Influence of Greenhouse Warming on Tropical Cyclone Frequency

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Abstract

Influences of sea surface temperature (SST) spatial patterns and cumulus parameterizations on tropical cyclone (TC) frequency, in the context of global warming impacts, are investigated using an atmospheric general circulation model at T106 horizontal resolution. Simulated TCs in this high-resolution model are categorized into tropical storms (TSs) and tropical depressions (TDs). Model TSs are defined as TCs with maximum surface wind speed more than, or equal to 16 m s\(^{-1}\), for experiments with an Arakawa-Schubert cumulus parameterization. Another threshold of 14 m s\(^{-1}\) is used for those with a Kuo cumulus parameterization. Model TDs are defined as weaker TCs. Although the maximum wind speed, and the minimum central pressures of intense TCs are not realistically simulated in the model, geographical patterns of TS formation seem to be realistically simulated, with climatological and El Niño/La Niña SST conditions.

A series of experiments is conducted with doubled CO\(_2\) and with increased SSTs. A spatial pattern of SST, made by uniform 2 K warming, is used for experiments with both of the cumulus parameterizations. El Niño-like and La Niña-like warming patterns of SSTs, are used with the Arakawa-Schubert scheme. In these global warming experiments, frequency of TS formation decreases by 9.0–18.4% globally, and some of these changes are statistically significant. While no coherent changes in global frequency of relatively intense TCs (e.g., maximum surface wind \(\geq 25\) m s\(^{-1}\)) are found in the warm-climate experiments, significant reduction in the total frequency of TSs and TDs resulted from all of these experiments. The results suggest that global frequency of relatively weak TCs may decrease in the future warm climate, but frequency of intense storms may either decrease or increase. Mean precipitation near TC centers is significantly heavier in the warming experiments than in the present-day experiments, as compared for TCs with the same maximum wind speed.

1. Introduction

Tropical cyclones (TCs), including typhoons and hurricanes, are probably the most devastating phenomena in the atmosphere, and sometimes cause large number of human deaths and huge economic losses. Therefore, it is very important for society to know how frequency and intensity of TCs change in response to projected global warming in the 21st century. The Intergovernmental Panel on Climate Change (IPCC 1990, 1996, 2001) has made efforts to assess the impact, and has concluded in the most recent report (Chapter 10 in IPCC 2001) that “there is some evidence that regional frequencies of tropical cyclones may change but none that their locations will change. There is also evidence that the peak intensity may increase by 5–10% and precipitation rates may increase by 20–30%. There is a need for much more work in this area to provide more robust
results." In the present paper, we attempt to deal with the frequency aspect of this issue, by a sensitivity study using an atmospheric general circulation model, with a high horizontal resolution.

An empirical relationship between large-scale atmospheric/oceanic fields, and frequency of climatological TC formation, was established by Gray (1979). But the relationship may change significantly in response to global warming (e.g., Ryan et al. 1992), so it is inappropriate to use for estimation of TC frequency from large-scale fields in warm climates simulated with numerical models, even if the simulated large-scale fields are assumed to be accurate. Royer et al. (1998) proposed a modification of the relationship, which could be a useful diagnostic of TC frequency change, due to greenhouse warming. Druyan et al. (1999) pointed out that, in response to CO₂ doubling, vertical lapse rate of the equivalent potential temperature, which is included in Gray’s (1979) relationship, is simulated to be more favorable for TC genesis.

Another hopeful method for projection of future changes in TC frequency, is to use climate models which can directly resolve the basic dynamics and thermodynamics of individual TCs. There are several studies with this method using relatively-low-resolution global climate models (e.g., Broccoli and Manabe 1990; Haarsma et al. 1993; Tsutsui 2002), and a few ones using higher-resolution global models, of about 100-km horizontal grid spacing (e.g., Bengtsson et al. 1996; Sugi et al. 2002).

Bengtsson et al. (1996) and Sugi et al. (2002) found in their high-resolution-model experiments, that global TC frequency decreases significantly in response to greenhouse warming. Sugi et al. pointed out that weakening of atmospheric circulation in the tropics, associated with stabilization of the troposphere (or increase in dry static stability), is a possible reason for the reduction in global TC frequency due to global warming. On the other hand, regional variations of the TC frequency in the warm climates are significantly different between the two studies. Such a difference may be attributable to differences in simulated spatial patterns of global warming, and parameterization of physical processes in the respective GCMs.

High-resolution models are generally desirable to simulate TCs, because a typical horizontal scale of TCs, is one thousand kilometers or less. The scales of convective cloud bands, and the inner-core structures of mature-stage TCs are, however, on the order of dozens of kilometers (e.g., Croxford and Barnes 2002), and they cannot be resolved even with higher-resolution models of 100-km grid spacing. The maximum wind speed of TCs, which is related to the inner-core structures, is not usually reproduced with such models either. Nevertheless, we believe that it is possible to realistically simulate the TC formation, on the larger horizontal scales, and its frequency using climate models.

