Influence of the Global Warming on Tropical Cyclone Climatology: An Experiment with the JMA Global Model

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Abstract

The influence of the global warming on tropical cyclones has been examined using a high resolution AGCM. Two ten-year integrations were performed with the JMA global model at T106 horizontal resolution. For the control experiment, the observed SST for the period 1979–1988 is prescribed, while for the doubling CO2 (2 × CO2) experiment, SST anomaly due to the global warming estimated from a coupled model transient CO2 experiment (Tokioka et al. 1995) is added to the SST used in the control experiment.

The results of experiments show that a significant reduction in the frequency of tropical cyclones is possible in response to the greenhouse gas-induced global warming. The most significant decrease is indicated over the North Pacific. On the other hand, a considerable increase in tropical cyclone frequency is indicated for the North Atlantic. As for the maximum intensity of tropical cyclones, no significant change has been noted.

It has been found that the regional change in tropical cyclone frequency is closely related to the distribution of the SST anomaly, and the change in convective activity associated with it. The results of the experiment indicate that the change in tropical cyclogenesis is strongly controlled by dynamical factors associated with the change in SST distribution, rather than the thermodynamical factors associated with the change in absolute value of local SST.

On the other hand, for the decrease in the global total number of tropical cyclones on doubling CO2, a weakening of tropical circulation associated with the stabilization of the atmosphere (the increase in dry static stability), seems to be responsible. It is found that the rate of increase in the tropical precipitation due to the global warming is much less than the rate of increase in the atmospheric moisture. With this little increase in precipitation (convective heating), a considerable increase in the dry static stability of the atmosphere leads to a weakening of the tropical circulation.

1. Introduction

The influence on the tropical cyclone climatology of the global warming, due to the increase of carbon dioxide and other greenhouse gases, is not only a scientifically interesting problem but also of particular importance from the viewpoint of societal concern. From a simple thought, it is expected that the increase in atmospheric moisture due to the global warming may lead to an increase in the intensity, and frequency of tropical cyclones. Is this simple thought true? To answer this question, nu-
merous studies have been conducted. Emanuel (1987) suggested, based on his simple theoretical model, that the possible maximum intensity of a tropical cyclone might be increased by the global warming. Ryan (1992) and Royer et al. (1998) applied an empirical formula by Gray (1975) to the results of global warming experiments using GCMs. They found a significant increase in tropical cyclone frequency due to the global warming, but noted a limitation of applying an empirical formula based on the present climate to the double CO2 climate that is substantially different from the present climate.

An alternative approach to the problem is to utilize a GCM, although grid sizes of the most GCMs are too coarse to simulate tropical cyclones realistically. Broccoli and Manabe (1992) first conducted such an experiment using the GFDL model with two different resolutions (R15 and R30, corresponding grid sizes are 5° × 7.5° and 2.5° × 3.75°, respectively). In their experiment, they employed two different treatments of clouds: fixed and variable. The different treatment of clouds, regardless of the resolution of the model, gave an opposite answer to the problem: the model with fixed cloud has shown an increase, while the model with variable cloud has shown a decrease in the simulated tropical cyclone frequency on doubling CO2. Haarsma et al. (1993) has conducted a similar experiment using the UKMO GCM with 2.5° × 3.75° latitude-longitude grid. Their experiment has shown a significant increase in tropical cyclone frequency and intensity on doubling CO2.

There have been criticisms on the GCM experiments of the influence of the global warming on tropical cyclones (Evans 1992; Lighthill et al. 1994). They argue that the resolution of the GCMs is too coarse to simulate tropical cyclones properly, and the results of experiments are not physically acceptable. Recently, Bengtsson et al. (1996) reported the results of an experiment with a high resolution (T106, 1.1° × 1.1° grid) ECHAM model. In contrast to Haarsma et al. (1993), they found a significant reduction in the number of tropical cyclones on doubling CO2. Sugi et al. (1997) also conducted an experiment using a high resolution (T106, 1.1° × 1.1° grid) JMA model. They also found a significant decrease in the tropical cyclone frequency on doubling CO2. More recently, Yoshimura et al. (1999), May and Andersen (1999) and McDonald (1999) conducted experiments using high resolution models: JMA model, ECHAM model and UKMO model, respectively, to study the influence of the global warming on tropical cyclone climatology. An experiment with a high resolution regional climate model (Walsh and Ryan 2000), and an experiment using a very high resolution hurricane model (Knutson et al. 1998), were performed to study the possible change in tropical cyclone intensity due to the global warming.

In the present study, the influence of the global warming on tropical cyclones has been examined using a high resolution AGCM. After a brief description of the model and experiments in section 2, the results of the experiment briefly reported in Sugi et al. (1997) will be presented more in detail in section 3. The interpretation of the experimental results: the possible reasons for the reduction in the tropical cyclone frequency due to the global warming is discussed in section 4, followed by a summary in section 5.

2. Model and experiments

2.1 Model

The model used for this study is the T106 GCM version of the Japan Meteorological Agency (JMA) global model JMA-GSM8911 (JMA 1993; Sugi et al. 1990). The model is a spectral model with triangular truncation at total wave number 106. It has unevenly spaced 21 levels, the top of which is placed at 10 hPa. The model includes comprehensive physical processes: radiation (Sugi et al. 1990; Lacis and Hansen 1974), cumulus convection (Kuo 1974), large scale condensation, planetary boundary layer processes (Louis et al. 1982), simplified biosphere (SiB) (Sellers et al. 1986; Sato et al. 1989), and orographic gravity wave drag (Iwaski et al. 1989).

The simulated climate with the T42 version of the JMA global model is reported by Sugi et al. (1995a, b). Generally, the model simulates large-scale features of the climate reasonably well. One major deficiency of the model that is most relevant to the present study is the simulation of tropical precipitation. The model tends to underestimate precipitation over the western North Pacific in the summer season.
The sensitivity of the simulated climate to horizontal resolution, and cumulus parameterization scheme has been examined. It was found that large-scale features of the simulated climate generally is not very sensitive to horizontal resolution of the model, whereas the tropical precipitation of the model is much more sensitive to a cumulus parameterization scheme, than to horizontal resolution (Sugi et al. 1995c).

