A Climatology of Sea Surface Temperature and the Maximum Intensity of Western North Pacific Tropical Cyclones

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Abstract

Using a 31-year (1960–1990) sample of western North Pacific tropical cyclones and monthly mean sea surface temperature (SST) for each year, an empirical relationship between SST and the maximum intensity of western North Pacific storms is determined and used to calculate the relative intensity, a measure of how close a storm reaches its maximum potential intensity. The analysis method in this study follows that by DeMaria and Kaplan (1994b) and results are compared with observations over the North Atlantic and theoretical studies.

Similar to previous studies, an upper bound of storm intensity for a given SST was determined. It is shown that a larger fraction of Pacific storms are observed over warm waters than Atlantic storms and the maximum potential intensity of Pacific storms tends to be stronger than that of Atlantic storms or theoretically calculated storms. The analyses of the relative intensity at the time of each storm's life-time maximum intensity indicate that the maximum intensity of Pacific storms is well below the maximum potential intensity. The average relative intensity of the total sample is 37 % (47 %) when the regression curve for the maximum (99th) intensity percentile is used to compute the relative intensity, implying that environmental influences appear to be more important than SST in determining the maximum intensity of Pacific storms. The average relative intensity of late-season storms tends to be, as in the Atlantic, larger than that of early-season storms, and the yearly-averaged relative intensity shows to some extent interannual variability but with little correlation either with quasi-biennial oscillation or with El Niño.

1. Introduction

Since the energy source to drive tropical cyclone comes mainly from the underlying ocean in a form of surface heat fluxes, it is natural to assume that tropical cyclone intensity is related to the sea surface temperature (SST). Based upon the thermodynamic calculation and observation, Miller (1958) determined the minimum surface pressure that a tropical cyclone can attain for a given SST. The hydrostatic pressure reduction could be obtained by lifting an air parcel moist adiabatically from the surface in the eyewall and then subsiding it dry adiabatically in the eye while allowing for some mixing with the eyewall air to match observed relative humidities in the eye. Merrill (1987) investigated the relationship between the climatological SST and the intensity of North Atlantic storms for a 12-year period of 1974–1985. He determined a capping function, which describes an empirical maximum intensity of a storm for a given SST. By treating secondary circulation as a Carnot cycle, Emanuel (1988) developed an exact equation that governs the maximum possible pressure fall in tropical cyclones. His theory states that the maximum intensity of tropical cyclones is a function of the SST, mean outflow temperature, and relative humidity in the boundary layer.

The results of Miller (1958), Merrill (1987), and Emanuel (1988) indicate that the stronger storms occur over regions with higher SST and SST imposes an upper bound of tropical cyclone intensity. However, environmental factors other than SST also influence tropical cyclone intensity (Merrill, 1987). Evans (1993) related the monthly mean SST to the maximum intensity of storms over five ocean basins for a 20-year period of 1967–1986. She showed that SST is not the dominant factor in determining the maximum storm intensity and there are many other possible influences. This is consistent with Merrill's (1987) result. She also showed that compared with

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storms over the North Atlantic there is no obvious relationship between SST and the maximum intensity of storms in other ocean basins, including the western North Pacific, suggesting some different characteristics for different ocean basins.

DeMaria and Kaplan (1994b), hereafter DK, elaborated previous observational studies (Merrill, 1987; Evans, 1993) using the climatological SST and a 31-year sample (1962–1992) of North Atlantic tropical cyclones. Their observations are in agreement with the theoretical results by Emanuel (1988) over a wide range of SST, provided that the tropopause temperature is assumed to be a function of SST. They showed that only about 20 % of Atlantic storms reach 80 % or more of their maximum potential intensity and on average storms reach about 55 % of their maximum potential intensity when the total sample is considered.

A similar elaboration is yet to be done to characterize the relationship between SST and the maximum intensity of western North Pacific storms in comparison with North Atlantic storms. In this paper, following the analysis method described by DK, a relationship between SST and the maximum intensity of western North Pacific tropical cyclones is determined and used to compute the relative intensity. Results are compared with observations over the North Atlantic and theoretical studies. For this purpose, a 31-year sample of western North Pacific tropical cyclones and monthly mean SST for each year are used for analyses.