Landsea (1997), however, commented on the paper by Bengtsson et al. (1996). One of the major concerns of Landsea is the possibility of inconsistency between a low-resolution simulation, from which warm sea surface temperature (SST) data were taken, and the high-resolution simulation of the warmed atmosphere. Landsea pointed out that the doubled CO₂ SSTs could be substantially different, if a long-term simulation is conducted using the high-resolution model coupled with an ocean GCM. In addition to the SST issue, the dependence of TC simulation on cumulus parameterization has been a key issue for TC modeling (e.g., Vitart et al. 2001; Tsutsui and Kasahara 2000; Ueno and Yoshimura 2002). Vitart et al. (2001) showed strong sensitivity of model TC frequency to changes in convective parameterization, especially to modifications that place restrictions on the production of deep convection (e.g., Toikioka et al. 1988), and Vitart et al. indicated that parameterization-induced changes in the tropical mean atmosphere may be responsible for the changes in simulated TC climatology. Tsutsui and Kasahara (2000), also obtained results of strong sensitivity to a relative humidity threshold, as a restriction on deep convection. Ueno and Yoshimura (2002) found significant sensitivity of simulated TCs to radiative processes, as well as sensitivity to cumulus parameterization.

The purpose of this paper is to investigate the influences of SST spatial patterns, and cumulus parameterizations on simulated TC frequency in global warming experiments, using a high-resolution atmospheric model, with the
same horizontal resolution as Bengtsson et al. (1996) and Sugi et al. (2002). We examine whether the reduction of TC frequency reported by them is a robust result or not, by a series of numerical experiments using various SST patterns and two types of cumulus parameterizations by Kuo (1974), and Arakawa and Schubert (1974). The present paper covers a wider range of SST patterns than those of the past studies. The model of this paper is a newer version of the JMA model used by Sugi et al. (2002), and have been improved in terms of active TC formation.

After brief description of the model, and the numerical experiments in Section 2, the definition of TCs in the model is presented in Section 3. The results are shown in Sections 4 for the present climate, and 5 for the greenhouse-warmed climate, followed by the discussion in Section 6. We conclude this study in Section 7.

2. Model and experimental design

2.1 Model

The high-resolution atmospheric general circulation model (GCM) used for the present study is a T106 version of the Japan Meteorological Agency Global Spectral Model (GSM9603), which was developed at the Japan Meteorological Agency (JMA) for operational numerical weather prediction (JMA 1997). The model is a spectral model, with triangular truncation at total wave number 106 (equivalent to approximately 110-km grid spacing), and it has unevenly spaced 21 vertical levels. Comprehensive physical processes are included in the model.

Sugi et al. (2002) used a previous version of the JMA model (GSM8911) for their study on changes of TC climatology. A Kuo-type scheme (Kuo 1974) is adopted for deep convection in this earlier version, while three cumulus schemes can be used alternatively for the newer version in the present study. Other differences between the two versions include changes in cloud radiation processes.

A convection scheme used for operational weather forecast with GSM9603, is based on an Arakawa-Schubert-type scheme (Arakawa and Schubert 1974), with a prognostic closure (Randall and Pan 1993). In our preliminary experiment (Yoshimura and Sugi 1997) with the prognostic Arakawa-Schubert scheme, formation of TCs is not active (only 25 TCs formed in a year based on criteria of the next Section). Therefore we adopted the other two alternative schemes for cumulus parameterization in the present study: an original Arakawa-Schubert scheme, modified by Kuma (1997), and the Kuo scheme of GSM8911. As described in Section 4, active formation of TCs is simulated in experiments with each of these two schemes.

2.2 Numerical experiments

We conducted six numerical experiments with the Arakawa-Schubert cumulus scheme, and two experiments with the Kuo scheme, as summarized in Table 1. Integration lengths are 10 years for ASCL and AS2U, and 5 years for the other experiments. For the greenhouse-warmed climate simulations (AS2U, AS2E, AS2L and Kuo2U), CO2 concentration is doubled in the model atmosphere.

For the present-day climate simulations, three prescribed SST patterns are used:

1. Climatological SST from observation. (Used for ASCL and KuoCL)
2. A composite El Niño pattern. (Used for ASEN)
3. A composite La Niña pattern. (Used for ASLN)

Observed SST data from 1951 to 1993 are used to make the composite El Niño and La Niña monthly SST anomalies, which varies seasonally but not interannually.

Three other SST patterns are used for the warm climate simulations:

4. SST is increased by 2 K uniformly from the climatological SST. This pattern is used for AS2U and Kuo2U. Such experiments changing SST uniformly by ±2 K have been performed so far to investigate climate feedback processes (e.g., Cess et al. 1990, 1996; IPCC 1996).
5. A natural variation pattern of SST, generated by a coupled GCM, is added to the uniformly increased SST (4). The variation pattern is the first mode of EOFs of inter-decadal variation of annual-mean SST in the low and middle latitudes (Tokioka et al. 1995), simulated by the Meteorological Research Institute (MRI) coupled ocean-atmosphere GCM. The pattern is used for all months of the AS2E integration. Be-
cause it is somewhat similar to the El Niño pattern, this pattern is called ‘El Niño-like warming pattern’ in this paper. Greenhouse-warming patterns of SST simulated by some coupled models are somewhat similar to this pattern (e.g., Knutson and Manabe 1998).

(6) The natural-variation pattern of SST, same as (5), is subtracted from the uniformly increased SST (4). The pattern is used for all months of the AS2L integration. It is called ‘La Niña-like warming pattern’ in this paper. The greenhouse-warming pattern simulated by the MRI coupled model (Tokioka et al. 1995) is similar to this pattern.

For ASEN, ASLN, AS2E and AS2L, the SST patterns of differences from that of ASCL are shown in Fig. 1. We assume that the TC climatology is not largely affected by changes of sea ice, and we used the same sea ice data (climatology from observation) in all of the experiments.