2.2 Experiments

A preliminary one-year integration of the model was first conducted to examine the ability of the model in simulating tropical cyclones. The integration was conducted for the period from April 1988 to March 1989, using the observed SST for the same period. Several realistic typhoons were simulated in this integration (Fig. 1). The lowest central pressure of the typhoons in the model is 980 hPa, maximum wind is 30 m s$^{-1}$, temperature anomaly of the warm core is 5°C, and the radius of the maximum wind is about 200 km. Such structure of the simulated typhoon is not sufficient, but much better than the simulation with lower resolution models (Sugi et al. 1994).

To simulate present tropical cyclone climatology, a ten-year integration of the model (control experiment) has been conducted using the AMIP observed SST for the period from 1979 to 1988. To examine the influence of global warming, another ten-year integration corresponding to the 2×CO2 climate (2×CO2 experiment) has been performed. For this purpose, the results of the transient CO2 experiment with MRI-CGCM (Tokioka et al. 1995) have been
used. First, the linear trend of the annual mean SST is computed to remove the natural decadal scale variation of SST in the transient CO2 experiment. Then, the increase of annual mean SST is computed for the period of 60–70 years of the transient CO2 experiment that corresponds to the period of doubling CO2. The SST increase due to the global warming thus computed (Fig. 2), together with the mean seasonal variation of SST increase during the period of 60–70 years, is added to the SST used in the control experiment. Using this SST and the doubled CO2 concentration in the atmosphere, the second integration has been conducted from the same atmospheric initial condition as the control experiment. A climatological sea ice distribution is used for both the control and $2\times CO2$ experiments. We believe that the influence of sea ice change due to the global warming may be negligible on the tropical cyclone climatology.

2.3 Criteria for selecting tropical cyclones

Because of the insufficient resolution of the model for simulating tropical cyclones, the structure and intensity of simulated tropical cyclones are different from the real tropical cyclones. Therefore, the criteria for selecting simulated tropical cyclones should be different from that of real tropical cyclones. Broccoli and Manabe (1992), Haarsma et al. (1993), Bengtsson (1995), Tsusui and Kasahara (1995) employed different criteria. We followed Bengtsson et al. (1995):

1) Sea level pressure: Candidate of tropical cyclone center is a grid point at which the sea level pressure takes local minimum, and the value is less than 1020 hPa.
2) Vorticity: Near the cyclone center, 850 hPa vorticity $\xi_{850} \geq 35 \times 10^{-6}\text{ sec}^{-1}$.
3) Maximum wind: Maximum wind speed at 850 hPa near the cyclone center $V_{max} \geq 15 \text{ m sec}^{-1}$.
4) Warm core: The average temperature difference from the area mean of surrounding region at 300 hPa, 500 hPa, 700 hPa and 850 hPa exceeds 3°C.
5) Upper level wind speed: Maximum wind speed at 300 hPa is less than the maximum wind speed at 850 hPa near the cyclone center.
6) Duration: Tropical cyclones should continue at least 2 days.

The criterion 5) effectively removes the extratropical cyclones (Bengtsson et al. 1995), and we do not need any prescribed geographical restriction for selecting tropical cyclones in
the model. The results of the simulation were stored every 24 hours. An automatic tracking program was developed to check the above criteria, and search a tropical cyclone near the location of the tropical cyclone found on the previous day.

3. Results

3.1 Control experiment

Figures 3(a) and 3(b) show the tracks of observed and simulated tropical cyclones for the 10-year period from 1979 to 1988. The total
The number of simulated tropical cyclones is 881 for the 10-year period, which is a little more than the observed number 812 in the same period (see Table 1). Overall geographical distribution of tropical cyclone is simulated well, although there are some discrepancies. In the model, tropical cyclones are simulated over the southeast coast of Brazil where no tropical cyclones are observed in reality. Tropical cyclones are also simulated but not observed around the Malay peninsula. From Fig. 3(a) and 3(b), we see that the density of simulated tracks is less than the observation over the North Pacific, and North Atlantic. In the North Pacific, the number of simulated tropical cyclones is considerably less than the observed (see Table 1). Moreover, the simulated tracks are generally shorter than those of the observed. The average lifetime of the simulated tropical cyclones is 3.4 days, which is considerably shorter than 5.2 days, the average observed lifetime of typhoons.

Table 1 shows the number of observed (column OBS), and simulated (column CNTL), tropical cyclones in each ocean basin. The number of simulated tropical cyclones in the North West Pacific are considerably less than the observed number, while those in the North Indian Ocean and South West Pacific Ocean are much larger than those observed. About 18% less tropical cyclones are simulated than the observation in the Northern Hemisphere, while significantly larger number of tropical cyclones are simulated than the observation in the Southern Hemisphere. Note that the difference in the numbers of simulated, and observed tropical cyclones in some ocean basin is larger than the standard deviation (shown in the parenthesis) of the interannual variation of the number of simulated or observed tropical cyclones, and therefore the difference is statistically significant. The difference in the global total numbers of simulated and observed tropical cyclones, is less than the corresponding standard deviations.

The seasonal variation of the number of simulated and observed tropical cyclones in each
ocean basin is plotted in Fig. 4. As noted above, the total number of simulated tropical cyclones is much less than those observed in the Northern Hemisphere, while it is much larger than those observed in the Southern Hemisphere.

