDI rises by about 50% in the warm climate. The ratio of the PDI increase to the magnitude of tropical ocean surface warming may be a useful measure of the effectiveness of tropical storm–induced ocean heat transport to stabilize tropical climate, as proposed by Emanuel (2001). This ratio as obtained in our simulations may also be compared with those from observational studies that attribute increasing PDI to observed tropical SST trends (e.g., Emanuel 2005).

Since the results of this regional climate model study depend on the forcing of the global model employed, it would be desirable to drive the regional model with results from more than one GCM. The recent effort of the International Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) climate model intercomparison provides the community with a large number of global climate model outputs that are suitable for the use in a similar study. It is planned to investigate the dependence of the results obtained in this paper on the choice of the global model used and the resolution of the regional model itself.

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REFERENCES

Bengtsson, L., 1996: Will greenhouse gas-induced warming over the next 50 years lead to higher frequency and greater intensity of hurricanes? Tellus, 48A, 57–73.


dium-Range Weather Forecasts, Reading, United Kingdom, 63 pp.