The initial conditions for all the experiments were taken from the GANAL dataset (global objective analysis of the JMA) for 1 January 1995. The timescales of transition from the initial state to the equilibrium state should be relatively short (within a few weeks) for the atmospheric GCM, because long-term processes of the ocean circulation are not included in it. Therefore, the model output of all the integration length (5 or 10 years) are analyzed in the present study. To confirm the influence of the initial condition, we examined tropospheric temperature changes in the first one-year of AS2U and ASCL, and found that temperature is apparently affected in the first month only.

3. Definition of TCs in the model

3.1 Selection of TCs

The following criteria are used for selection of TCs in the model:

1. A grid point with local-minimum surface pressure, and with magnitude of relative vorticity at 850 hPa above $3.5 \times 10^{-5} \text{ s}^{-1}$, is selected as a TC-center candidate.

2. Maximum wind speed at 850 hPa near the point, is larger than 15 m s$^{-1}$.

3. Warm-core criterion: $\Delta T_{sum} = \Delta T_{300} + \Delta T_{500} + \Delta T_{700} \leq 2.0 \text{ K}$. Here $\Delta T_{300}$, $\Delta T_{500}$, and $\Delta T_{700}$ are temperature deviations at 300, 500, and 700 hPa, respectively, from the averages over a $12.375^\circ \times 12.375^\circ$
square around the point of minimum pressure.

[4] Maximum wind speed at 850 hPa near the point is larger than that at 300 hPa.

[5] Duration of a TC is not shorter than two days.

We used GCM results at 00 UTC on each day during the model integration. In order to determine duration of each TC (for use in the above criterion [5]), temporal continuity of a TC is defined as follows: In the next-day data, events which satisfy the criteria [1]–[4] are searched within 800 km of the original TC position. If one event is found, it is regarded as the continued one. If no events are found, the TC is regarded as ended. If two or more events are found (such a case is relatively rare), the one

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**Fig. 1.** SST anomaly: (a) ASEN minus ASCL, (b) ASLN minus ASCL, (c) AS2E minus ASCL, and (d) AS2L minus ASCL. Shading indicates large warming above 2 K. The unit is K.
nearest to the original TC position, is regarded as the continued one.

These criteria are the same as those of Sugi et al. (2002), which is based on the criteria of Bengtsson et al. (1996). Although Bengtsson et al. and Sugi et al. limited the search to ocean areas only, we did not put any geographical restriction.

Note that surface wind speed, which is often used for the observational data, is mainly used in the following sections as a measure of TC intensity in the simulations, while 850-hPa wind data are used in the criterion [2] above.

3.2 Tropical storms and tropical depressions

Observed TCs with maximum sustained wind speed exceeding 17 m s\(^{-1}\) (34 kt) are generally classified as tropical storms (TSs). Although storms of intensity 33 m s\(^{-1}\) (64 kt) or more have different regional names (e.g., typhoons and hurricanes), we will simply refer to these storms, also, as TSs in this paper.
However, the criterion of 17 m s$^{-1}$ would not be appropriate for definition of TSs simulated in atmospheric models of 100-km grid spacing, because the inner-core structures of TCs are not reproduced with such models. Sugi et al. (2002) adjusted their criteria so that the annual number of counted TSs in the control experiment should be near the number (approximately 80) of observed TSs. In the present study, the annual-mean numbers of TCs are more than 150 in the present-day climate experiments (see Section 4.2 for details), using the same criteria (mentioned in Section 3.1) as those of Sugi et al. Although this is much larger than that of the observed TSs, we interpret the results to include the numbers of weak disturbances which correspond to observed tropical depressions.

Here we adopt additional criteria for definition of TSs in the model, so that the annual numbers of TSs in the present-day experiments, should be comparable to the observation. A threshold of 16 m s$^{-1}$ is used for maximum surface wind speed of TSs in the experiments with the Arakawa-Schubert scheme, and another threshold of 14 m s$^{-1}$ is used in the experiments with the Kuo scheme. Weaker TCs below these thresholds, are classified as tropical depressions (TDs) in the model.

4. TCs in the present-day climate simulations

4.1 Geographical distribution

Positions of TS formation simulated in the two control runs, with the climatological SST (ASCL and KuoCL) are shown in Fig. 2. As compared with the observed distribution of TSs (Fig. 2a), reasonably realistic geographical pattern is reproduced, with both of the cumulus parameterizations (Fig. 2b for the Arakawa-Schubert scheme and Fig. 2c for the Kuo scheme). In the South Atlantic Ocean, there
are a few TSs in the simulations. In the real atmosphere, also, an apparent hurricane was observed near the coast of Brazil in March 2004 (Comte 2005), while such an event is very rare. Although a few TDs at unrealistic positions are simulated (e.g., near the Himalaya Mountains, figure not shown), some of them may be removed by changing the criteria for definition of TCs.

### 4.2 Total frequency

Annual total frequencies of simulated TS and TD formation are shown in Fig. 3. Surface wind intensities are indicated as shadings in the figure. The yearly numbers of the simulated TSs and TDs, also, are shown in Fig. 4. Annual-mean global TS and TD frequencies of ASCL, ASEN, ASLN and KuoCL are 153.4, 158.4, 156.4 and 170.0 respectively. In both of ASEN and ASLN, total frequency of TSs and TDs is not significantly different from that of the control run (ASCL). Total frequency of TSs and TDs in KuoCL is larger than that of ASCL, and the difference is statistically significant at 95% confidence level. Here, two-sided Student's $t$-test is applied to the differences between the 5-year data of KuoCL, and the first 5 years of 10-year data of ASCL. In later parts of this paper, statistical significance is tested in the same manner.