The model fails to simulate the late summer peak in the Northern Hemisphere, while it simulates too many tropical cyclones in the Southern Hemisphere winter and spring seasons. In the North West Pacific, the maximum frequencies of tropical cyclones are simulated in October and November, two months later than the observation. The value of the peak frequency is significantly less in the North West Pacific, North Atlantic and North Eastern Pacific. Therefore, the amplitude of seasonal variation of simulated tropical cyclone frequency in the Northern Hemisphere, is much less than that observed. The amplitude of seasonal variation of the simulated tropical cyclone is also less than the observation in the Southern Hemisphere, because too many tropical cyclones are simulated from April to December in the South West Pacific. In the North Indian Ocean, the double peaks in May and November are simulated well, although the number of simulated tropical cyclones is considerably larger than the observed number.

Figure 5 shows the inter-annual variation of the global total number of simulated and observed tropical cyclones during the 10-year period. The standard deviation of the inter-annual variation of simulated tropical cyclone frequency is 10.4, and comparable to the observed standard deviation 12.3 (see Table 1). It should be noted that the similar magnitude of inter-annual variation of tropical cyclone frequency has been simulated with a GCM using a climatological SST as a lower boundary condition (Bengtsson et al. 1995; Tsutsui et al. 1996). This suggests that the major portion of inter-annual variation of tropical cyclone frequency is not forced by inter-annual variation of SST. Therefore, it is not so unreasonable that the correlation between the simulated and observed inter-annual variation of tropical cyclone frequency is weak (0.15), although an observed SST is used as a lower boundary condition in the present study.

Figure 6 shows the frequency distributions of simulated, and observed tropical cyclones as a function of maximum wind speed for each one. The frequency distribution of simulated tropical cyclones exhibits a sharp peak at 20–25 m s\(^{-1}\). The largest maximum wind speed of simulated tropical cyclones is 40–50 m s\(^{-1}\). On the other hand, the frequency distribution of observed tropical cyclones has a weak peak at 20–25 m s\(^{-1}\), and extends over a broad range of maximum wind speed up to 80–85 m s\(^{-1}\). It is clear from the figure that the model fails to simulate very intense storms. This indicates that the horizontal resolution of the model is still too coarse to simulate tropical cyclones very realistically. To summarize the results of control experiment; it may be said that the model generally simulates the observed tropical cyclone climatology well, although there are some deficiencies in the simulation of geographical distribution, seasonal variation and intensities of tropical cyclones. The reason for the discrepancy between the observed and simulated tropical cyclone climatology may be an important subject. Generally speaking, it may be related to deficiencies of the model in simulating tropical convection and tropical cyclone development. The analysis of more specific reason for each discrepancy is left for a future study.

### 3.2 2 × CO2 experiment

Figure 3(c) shows the tracks of tropical cyclones simulated in the 2 × CO2 experiment. Compared with the result of the control experiment shown in Fig 3(b), we see a significant reduction in the number of simulated tropical cyclones over the North West Pacific and North East Pacific Oceans. In contrast, over the North Atlantic Ocean, we can see more tropical cyclones simulated in the 2 × CO2 experiment than the control experiment. The number of tropical cyclones simulated in the 2 × CO2 experiment is shown in the third column of Table 1. In the Table 1, also shown are the difference (2 × CO2 – control) and the ratio (2 × CO2/control) of the numbers of tropical cyclones in the two experiments. Most significant reduction is seen over the North West Pacific, where the number of simulated tropical cyclones in the 2 × CO2 experiment is 7.5 per year, which is one-third of the number for the control experiment (21.9). A similar reduction rate is seen over the North East Pacific, but it is not statistically significant because of the large inter-
Fig. 4. Seasonal variation of tropical cyclone frequency by ocean basin. Dotted lines indicate the observed frequency. (a) Northern Hemisphere, (b) North Indian Ocean, (c) North Western Pacific, (d) North Atlantic and North East Pacific, (e) Southern Hemisphere, (f) South Indian Ocean, (g) South West Pacific, and (h) South Atlantic and South East Pacific (see Table 1 for the definition of regions). Solid and dashed lines show the frequencies of simulated tropical cyclones in the control experiment and the $2 \times$ CO2 experiment, respectively.
annual variation over this region in the control experiment. The South Indian Ocean also shows a substantial reduction in the tropical cyclone frequency on doubling CO2. On the other hand, over the North Atlantic Ocean, we see a significant increase in the number of tropical cyclones, from 7.4 per year in the control experiment to 11.4 per year in the 2×CO2 experiment, which is more than 60% increase. Over the North Indian Ocean, there is a slight increase, although it is not statistically significant. From Table 1, we note that a considerable reduction in the tropical cyclone frequency over the South Pacific Ocean on doubling CO2. Figure 3 shows that this reduction mostly occurs in the northern part of the South Pacific Ocean. In the southern part of the South Pacific Ocean to the east of Australia, we can see more tropical cyclones simulated in the 2×CO2 experiment than in the control experiment. This seems to be associated with the southward shift of the region of tropical cyclogenesis (probably SPCZ).

In Fig. 4, the seasonal variation of tropical cyclone frequency over each ocean basin simulated in the 2×CO2 experiment is shown by dashed curves. In the North West Pacific, the reduction in the tropical cyclone frequency in the 2×CO2 experiment is seen in all seasons. In the North Atlantic and North East Pacific, more tropical cyclones are simulated in the 2×CO2 experiment than the control experiment in late autumn (October and November), while less tropical cyclones are simulated in early summer (June and July). In the North Indian Ocean, the peak in May is more prominent in the 2×CO2 experiment.

The inter-annual variation of the number of tropical cyclones simulated in the 2×CO2 experiment is shown by long-dashed curve in Fig. 5. We can see a clear separation of this curve from the solid curve (control experiment), and the short-dashed curve (observed), although the maximum number occurred in the year 1985 of the 2×CO2 experiment is slightly larger than the minimum number occurred in 1986 of the control experiment. The global total number of simulated tropical cyclone in the 2×CO2 ex-

![Fig. 5. Inter-annual variation of tropical cyclone frequency. Dotted curve indicates the observed frequency based on the US Navy Best Track Dataset. Solid and dashed lines show the frequencies of simulated tropical cyclones in the control experiment and the 2×CO2 experiment, respectively.](image)

![Fig. 6. Tropical cyclone frequency distribution as a function of maximum wind speed of each cyclone. (a) Raw frequency during the ten year period. (b) Normalized frequency. Dotted line indicates the observed frequency. Solid and dashed lines show the frequencies of simulated tropical cyclones in the control experiment and the 2×CO2 experiment, respectively.](image)
periment is 58.5 per year, which is about 34% less than the control experiment (see Table 1). The standard deviation of the inter-annual variation of the number of tropical cyclones simulated in the $2 \times CO_2$ experiment is 7.7, which is a little less than that of the control experiment. The difference in the global total number of tropical cyclones simulated in the two experiments is 29.6 per year, which is three times larger as the standard deviation, and therefore, the difference is highly statistically significant.