2. Data and analysis method

The data used in this study consist of the 31-year sample of tropical cyclones (tropical storm or typhoon strength) from 1960 to 1990 over the western North Pacific. The positions and intensities of tropical cyclones in every 6-h interval are archived from the RSMC (Regional/Specialized Meteorological Centers) Tokyo — Typhoon Center. The storm intensity in this data set is expressed in the minimum surface pressure and the maximum wind. In this study, the storm intensity is represented by the minimum surface pressure. The observations where the storms are over lands or islands are excluded from the present analysis because this study relates SST at the storm center to the storm intensity. Whether a storm is located over ocean or land or island is determined using $10'$ latitude by $10'$ longitude data that depict coastlines. The observations where SST at the storm center is less than 14.5°C are excluded for a comparison with DK's results. However, the number of observations of SST < 14.5°C is very small compared with the total observation number. The final data contain 15,819 observations of 857 tropical cyclones.

The monthly mean SST data for each year from 1960 to 1990 are archived from the U.S.A. NCEP (National Centers for Environmental Prediction). These SST data products are derived from ship, satellite, and sea ice limits data and in this study the reconstructed historical monthly analyses on a 2° latitude by 2° longitude grid are used. To obtain SST at a storm center, the gridded SST values are linearly interpolated in both space and time to the storm position and date, assuming that the SST analysis fields represent the middle of each month.

3. Results and discussion

Figure 1 shows the scatter plot of SST versus storm intensity for all time periods (15,819 observations). Note that the RSMC Tokyo — Typhoon Center assigns intensities for a majority of storms below 990 hPa in 5 hPa interval (last digit of the minimum surface pressure is 0 or 5). It is clear from this figure that weak storms are observed over regions with a wide range of SSTs, while strong storms are observed only over regions with a narrow range of high SSTs. For example, storms with about 980 hPa strength exist over regions with SSTs between 14.5° and 30.5°C, while storms with about 900 hPa strength exist only over regions with SSTs above 27°C. Similar to previous observational studies (Merrill, 1987; Evans, 1993), this figure shows that there is an upper bound of tropical cyclone intensity for a given SST. Although the number of storms in the SST range of 14.5°–20.5°C is very small compared with the total number of observations (see Table 1), this low SST range is included in this study mainly for comparisons with North Atlantic storms (DK).

Following DK, the observations are divided into 16 SST groups in a 1°C interval from 14.5° to 30.5°C. Table 1 lists the SST midpoint, number of observations, average SST, and average intensity for each SST group. Also, listed is the number of storms in the SST group which contains SST at the location where a storm reaches its life-time maximum intensity. Since there are cases in which the same maximum intensity is recorded more than one time for a life-time period of a storm, in this study the life-time maximum intensity of a storm is identified as the intensity at the time when the storm first reaches its maximum intensity for simplicity. The 82 % of the observations belong to the 27°–29°C SST groups with a maximum frequency (43 %) in the 29°C group. North Atlantic storms have a maximum frequency (29 %) in the 28°C group (DK). A comparison of Table 1 with the corresponding table in DK indicates that a larger fraction of western North Pacific storms are observed over warm waters, say, SST > 27.5°C, than North Atlantic storms. The 87 % of the observed storms have their maximum life-time intensities in the 27°–29°C SST groups with a maximum frequency (44 %) in the 29°C group. Only about 8 % of storms at-

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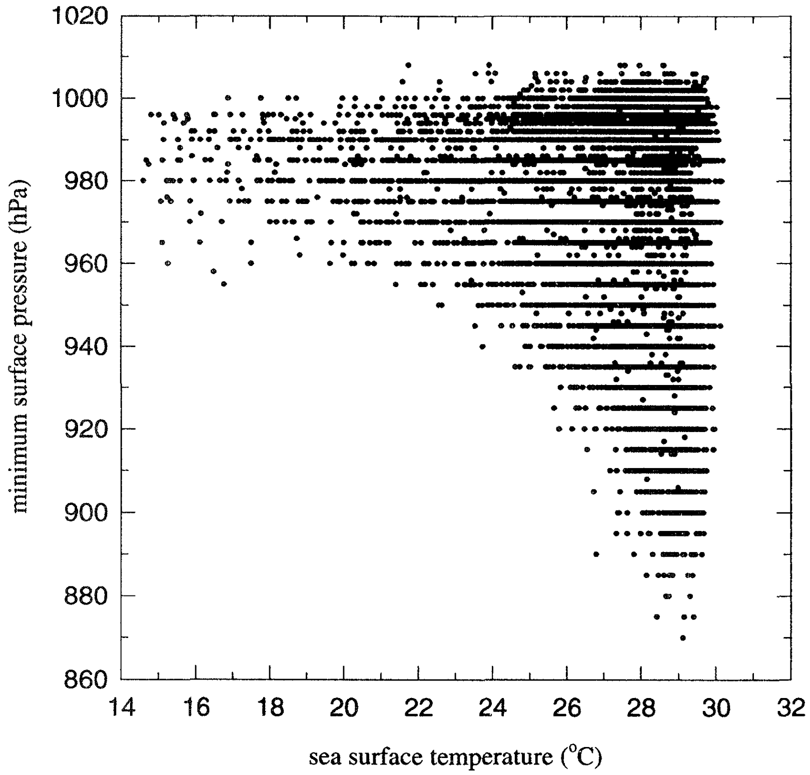