Using a previous version of the model (JMA-GSM8911), Sugi et al. (2002) simulated a fewer number of TC formation, with the same criteria for TC definition (about 88 TCs in a year) in their control experiment. The Kuo cumulus parameterization used in their experiments is the same as that of the present study for KuoCL and Kuo2U. Some modification made by the JMA for the newer version, including changes in cloud radiation processes, seems to be responsible for the difference in TC frequency.

Strong sensitivity of simulated TC frequency to changes in physical processes has been reported by a few earlier studies (e.g., Vitart...
The above results of the present study, also, indicate significant sensitivity to physical processes. We speculate that such sensitivity could be attributed to changes in climatological precipitation, static stability, vertical distribution of convective heating, etc. Further investigation of sensitivity to physical processes is one of the most important issues for numerical simulations of TC genesis.

TC frequency is also sensitive to changes in the criteria of TC definition. For example, Table 2 shows results based on additional criteria on the maximum surface wind speed.

4.3 Seasonal variations

In Fig. 5, seasonal variations of TS formation are shown for the observation and for the numerical experiments. The observed TS frequency is very low from January to April in the Northern Hemisphere (North of 10°N), and from June to October in the Southern Hemisphere (South of 10°S). For all the experiments, the simulated TS frequency is also very low in the wintertime, although the wintertime TS frequency is still larger than the seasonal minimums of the observed TSs. Amplitude of the simulated seasonal variations is smaller than that of the observation.

Sugi et al. (2002), also indicated that amplitude of simulated seasonal variations is too small for the control experiment, using the previous version of the model (JMA-GSM8911).

4.4 Impacts of El Niño and La Niña

In Fig. 6, impacts of El Niño and La Niña are shown as longitudinal distribution of TS frequency differences for the observation, and the numerical experiments. For the observational data (Figs. 6 b–c), El Niño (La Niña) is defined as periods when monthly SST anomaly in the Nino-3 region (4°N–4°S, 150°W–90°W) is higher (lower) by more than 70% of standard deviation. In this definition of El Niño/La Niña, we used 5-month-running-mean SST data, obtained from Tokyo Climate Center at the JMA, for a period of 1979–1998, and the standard deviation is calculated for each

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Relative changes (%)

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**Statistically significant decrease at 99% confidence level.
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month. Note that this period (1979–1998) was chosen for analyses of observational TS track data, and it is different from the period (1951–1993) used to make the composite SST data for the El Niño and La Niña experiments (ASEN and ASLN).

In an eastern part of the Western North Pacific (around 150°E–180°; in Fig. 6e), TS frequency increased in the El Niño experiment (shown as ASEN minus ASCL), and decreased in the La Niña experiment (shown as ASLN minus ASCL). In the North Atlantic basin (around 60°W; in Fig. 6e), TS frequency decreased in ASEN. These changes due to El Niño/La Niña are generally consistent with observational studies (e.g., Gray 1984, 1993; Chan 1985; Lander 1994), and mostly similar to changes in the observational graph (Fig. 6b). In the Southern Hemisphere, an eastern shift of TS formation near the date line in ASEN (Fig. 6f), is
similar to those in the observational graph (Fig. 6c). However, regions where simulated changes are statistically significant are limited, because the integration length (5 years for ASEN/ASLN) is not long enough.

4.5 Frequency of intense TCs

In this subsection, relatively intense TCs in the present-day simulations are examined. The maximum surface wind, and minimum sea level pressure of TCs during the ASCL experiment...
are 40 m s\(^{-1}\) and 952 hPa, respectively. They are 35 m s\(^{-1}\) and 959 hPa, respectively, during the KuoCL run. Extremely high wind speed, sometimes observed near the centers of TCs in the real atmosphere (e.g., around 70–80 m s\(^{-1}\)), is not reproduced with our model. Very deep central pressures (e.g., below 900 hPa) of the most intense typhoons and hurricanes, are not simulated in the model.

It has been suggested from observational studies (e.g., Merrill 1988) that the upper limit of TC intensity increases with SST. Theoretical and modeling studies (e.g., Emanuel 1987; Holland 1997; Knutson et al. 1998) also suggest a similar relationship. Scatter diagrams of SST against TC intensity (in terms of maximum wind speed) are shown in Fig. 7 for the simulations of the present study, and the observation from Evans (1993). While the relationship between SST and the maximum intensity of TCs can be clearly seen in the observational data (Figs. 7 c–d), the relationship of the simula-

Fig. 7. Scatter diagrams of TCs: SST versus maximum surface wind.
(a, b) Global data from the control experiments (ASCL and KuoCL).
(c, d) Observation for the North Atlantic and the western North Pacific basins (from Evans 1993).
These results indicate that our model does not have the capability to reproduce realistic intensity of strong TCs. This is most probably due to insufficient horizontal resolution (~110 km) of the model, in which the inner-core structures of TCs are not reproduced.

It should be noted that the time-mean (climatological) SST data are used in these experiments, while the real TC develops over the ocean surface, which varies significantly with time. The climatological SST distribution should have weaker horizontal gradients and lower peaks than the real “snapshot” SST distribution. The effect of storm-ocean coupling is not included in the model. These factors may affect the TC intensity in the simulations.