In Fig. 6, the frequency distribution as a function of maximum wind speed for the $2 \times CO_2$ experiment is plotted by a dotted curve. From Fig. 6(a), we can see the overall reduction in the frequency of tropical cyclones on doubling CO2. From the normalized frequency distribution curves in Fig. 6(b), it is noted that the relative frequency distributions are very similar for the two experiments. There is a slight increase in the normalized frequency on doubling CO2 at the region of maximum wind speed exceeding 40 m s$^{-1}$, but it is not statistically significant. From Fig. 6, we may conclude that there is no significant change in the intensity of tropical cyclones on doubling CO2.

From Fig. 6, we note that the simulated tropical cyclones are much weaker than the observed, and this raises a concern about the selection criteria for simulated tropical cyclones. If we employ different criteria, would the results be different? Figure 7 shows the frequency distribution of tropical disturbances as a function of intensity of the vorticity at 850 hPa. Here, the center of a “tropical disturbance” is defined as a grid point where the 850 hPa vorticity takes local maximum value on the $2.5^\circ \times 2.5^\circ$ latitude-longitude grid between 30$^\circ$N and 30$^\circ$S. For the entire tropical belt and annual average case, the frequency distribution is plotted in Fig. 7(a). In this case, the frequency of tropical disturbances, with vorticity intensity exceeding $23 \times 10^{-6}$ sec$^{-1}$, is less in the $2 \times CO_2$ experiment than in the control experiment. Note that the selection criteria used in this study employs $35 \times 10^{-6}$ sec$^{-1}$ as a threshold value of 850 hPa vorticity intensity. From Fig. 7(a) we can see that a change in this threshold value would not affect the conclusion that the number of tropical cyclone simulated in the $2 \times CO_2$ experiment, is significantly less than the control experiment.

Figures 7(b) and Fig. 7(c) show the frequency distributions of tropical disturbances over the North West Pacific Ocean, and the North At-
Atlantic Ocean, in the autumn season (September–November). In the Pacific Ocean case, we see a significant reduction on doubling CO2 in the frequency of tropical disturbances with vorticity larger than $30 \times 10^{-6}$ sec$^{-1}$. In contrast, in the Atlantic Ocean case, an increase in the frequency on doubling CO2 is seen over a wide range of vorticity intensity. Thus, the change in frequency distribution of vorticity over the North West Pacific, and North Atlantic, is in agreement with the change in tropical cyclone frequency over these regions shown in Fig. 4 and Table 1. It may be said, therefore, that the regional changes in tropical cyclone frequency also would not be much affected by the value of vorticity criteria for tropical cyclones.

4. Discussion

The results of experiments indicate that the number of tropical cyclones may significantly be reduced due to the global warming. On the other hand, there is not a clear evidence showing significant change in the intensity of tropical cyclones. In this section, we will examine possible reasons for the reduction in tropical cyclone frequencies on doubling CO2. First, we discuss the regional change, and then the global change in the tropical cyclone frequencies.

4.1 Regional change

From Fig. 3, we see the most significant reduction in the number of tropical cyclones over the North Pacific Ocean, while we see a considerable increase over the North Atlantic Ocean. From Fig. 2, we note that the SST anomaly given to the $2 \times \text{CO}_2$ experiment is relatively small over the central part of the Pacific Ocean compared with the other regions of tropical ocean. Also we note that the SST anomaly is large over the Atlantic Ocean. We can see that the regions where the reductions in the tropical cyclone frequency occur in Fig. 3 agree with the regions where the SST anomalies are less than 1°C (lightly shaded regions). On the other hand, over the regions where the SST anomalies are larger than 1°C (medium dark areas), more tropical cyclones are simulated in the $2 \times \text{CO}_2$ experiment than the control experiment. The agreement between the SST anomaly pattern and the tropical cyclone frequency change pattern indicates that the anomalous tropical circulation associated with the SST anomaly play an important role in the tropical cyclogenesis. It should be noted that the SST anomaly over the Pacific Ocean is positive, though it is less than the other regions. Therefore, if the absolute value of SST is a dominant factor for the tropical cyclogenesis, more tropical cyclones should be simulated in the $2 \times \text{CO}_2$ experiment than the control experiment even over the North Pacific Ocean. The result of experiment indicates that it is not the absolute value of SST that dominates the tropical cyclogenesis. Rather, a tropical circulation pattern associated with the SST distribution is a dominant factor for the regional tropical cyclogenesis.

Figure 8(a) shows the difference between the annual mean precipitation rate of the control experiment, and that of the $2 \times \text{CO}_2$ experiment, while Fig. 8(b) shows the SST anomaly (minus 1°C) given to the $2 \times \text{CO}_2$ experiment. From Fig. 8(a) and 8(b), we can see a good agreement between the pattern of precipitation difference, and the pattern of SST anomaly. As we have noted above, the pattern of tropical cyclone frequency change from control experiment to $2 \times \text{CO}_2$ experiment resembles the pattern of SST anomaly. Thus, the patterns of tropical cyclone frequency change, precipitation change and SST anomaly resemble each other. This suggests a chain of links from tropical SST anomaly to convective activity (or precipitation), from convective activity to tropical circulation, and from tropical circulation to tropical cyclogenesis.