Much larger number of relatively intense TCs (≥20 or ≥25 m s⁻¹) are simulated in the experiments with the Arakawa-Schubert scheme than those with the Kuo scheme (see Fig. 3). This suggests that occurrence of relatively intense TCs strongly depends on the cumulus parameterizations used in numerical models.
5. Influence of greenhouse warming on simulated TCs

As stated in the Section 4, geographical distribution of TSs in the control experiments is realistic, and effect of El Niño/La Niña SST anomaly on TS frequency in the model also seems realistic. These results encourage us to investigate impacts of greenhouse warming on TSs (and TDs) simulated with the model.

5.1 Changes in TC frequency associated with uniform SST warming

In Fig. 3, the greenhouse-warming experiment, with uniformly high SST using the Arakawa-Schubert scheme (AS2U) shows significant reduction (by about 20%) in the total frequency of TSs and TDs, as compared with those for the present-day climate (ASCL, ASEN and ASLN). With the Kuo scheme, also, the result of the uniform-warming experiment (Kuo2U) shows smaller frequency of TSs and TDs than that of the control run (KuoCL) by about 20%. As shown in the panels of “ALL TCs” in Fig. 4, interannual fluctuations and trends are not significantly larger than these global-warming impacts on TS and TD frequency.

Table 2 shows the total frequency of TSs and TDs based on additional criteria on the maximum surface wind speed. For the experiments with the Arakawa-Schubert scheme, 14.8% reduction in global frequency of TSs (≥ 16 m s⁻¹) results from the uniform SST warming (ASCL → AS2U). For the experiments with the Kuo scheme, 9.0% reduction in the global number of TSs (≥ 14 m s⁻¹) results from the uniform warming (KuoCL → Kuo2U), but this change is not statistically significant.

Longitudinal distribution of TS and TD frequency in ASCL and AS2U is shown in Fig. 8a. The simulated reduction in TS and TD frequency, due to the uniform warming, is not confined to specific regions, but it is seen almost all over the longitudes where TCs occur. Latitudinal distribution of TS and TD frequency is shown in Fig. 8b, and the simulated frequency is reduced both in the Northern and Southern Hemispheres in response to the uniform warming. For TCs whose maximum wind reaches 15 m s⁻¹ or more, similar reduction is found in longitudinal and latitudinal distributions (Figs. 8 c–d).

![TC frequency difference](image)

In Fig. 9, difference in TS and TD frequency associated with uniform SST warming is shown as ‘AS2U minus ASCL’ and ‘Kuo2U minus KuoCL’ with information on statistical significance. In both of the hemispheres for the two
warming experiments (Figs. 9b and 9c), reduction in TS and TD frequency is seen almost all over the ocean basins, although the changes are statistically significant only in some of the basins. One exception is the increase in the North Indian Ocean for the Kuo2U experiment (see the broken line in Fig. 9b).

According to empirical relationship (e.g., Gray 1979), sea surface temperature (and subsurface ocean temperature) should be above 26°C for TC formation in the present-day climate. Similar relationship is seen in our numerical experiments for present-day climate (Fig. 10). If such relationship with sea surface temperature is assumed to be applicable to warmer climate, a large increase in global TC frequency will be expected in response to greenhouse warming. However, this assumption is widely thought to be unrealistic (e.g., Ryan et al. 1992; Henderson-Sellers et al. 1998).

Here we investigate the relationship between TC frequency and SSTs in our greenhouse-warming experiments. The result of AS2U shows that, as compared with those of the three experiments for present-day climate (ASCL, ASEN and ASLN), frequency distribution of TC formation shifts simply toward warmer temperatures by about 2 K, due to greenhouse warming (Fig. 10a). The lower limits for TS and TD formation also shift toward warmer temperatures. Similar shifts can be also seen in Fig. 10b between Kuo2U and KuoCL.

5.2 Changes in TC frequency associated with non-uniform SST warming

In Fig. 3 (see also Fig. 4), both experiments with the non-uniform SST warming, using the Arakawa-Schubert scheme (AS2E and AS2L), show significant reduction in total frequency of TSs and TDs, as compared with those for the present-day climate. Table 2 shows 9.2% (ASCL → AS2E) and 18.4% (ASCL → AS2L) reductions in global frequency of TSs (≥ 16 m s⁻¹), but only the latter is statistically significant. Reductions in global frequency of TSs were also shown by other studies with non-uniform SST warming, using high-resolution atmospheric GCMs (Bengtsson et al. 1996; Sugi et al. 2002).

Changes in longitudinal distribution of TS and TD frequency associated with the non-uniform SST warming, are shown in Fig. 11. Unlike the results with the uniform warming
these two experiments show rather confined regions of reductions in TSs and TDs: The Pacific Ocean in AS2L, and the Indian Ocean and the Atlantic Ocean in AS2E (Fig. 11a). In these ocean basins for the two experiments, the decreases are statistically significant in many longitudinal bands. For AS2L, the reduction in the Pacific Ocean is seen mainly in the Northern Hemisphere (Fig. 11b). The reduction in the Indian Ocean for AS2E occurs mainly in the Southern Hemisphere (Fig. 11c).

Relationship between TC frequency and SST for the two experiments with the non-uniform warming is similar to that of the uniform-warming experiment (Fig. 10a). Thus, all results of the four greenhouse-warming experiments (AS2U, AS2E, AS2L and Kuo2U) show shifts of TC formation toward warmer temperatures by about 2 K, as compared with those of the present-day climate experiments.

5.3 Changes in frequency of intense TCs

Knutson and Tuleya (1999, 2004) found that intense (i.e., damaging) TCs intensify further in global warming experiments, using a regional high-resolution hurricane model, which is forced with boundary conditions derived from warm-climate experiments with global ocean-atmosphere GCMs. Although the model of the present study cannot reproduce such highly intense TCs, it should be meaningful to examine relatively intense TCs in the model. In this subsection, changes in relatively intense TCs in the global-warming experiments, are examined by excluding weaker disturbances from the above analyses.

If the slightly severe threshold of 17 m s\(^{-1}\) is chosen in Table 2, the decreases in TS frequency is smaller than those of all TCs, and statistically significant reduction is found only in one case (ASCL → AS2U).

As shown in Figs. 3 and 4, global frequency of more intense TCs (maximum surface wind \(\geq 20\) m s\(^{-1}\), or \(\geq 25\) m s\(^{-1}\)) is not significantly changed in AS2U, substantially increased in AS2E, and slightly decreased in AS2L, as compared with ASCL. It is slightly larger in Kuo2U than in KuoCL.

Significant reduction in the total frequency of TSs and TDs is common to all of the greenhouse-warming experiments in this study. However, the above results indicate that frequency of intense TCs does not necessarily decrease due to greenhouse warming. It may depend on geographical patterns of sea surface temperature (and possibly on cumulus parameterization). For an example of AS2E, as shown in Fig. 12, increase of intense TCs is seen mainly in the western Pacific, which implies that such changes may not be a global feature, but a regional phenomenon.
5.4 Changes in precipitation rate near the TC centers

According to simulations with the regional hurricane model of Knutson and Tuleya (1999, 2004), near-TC precipitation becomes significantly heavier associated with global warming. However, their study was based on simulations of relatively intense TCs only, and dependence of TC precipitation on TC wind speed was not considered. In the present study, relationship between near-TC precipitation, and maximum wind speed is investigated.

Figures 13a and 14a show mean precipitation averaged near centers of TSs and TDs, as functions of maximum surface wind speed. For each experiment, precipitation almost monotonically increases as maximum wind speed. Qualitatively similar relationship, between average TC precipitation and maximum wind speed, has been shown from observational data (Cerveny and Newman 2000).

The greenhouse-warming experiments with the Arakawa-Schubert scheme (AS2U, AS2L and AS2E) show heavier precipitation for each wind-speed band (Fig. 13a), as compared with the present-day climate experiments (ASCL, ASEN and ASLN). Using the Kuo scheme, also, precipitation is heavier in Kuo2U than in KuoCL (Fig. 14a). Precipitation ratio of AS2U to ASCL is around 1.1 (Fig. 13b), and that of Kuo2U to KuoCL is around 1.1 or 1.2 (Fig. 14b). Those ratios seem to be independent of maximum wind speed.

6. Discussion

6.1 Validity of the experimental design

In the coupled atmosphere-ocean system, enhancement of greenhouse gases, including CO₂, causes gradual rise in SST on long time scales of decades or more. In response to SST warming, atmospheric variables (e.g., temperature, moisture and convective heating) reach a quasi-equilibrium state in significantly shorter time scales. In the present study, we focus on the quasi-equilibrium state of the atmosphere by the experimental design, in which an atmospheric GCM is used, and SST data are prescribed as an external forcing. This type of experimental designs is sometimes called a ‘time-slice’ experiment. Such experiments have been conducted by Bengtsson et al. (1996), Sugi et al. (2002) and others to obtain finer-resolution information about global warming under limited computing resources.

Landsea (1997) pointed out the possibility of inconsistency, between a low-resolution coupled...
model, from which warm SST data were taken, and the high-resolution atmospheric model, as a comment on the time-slice experiment of Bengtsson et al. (1996). According to Landsea, weakening of an intertropical convergence zone (ITCZ), and hydrological cycle in the high-resolution simulation of the warm climate, is an indication of the inconsistency, because such weakening is not seen in lower-resolution simulations of the coupled atmosphere-ocean system.

Here, we confirm that our results of the greenhouse warming experiments are consistent with those from lower-resolution coupled GCM experiments, (such as reported by IPCC, 2001). In Fig. 15, changes in potential temperature and equivalent potential temperature, are shown as the difference between AS2U and ASCL. Change in zonal mean precipitation is shown in Fig. 16. These figures indicate an increase in dry static stability (in the upper panel of Fig. 15), a decrease in moist static stability in the lower half of the tropical atmosphere (in the lower panel of Fig. 15), and an increase in precipitation in the latitudes where TCs are active (in Fig. 16). All of these results are generally consistent with lower-resolution experiments of global warming (e.g., IPCC 2001; Henderson-Sellers et al. 1998; Druyan et al. 1999). Possible influence of the increase in dry static stability, is discussed below in Section 6.4.

6.2 Robustness of the result

Significant reduction in total frequency of TSs and TDs is common to all of the global warming experiments in the present study. Robustness of this result is discussed in this subsection.

As shown in Section 5.3, there are no coherent changes in frequency of relatively intense TCs in the warm-climate experiments. The simulated changes in relatively intense
TCs (maximum surface wind $\geq 20$ m s$^{-1}$ or $\geq 25$ m s$^{-1}$) seem to depend on SST spatial patterns, and cumulus parameterizations.