To understand the regional changes in the convective activity in the tropics, the changes in the distribution of the mean fields of quantities associated with tropical convection are examined. Figure 9 shows the differences ($2 \times \text{CO}_2$ – control) in the annual mean vertical $p$-velocity at 500 hPa, surface specific humidity, dry static stability and convective available potential energy (CAPE). In Fig. 9(a), anomalous upward motions are seen over the Atlantic ocean and the region extending south-eastward from North Indian Ocean, to the western south Pacific Ocean. Over these regions, anomalous convection is positive in Fig. 8(a). On the other hand, anomalous downward motions are seen over the most part of the Pacific Ocean and South Indian Ocean, where anomalous convection is negative. Figure 9(b) shows that the
surface specific humidity over the ocean is 5–15% larger in the $2 \times CO_2$ experiment than in the control experiment. Although the difference in the surface specific humidity is positive everywhere, a large difference of more than 10% is seen over the regions where the anomalous convection is positive in Fig. 8 (a).

The change in dry static stability is shown in Fig. 9 (c). Here the dry static stability is defined as the difference in potential temperature at 200 hPa and 1000 hPa. The figure shows that the dry static stability is 6–8% larger in the $2 \times CO_2$ experiment than the control experiment over the most part of the tropical belt. It should be noted that the increase in the dry static stability is closely related to the increase in the moist adiabatic lapse rate in the warmer and wetter atmosphere. In contrast to the rather uniform increase in the dry static stability, the change in the convective instability as measured by CAPE shows distinct regional distribution (Fig. 9(d)). It increases over the regions where the anomalous convection is positive in Fig 8a. It should be noted that the CAPE decreases over the most part of the Pacific Ocean, and South Indian Ocean, although the surface moisture, and therefore, the moist static energy of the surface air increases everywhere in the tropics. The reason for the decrease in CAPE over these regions is that the reduction in CAPE due to the increased dry static stability is larger than the increase in CAPE due to the increase of moist static energy of surface air. To further confirm this idea, the vertical profiles of potential temperature and moist static energy at the grid point in the western Pacific, where the change in CAPE due to the global warming is close to zero, are shown in Fig. 10. The potential temperature difference is about one degree at 1000 hPa and it increases with height up to 200 hPa, where the difference is about four degrees. On the other hand, the difference in the saturated moist static energy is nearly the same at all levels between the surface and 200 hPa. The magnitude of the difference is almost the same as the difference in the moist static energy at the surface. Thus, the difference in the CAPE at this grid point is close to zero. At the other grid point where the CPAE increases (decreases) due to the global warming, the effect of moist static energy increase at the surface is larger (less) than the
effect of dry static energy increases in the upper levels.

In order to understand the regional change in the tropical cyclone frequency in response to the CO2 increase, yearly genesis parameter (YGP) proposed by Gray (1975) is examined. Gray identified six factors which contribute to tropical cyclogenesis. They are three dynamical factors: Coriolis parameter, low level relative vorticity and vertical shear of horizontal wind; and three thermodynamical factors: thermal energy of ocean (or SST), moist instability of lower troposphere and relative humidity. Gray combined these six factors into a single param-

Fig. 9. Differences in the annual mean quantities related to tropical circulation between the control experiment and the $2 \times \text{CO}_2$ experiment. (a) vertical $p$-velocity at 500 hPa (hPa/h), (b) surface specific humidity (difference in %), (c) dry static stability (difference in %), (d) CAPE (J kg$^{-1}$).
eter called YGP, which accounts for the climatological geographical distribution of tropical cyclogenesis fairly well.

To find out which factor most contributes to the change in tropical cyclone frequency, YGP is computed for the control experiment, and the 2 × CO2 experiment. The difference in the YGP of the two experiments is shown in Fig. 11 together with the contributions to the YGP change from the five factors constituting the YGP (no contribution from Coriolis factors which does not change on doubling CO2). Since the YGP is a sum of four seasonal genesis parameters (SGP), the contribution from each factor to the SGP change is calculated first by calculating the SGP change associated with the change in the respective factor. Although the second order terms are neglected in this calculation of SGP change, the sum of contributions from the five factors (Fig. 11(b)–(f)) calculated in this way is nearly the same as the total change in the YGP (Fig. 11(a)).

From Fig. 11(a), it is seen that the change in the YGP is positive over most regions, except for the western central Pacific. The total number of tropical cyclones estimated from YGP is 83.5 per year for the control experiment, and 130.4 per year for the 2 × CO2 experiment. The change in YGP indicates a 56% increase in the number of tropical cyclone due to the CO2 increase, in spite of the 34% decrease in the number of simulated tropical cyclones. The YGP clearly overestimates the number of tropical cyclones in the 2 × CO2 experiment. This overestimation is mainly due to the ocean thermal energy factor (SST factor) as shown in Fig. 11(d).

It is interesting to note that the changes in the two dynamical factors, low level relative vorticity (Fig. 11(b)) and vertical shear of horizontal wind (c), contribute to the reduction of the YGP, particularly over the Pacific. This indicates that the dynamical factors play important roles in the significant reduction in the simulated tropical cyclone frequency, particularly over the Pacific. On the other hand, the three thermodynamical factors, SST (d), moist instability (e) and relative humidity (f), mostly contribute to the increase of the YGP. Positive SST anomalies are considered to be favorable for tropical cyclogenesis as indicated by the thermodynamical factors in the YGP. However, relatively little SST anomaly compared with the surrounding regions, even though it is positive, would lead to a circulation that is unfavorable for tropical cyclogenesis. The results of the experiment indicate that the change in tropical cyclogenesis is strongly controlled by dynamical

Fig. 10. Annual mean vertical profile of (a) potential temperature (K), and (b) moist static energy \((10^3 \text{ J kg}^{-1})\) at 10°N, 160°E. Solid and dashed lines indicate the control experiment and 2 × CO2 experiments. Thick lines in (b) show the saturated moist static energy.
factors associated with the change in SST distribution, rather than the thermodynamical factors associated with the change in absolute value of local SST.