The average altitude of the TC's warm-core maximum may not be constant under global warming. This suggests a possibility that the details of the warm-core criterion has some artificial effects on TC frequency changes in the warm-climate experiments. As sensitivity tests, three additional warm-core criteria are applied to the simulated TSs and TDs, in addition to the original criteria (Section 3.1), in the following cases:

“Case B”—Base case.

$$\Delta T_{\text{sum}} \equiv \Delta T_{300} + \Delta T_{500} + \Delta T_{700} \geq 2.5 \text{ K}$$

“Case U”—Upper troposphere weighted.

$$\Delta T'_{\text{sum}} = 1.5 \times \Delta T_{300} + 0.75 \times (\Delta T_{500} + \Delta T_{700}) \geq 2.5 \text{ K}$$

“Case L”—Lower troposphere weighted.

$$\Delta T''_{\text{sum}} = 0.5 \times \Delta T_{300} + 1.25 \times (\Delta T_{500} + \Delta T_{700}) \geq 2.5 \text{ K}$$

As shown in Table 3, the results demonstrate that percentage changes in the warm-climate experiments are very similar for each case. Therefore, the details of the warm-core criterion should have only minor influence on the TC frequency changes.

### 6.3 Comparison with other studies using high-resolution models

In this subsection, the present study is compared with two studies using atmospheric GCMs of the same horizontal resolution (Bengtsson et al. 1996; Sugi et al. 2002). The models, and global-warming impacts on global TS frequency simulated in these papers, are summarized in Table 4. In terms of global-scale change, all the results of the high-resolution numerical experiments in the three studies show reduction in TS formation frequency, in response to greenhouse warming.
The present study have the advantage of covering a wider range of SST patterns over the other two papers. As compared to the Sugi et al. study, the model has also been improved in terms of active TC formation.

Bengtsson et al. (1996) showed that, as a result of global warming, TS frequency decreased in all of the seven ocean basins they defined, although the changes were not statistically significant in some of the basins. Their result is similar to the regional variation of TS and TD frequency of AS2U and Kuo2U (see Fig. 9). The SST spatial pattern used for the greenhouse-warming experiment by Bengtsson et al. (1996) was somewhat similar to the uniform pattern used for AS2U and Kuo2U. In the results of the global warming experiment by Sugi et al. (2002), regional TS frequency decreased largely in the Western North Pacific and in the Eastern North Pacific, while TSs increased in the North Atlantic basin. For the results of AS2L, there was also significant reduction in TS and TD frequency over the western and eastern regions of the North Pacific, although significant changes are not seen in the North Atlantic basin (see Fig. 11). The SST spatial variation of the La Niña-like warming pattern (used for AS2L) is similar to that of the warming experiment by Sugi et al. (2002). Thus significant differences in regional-scale TC frequency could be largely attributed to the differences in SST patterns used in these numerical experiments.

Both of the warm-climate experiments by Bengtsson et al. (1996) and Sugi et al. (2002), showed larger reduction (more than 30%) in global TS frequency, with less SST warming (mostly less than 2 K increase in the tropics), as compared with the greenhouse-warming experiments of the present study. These quantitative differences could be attributed to the differences in model physics, including convective parameterizations and radiation processes, and also to some differences in SST patterns.

6.4 Influence of static stabilization of the atmosphere

The dry static stability in the tropical troposphere becomes significantly higher in response

<table>
<thead>
<tr>
<th>ALL</th>
<th>Case B</th>
<th>Case U</th>
<th>Case L</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASCL</td>
<td>153.4</td>
<td>135.6</td>
<td>135.6</td>
</tr>
<tr>
<td>KuoCL</td>
<td>170.0</td>
<td>143.0</td>
<td>143.6</td>
</tr>
</tbody>
</table>

Relative changes (%)

<table>
<thead>
<tr>
<th>ALL</th>
<th>Case B</th>
<th>Case U</th>
<th>Case L</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASCL</td>
<td>−17.2%</td>
<td>−17.5%</td>
<td>−17.5%</td>
</tr>
<tr>
<td>AS2U</td>
<td>−22.4%</td>
<td>−21.1%</td>
<td>−20.8%</td>
</tr>
<tr>
<td>ASCL</td>
<td>−21.6%</td>
<td>−21.1%</td>
<td>−20.8%</td>
</tr>
<tr>
<td>AS2E</td>
<td>−17.5%</td>
<td>−17.3%</td>
<td>−15.0%</td>
</tr>
</tbody>
</table>

Table 3. Annual-mean numbers of simulated TS and TD genesis and their changes due to global warming. Results from cases 'B', 'U', and 'L', based on additional criteria on the warm core, are shown. See text for details.

Table 4. Studies using high-resolution atmospheric GCMs.

<table>
<thead>
<tr>
<th>AGCM</th>
<th>Deep convection</th>
<th>Changes in global TS frequency in response to global warming</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bengtsson et al. (1996)</td>
<td>ECHAM3 (T106)</td>
<td>a mass flux scheme (Tiedtke 1989)</td>
</tr>
<tr>
<td>Sugi et al. (2002)</td>
<td>JMA GSM8911 (T106)</td>
<td>Kuo scheme</td>
</tr>
<tr>
<td>The present study</td>
<td>JMA GSM9603 (T106)</td>
<td>Arakawa-Schubert scheme</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kuo scheme</td>
</tr>
</tbody>
</table>

*Not statistically significant.
to greenhouse warming, because the environmental lapse rate is close to the moist-adiabatic lapse rate, which decreases with temperature. This effect is shown in Table 5 for vertical differences in potential temperature of our experiments, and also in the upper panel of Fig. 15 (as AS2U minus ASCL). Note that the middle troposphere warms more than near the surface, even though the prescribed changes were specified in the SST. Such stabilization is commonly seen in warming experiments with climate models (e.g., Henderson-Sellers et al. 1998).

The impact of static stabilization can be understood by the thermodynamic equation. In the tropics, where the horizontal temperature gradient is small, the approximate thermodynamic equation is written as

\[ \omega \frac{\partial \theta}{\partial p} \approx \frac{1}{T} \frac{Q}{C_p}, \]

where \( \omega \) and \( Q \) denote vertical \( p \)-velocity and diabatic heating, respectively (Holton 1979; Knutson and Manabe 1995). Knutson and Manabe (1995) showed, based on their experiments with GFDL coupled ocean-atmosphere GCM, that large-scale dynamical cooling, \( \omega \frac{\partial \theta}{\partial p} \), is enhanced by increasing static stability \( (\frac{\partial \theta}{\partial p}) \) without changing time-mean upward motion \( (\omega) \), over the western tropical Pacific in a global-warming condition.

Similar discussion can be given in the TC central region. Based on a heat budget analysis of a TC, in an axisymmetric atmospheric model, Kurihara (1975) has shown that, in the central region of the TC, the large heating due to condensation (and convection) is almost balanced by dynamical cooling, due to rising motion. Sugi et al. (2002) investigated frequency distribution of precipitation, upward motion (at 500 hPa) and vorticity (at 850 hPa), associated with disturbances in the tropics in their model outputs. They found that, for the same amount of precipitation, upward motion (and vorticity) of disturbances tends to be weaker in the global warming experiment, than in the control experiment. They attributed the weakening of disturbances to the increased dry static stability in the tropical atmosphere. They suggested that the reduction in global TC frequency, in their global warming experiment, is related to the weakening of the tropical disturbances, due to the static stabilization effect.

In the Section 5.4, we have found that precipitation is heavier (i.e., condensational heating should be enhanced) near TC centers in the warmer climates, if the maximum wind speed is the same. The thermodynamic equation indicates that if vertical velocity in the central regions of TCs is the same, the enhanced dynamical cooling, due to increased static stability, should be accompanied by the enhanced condensational heating, assuming the other heating terms are approximately constant.

### 7. Conclusions

In this study, we have conducted a series of numerical experiments using a T106 (equivalent to 110-km grid spacing) atmospheric GCM, in order to investigate the influences of SST spatial patterns, and cumulus parameterizations, on simulated TC frequency in global warming experiments. The results and conclusions are as follows:

1. Geographical distribution of TS formation is simulated realistically in the control experiments.
2. Impacts of El Niño/La Niña SST anomalies on TS frequency in the model also seems realistic.

<p>| Table 5. Difference in potential temperature between 500 hPa and the surface. Five-year-mean values, averaged between 30°N and 30°S, are shown. Anomalies from the control experiments are shown in the right column. |
|-----------------|-----------------|</p>
<table>
<thead>
<tr>
<th></th>
<th>( \Delta \theta ) (K)</th>
<th>Anomaly</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASCL</td>
<td>27.0</td>
<td>0</td>
</tr>
<tr>
<td>ASEN</td>
<td>27.2</td>
<td>0.2</td>
</tr>
<tr>
<td>ASLN</td>
<td>26.9</td>
<td>-0.1</td>
</tr>
<tr>
<td>AS2U</td>
<td>29.1</td>
<td>2.1</td>
</tr>
<tr>
<td>AS2E</td>
<td>29.4</td>
<td>2.4</td>
</tr>
<tr>
<td>AS2L</td>
<td>28.8</td>
<td>1.8</td>
</tr>
<tr>
<td>KuoCL</td>
<td>27.2</td>
<td>0</td>
</tr>
<tr>
<td>Kuo2U</td>
<td>29.0</td>
<td>1.8</td>
</tr>
</tbody>
</table>
The maximum wind speed and the minimum central pressures of intense TCs, are not realistically simulated in the model.

In the global warming experiments with the different SST patterns, and with the different cumulus schemes, frequency of TS formation decreases by 9.0–18.4% globally, and some of these changes are statistically significant. Total frequency of TSs and TDs decreases significantly in all of the warm-climate experiments. For relatively intense TCs (e.g., maximum surface wind $\geq 25$ m s$^{-1}$), there are no coherent changes in global frequency. These results suggest that global frequency of weak TCs, may decrease in the future warm climate, but frequency of intense TCs may either decrease or increase depending possibly on SST spatial variations.

Regional variation of TS and TD frequency in the global warming experiments depends largely on the SST spatial patterns. Therefore, in order to project regional change in TC frequency due to greenhouse warming, reliable projection of SST is understandably one of the crucial factors.

Mean precipitation near TC centers is heavier in the greenhouse-warming experiments than in the present-day-climate experiments, as compared for TCs with the same maximum wind speed. Since heavy rainfall near TCs sometimes causes severe floods, heavier precipitation could have important societal effects in a warmer climate.

Acknowledgements

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