In summary, the changes in tropical circulation associated with the SST anomaly distribution play an important role in the regional change of tropical cyclone frequency due to the global warming. In the previous section we have noted that the major portion of inter-annual variation of the simulated tropical cyclone frequency is not forced by SST variation. However, this does not exclude a possibility of an influence of SST anomaly on regional tropical cyclogenesis. It should be noted that the SST anomaly used in the $2 \times \text{CO}_2$ experiment is almost the same pattern throughout the 10 year period of the experiment, and therefore, very persistent. Such persistent SST anomaly has never been observed in the past. In this regard, it is interesting to note that the number of typhoons in 1998 summer is only 4, only one third of the normal year. The SST anomaly pattern in 1998 summer was persistent, and similar to the pattern of that of the $2 \times \text{CO}_2$ experiment. Therefore, a significant reduction in typhoon frequency indicated by the present study may not be unrealistic, if the SST anomaly is reasonable.

4.2 Global change

In the previous subsection, it has been shown that the regional change in the tropical cyclone frequency due to the global warming is closely related to the distribution of SST increase. Over the regions with relatively large SST increase, convective activity and tropical cyclone frequency also increase. On the other hand, the

Fig. 11. (a) Difference in the YGP between the control experiment and $2 \times \text{CO}_2$ experiment. (b)–(f) Contributions to the difference in YGP from the changes in constituting factors: (b) relative vorticity, (c) vertical shear of horizontal wind.
regions with relatively small SST increase, convective activity and tropical cyclone frequency decrease, even though the SST does increase. The result of the experiment shows a significant reduction in the global tropical cyclone frequency due to the global warming. Does this correspond to an overall weakening of convective activity?

To examine the change in overall convective activity, zonal mean precipitation and evaporation for the control experiment and the 2×CO2 experiment are shown in Fig. 12. Both the global average precipitation and evaporation increase about 2% due to the global warming. Both of them show larger increase in the extratropics. The precipitation and evaporation in the tropics increase 1% and 1.4%, respectively (Table 2). In the Southern Hemisphere tropics, the zonally averaged precipitation of the 2×CO2 experiment is less than that of the control experiment. On the other hand, precipitable water in the tropical atmosphere increases as much as 14% due to the global warming (Table 2). Why does the tropical precipitation increase only 1% in spite of the 14% increase of precipitable water? If the tropical circulation does not change on doubling CO2, the precipitation should increase with increasing atmospheric moisture. That the precipitation does not much increase in spite of a

Fig. 11. (d) ocean thermal energy (SST), (e) moist instability, (f) relative humidity.
significantly increased atmospheric moisture suggests a weakening of tropical circulation on doubling CO2.

To examine the change in the tropical circulation, mass flux at 500 hPa is computed (Table 3). The upward (downward) flux in the Table 3 is the vertical p-velocity averaged over the upward (downward) motion region within the tropics (30°N–30°S) multiplied by the fractional area. The instantaneous mass flux is computed using snapshots with 24 hour interval. Since upward motion and downward motion take place at the same grid point at different time, the magnitude of mass fluxes calculated from time averaged vertical velocity would be smaller than that calculated from the instantaneous vertical velocity. Indeed, the mass fluxes calculated from the monthly mean vertical velocity or annual mean vertical velocity in Table 3 are smaller than the instantaneous mass fluxes. By a similar reason, the intensity of spatially averaged circulation such as (zonally averaged) Hadley circulation or (meridionally averaged) Walker circulation, which are often used as a measure of tropical circulation intensity, would underestimate the intensity of tropical circulation by cancellation of upward motion and downward motion. Here we take the instantaneous mass flux as an index of tropical circulation intensity. As the tropical region (30°N–30°S) considered here is not a closed region, magnitudes of the upward mass flux and downward mass flux are not exactly the same. As the upward motion area of the

Table 2. Variables related to tropical circulation

<table>
<thead>
<tr>
<th>Variables</th>
<th>CNTL</th>
<th>2xCO2</th>
<th>Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation</td>
<td>3.43</td>
<td>3.47</td>
<td>+1.0</td>
</tr>
<tr>
<td>Evaporation</td>
<td>3.78</td>
<td>3.83</td>
<td>+1.4</td>
</tr>
<tr>
<td>Radiative Cooling</td>
<td>101.0</td>
<td>102.0</td>
<td>+1.0</td>
</tr>
<tr>
<td>Stability</td>
<td>47.2</td>
<td>49.8</td>
<td>+5.5</td>
</tr>
<tr>
<td>Mass flux</td>
<td>1.429</td>
<td>1.337</td>
<td>-6.4</td>
</tr>
<tr>
<td>Precipitable Water</td>
<td>3.35</td>
<td>3.81</td>
<td>+13.7</td>
</tr>
</tbody>
</table>
Table 3. Tropical mass fluxes

<table>
<thead>
<tr>
<th></th>
<th>CNTL</th>
<th></th>
<th>2xCO2</th>
<th></th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mass flux</td>
<td>Fractional area</td>
<td>Mass flux</td>
<td>Fractional area</td>
<td>Mass flux</td>
</tr>
<tr>
<td>Instantaneous</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upward</td>
<td>1.429</td>
<td>0.438</td>
<td>1.337</td>
<td>0.440</td>
<td>- 6.4</td>
</tr>
<tr>
<td>Downward</td>
<td>1.403</td>
<td>0.562</td>
<td>1.284</td>
<td>0.560</td>
<td>- 8.5</td>
</tr>
<tr>
<td>Monthly mean</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upward</td>
<td>0.697</td>
<td>0.438</td>
<td>0.672</td>
<td>0.438</td>
<td>- 3.6</td>
</tr>
<tr>
<td>Downward</td>
<td>0.672</td>
<td>0.562</td>
<td>0.638</td>
<td>0.562</td>
<td>- 5.1</td>
</tr>
<tr>
<td>Annual mean</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upward</td>
<td>0.497</td>
<td>0.462</td>
<td>0.487</td>
<td>0.473</td>
<td>- 2.0</td>
</tr>
<tr>
<td>Downward</td>
<td>0.472</td>
<td>0.538</td>
<td>0.453</td>
<td>0.527</td>
<td>- 4.0</td>
</tr>
</tbody>
</table>

The wakening of tropical circulation due to global warming may be understood by energy balance of the tropical atmosphere. The approximate energy equation in the tropics may be expressed as,

\[
\frac{\partial \theta}{\partial p} \approx \frac{\theta}{T \ C_p} \left( Q_c - Q_R \right) A_u, \tag{1}
\]

(Holton 1979; Kuntson and Manabe 1995). Here we consider a schematic tropical circulation as shown in Fig. 13. Then, the energy balance in the upward motion region and downward motion region at 500 hPa may be written as,

\[
M_u \frac{\partial \theta}{\partial p} \approx \frac{\theta}{T \ C_p} \left( Q_c - Q_R \right) A_u, \tag{2}
\]

\[
M_d \frac{\partial \theta}{\partial p} \approx \frac{\theta}{T \ C_p} Q_R A_d, \tag{3}
\]

where \(M_u(M_d)\) and \(A_u(A_d)\) are the mass flux, and fractional area of the upward (downward) motion area, \(Q_c\) and \(Q_R\) are the condensation heating rate and radiative cooling rate, respectively. The change in condensation heating rate due to the global warming may be proportional to the change in precipitation rate. Therefore, from Table 2, the \(Q_c\) would increase about 1% due to the global warming. We note that the
radiative cooling rate also increases about 1%. On the other hand, mass flux is about 6% less in the $2 \times CO_2$ experiment than the control experiment, while the atmospheric stability (defined as the potential temperature difference at 200 hPa and 1000 hPa) is about 6% larger in the $2 \times CO_2$ experiment. We note that the dry static stability is nearly uniform in the tropics, and we assumed for simplicity that the change in dry static stability over the upward motion region is the same as that over the downward motion region. Thus, we see that the increase in dry static stability and the decrease in mass flux nearly cancel each other and balance with the small increase in the heating terms.

In summary, there is a balance as illustrated in Fig. 14. The tropical atmosphere becomes more stable (dry static stability increases) in response to the global warming, while the condensation heating and radiative cooling, which are the driving force of tropical circulation, increase only a little. Hence, the tropical circulation weakens due to the global warming. On the other hand, as the atmospheric moisture in the tropics significantly increases with the global warming, condensation heating can increase even with weaker circulation. In other words, in the warmer tropical atmosphere with more moisture and larger dry static stability, the energy balance is achieved by a weakening of the atmospheric circulation.

A weakening of tropical circulation due to the global warming has been reported in several GCM studies. Bengtsson et al. (1996) pointed out the weakening of Hadley cell and tropical hydrological cycle on doubling CO2 in their experiment. Knutson and Manabe (1995) reported that the intensity of the upward motion over the tropical western Pacific slightly decreases when the atmospheric CO2 concentration is quadrupled, although the precipitation over the same region is significantly enhanced. Kitoh et al. (1996) has shown that the intensity of the south-westerly monsoon flow becomes weaker on doubling CO2, although the monsoon precipitation increases.

Next, we examine the influence of the global warming upon individual tropical disturbances. Figure 15 shows the frequency distribution of tropical disturbances as a function of precipitation intensity, vertical $p$-velocity at 500 hPa and vorticity at 850 hPa. Here, we have selected the grid points in the tropics ($30^\circ$N–$30^\circ$S) where the 850 hPa vorticity takes local maximum value, and we regard such grid points as the centers of tropical disturbances. For the values of precipitation and vertical velocity of a disturbance, we take the maximum value within the two grid distances (~500 km) from the center of the disturbance. The frequency distributions with respect to precipitation intensity are almost the same for control and $2 \times CO_2$ experiments. On the other hand, in the frequency distribution as a function of vertical velocity and vorticity, we can see a tendency that the frequency of tropical disturbances with strong (weak) vertical velocity or vorticity decreases (increases) on doubling CO2.

To further confirm this tendency, two dimensional frequency distributions of disturbances as a function of precipitation, vertical $p$-velocity at 500 hPa and vorticity at 850 hPa are shown in Fig. 16. The top panels show the frequency distribution with respect to precipitation and vertical velocity for the control and $2 \times CO_2$ experiments, and the difference between the two experiments. The pattern of distribution indicates a positive correlation between precipitation and vertical velocity of disturbances. It should be noted that both the axes are scaled in logarithm. Therefore, for a certain value of precipitation (vertical velocity), the values of vertical velocity (precipitation) extend over a considerably wide range. In the left panel, the distributions for control and $2 \times CO_2$ experiments mostly overlap. The difference between the two experiments on the right panel indicates that for the same value of precipitation the
vertical velocity of disturbances tends to be weaker in the $2 \times CO_2$ experiment than in the control experiment.

On the other hand, as shown in the middle panels, for the frequency distribution with respect to vertical velocity and vorticity, the difference between the two experiments extends along the 45 degree sloped line. Both the vertical velocity, and vorticity tend to be weaker in the $2 \times CO_2$ experiment than the control experiment. Thus the relationship between vorticity and vertical velocity is rather invariant on doubling CO2. As expected from the above results, for the same amount of precipitation, the vorticity of disturbances tends to be weaker in the $2 \times CO_2$ experiment than in the control experiment, as shown in the bottom panels. These results may be closely related to the stabilization of tropical atmosphere on doubling CO2. The equation (1) indicates that for the same amount of heating (precipitation), vertical velocity should be weaker in the stabilized atmosphere. As in the discussion for the energy balance of the tropical circulation, the condensation heating associated with a tropical disturbance would not necessarily decrease with weakening of the vertical velocity because of the atmospheric moisture increase on doubling CO2.

### 4.3 Inter-model difference

As mentioned in the introduction, there have been several studies using GCMs on the influence of global warming on tropical cyclone climatology. Some of them have shown an increase in global total number of tropical cyclones due to the global warming, while the others have shown a decrease. Regional change in the tropical cyclone frequency is also different from model to model. The reason for such differences among the models will be discussed in this subsection.

According to the discussion in the previous subsection, there are two opposing factors for the change in global tropical cyclone frequency: reduction due to the stabilization of the atmosphere (increase in dry static stability), and increase due to the enhanced precipitation (condensation heating). In the experiment of the present study, the increase in the tropical precipitation is not significant and the increase in dry static stability of the atmosphere is dominant, leading to a significant reduction in the
Fig. 16. Two dimensional normalized frequency distribution of tropical disturbances (%) as a function of (a) precipitation and vertical p-velocity at 500 hPa, (b) vertical p-velocity at 500 hPa and vorticity at 850 hPa, and (c) precipitation and vorticity at 850 hPa. Solid and dashed curves indicate the control experiment and 2×CO2 experiment, respectively. (d)–(f): differences between the control experiment and the 2×CO2 experiment.
global total number of tropical cyclones in the $2 \times CO2$ experiment. However, the amount of increase in tropical precipitation on doubling CO2 is very much different from model to model. It takes values between $-1\%$ and $5\%$ according to our calculation using the data from eight models collected at IPCC Data Distribution Center. In the present study, the increase in tropical precipitation is $1\%$ and relatively small compared with other models. It may be possible that the global frequency of tropical cyclone show less decrease or even increase in the model with more tropical precipitation increase on doubling CO2. It is noted that the amount of increase in tropical precipitation is well balanced with the amount of increase in the radiative cooling in the present study (Table 2). It is suggested that the change in the tropical precipitation due to the global warming is closely related to the change in the radiative cooling, and therefore the change in clouds. In this regard, it is interesting to note that the results of experiments by Broccoli and Manabe (1990) differ by different treatment of clouds in the same model. They found an increase of tropical cyclone frequency in the experiment with prescribed climatological clouds, while a decrease in the experiment with variable clouds.

As for the difference among the models in the regional change of tropical cyclone frequency, the difference in the distribution of SST anomaly used in the experiments seems to be a major factor. It is known that there are large intermodel differences in the SST anomaly distribution among the results of global warming experiments with coupled atmosphere-ocean GCMs (Noda et al. 1999). In section 4.1, it has been noted that the regions where the SST increase due to the global warming is relatively large (small) coincide with the regions where the convective activity (precipitation) and tropical cyclone frequency are increased (decreased).

From above discussion it is indicated that a reliable estimate of the distribution of SST change and a quantitative estimate of tropical precipitation change are essential for a reliable prediction of possible change in tropical cyclone frequency and distribution.

5. Summary

In the present study, the influence of the global warming on tropical cyclones has been examined using a high resolution AGCM. Two ten-year integrations were performed with the JMA global model at T106 horizontal resolution. For the control experiment, the observed SST for the period 1979–1988 is prescribed, while for the $2 \times CO2$ experiment, SST anomaly due to the global warming estimated from a coupled model transient CO2 experiment (Tokioka et al. 1995), is added to the SST used in the control experiment.

The results of experiments show that a significant decrease in the frequency of tropical cyclones is possible in response to the greenhouse gas-induced global warming. Most significant decrease is indicated over the North Pacific. On the other hand, a considerable increase in tropical cyclone frequency is indicated for the North Atlantic. As for the maximum intensity of tropical cyclones, no significant change has been noted.

It has been found that the regional change in tropical cyclone frequency is closely related to the distribution of the SST anomaly and the change in convective activity associated with it. Over the regions where the SST anomaly is relatively small compared with surrounding regions, even though it is positive, convective activity and tropical cyclone frequency tend to decrease. This indicates that the climate of convective activity, and tropical cyclone frequency distribution, depends on dynamical factors associated with the SST distribution, rather than thermodynamical factors associated with the absolute value of local SST.

As for the decrease in the global total number of tropical cyclones on doubling CO2, a weakening of tropical circulation associated with the stabilization of the atmosphere (the increase in dry static stability) seems to be responsible for it. It is found that the rate of increase in the tropical precipitation due to the global warming is much less than the rate of increase in the atmospheric moisture. With this little increase in precipitation (convective heating), a considerable increase in the dry static stability of the atmosphere leads to a weakening of the tropical circulation.

There are considerable differences among the global or regional climate changes in tropical cyclone frequency indicated by different GCMs. A large difference in the amount of tropical precipitation increase seems to be responsible
for the difference in the global climate change, while differences in the SST anomaly used in the respective model seem to be responsible for the difference in the regional climate change. In other words, for a reliable prediction of future climate change in tropical cyclone frequency, a reliable estimate of SST change and tropical precipitation change is necessary.

The model used for the present study is a high-resolution model as a present-day AGCM. However, it is clear that the resolution of the model is still too coarse for studying a possible climate change of tropical cyclones, particularly the change in intensity of mature stage tropical cyclones. In the near future, it would be possible to do experiments with much higher resolution global models. Then, we will be able to simulate tropical cyclones much more realistically, and increase reliability of the experiment. To simulate tropical cyclones realistically, however, not only increasing the resolution of models, but also introducing more sophisticated physics, particularly convection process, is necessary. For example, Yamasaki (1977) pointed out an important role of drag force and evaporation of rain drops in the early stage of development of a tropical cyclone. Such processes are not included in the parameterization of cumulus convection in the present-day GCMs, but may be required in the future to simulate development of tropical cyclones more realistically.

The present study indicates that, for a reliable prediction of future climate change in tropical cyclone climatology, not only a realistic simulation of tropical cyclones with high-resolution models, but also a reliable estimate of SST change with a low resolution coupled GCMs is needed. A reliable prediction of the change in large scale tropical circulation and precipitation is also important.

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