Fig. 1. The scatter diagram of sea surface temperature versus tropical cyclone intensity for all time periods.

Table 1. Characteristics of each SST group in an interval of 1°C.

SST midpoint (°C)	number of observations	number of storms in life-time maximum intensity	average SST (°C)	average intensity (hPa)
15.0	19	0	15.1	983
16.0	26	1	16.0	984
17.0	29	2	17.0	985
18.0	43	1	18.0	985
19.0	54	0	18.9	984
20.0	74	0	20.1	982
21.0	104	1	21.0	982
22.0	148	2	22.0	983
23.0	202	4	23.0	981
24.0	264	3	24.0	979
25.0	512	19	25.0	979
26.0	774	33	26.0	976
27.0	1722	106	27.1	975
28.0	4375	265	28.1	974
29.0	6858	377	29.0	973
30.0	615	43	29.7	973

tain their maximum life-time intensities over regions with SST $<$ 26.5°C. The average SST in each SST group is within 0.1°C of the midpoint value except for the 30°C group where the average SST is 29.7°C. This is because there are few cases with SST $>$ 30°C (14 cases) and the highest SST value at storm center is 30.2°C in this sample. For this reason, in the subsequent analysis the value of 29.7°C is used for the SST midpoint in the 30°C group.

Next, the scatter diagram of SST versus life-time maximum storm intensity (857 dots) was plotted (figure is not shown here). Results showed that over cool SST regions, say, below 24.5°C, there are few cases in which storms attain their life-time maxi-

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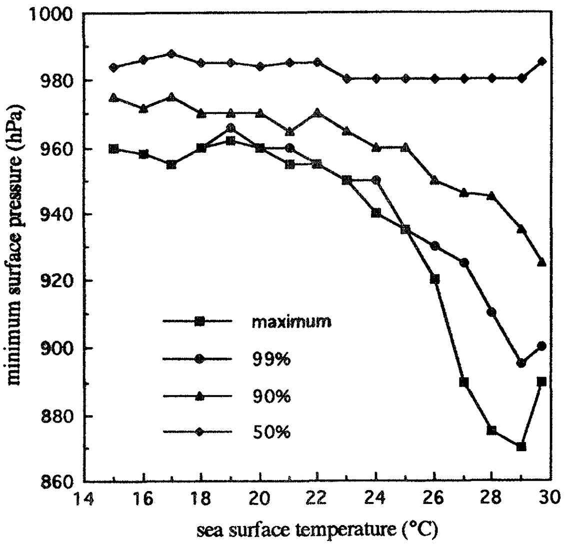

Fig. 2. The sea surface temperature versus the maximum (100th), 99th, 90th, and 50th storm intensity percentiles. The marks represent the midpoint of each SST group.

mum intensities (see the third column of Table 1) and there is no obvious relationship between SST and life-time maximum storm intensity. However, there appears to be an upper bound of life-time maximum storm intensity for a given high SST.

Figure 2 shows the maximum (100th), 99th, 90th, and 50th storm intensity percentiles for each SST group. The 50th intensity percentile exhibits very little change with SST variations. However, the maximum, 99th, and 90th intensity percentiles tend to be strongly dependent upon SST when SST is above about 23°C. In the 24°–27°C SST groups, the maximum storm intensity is more sensitive to SST changes at warmer SSTs. This is consistent with the results of Miller (1958) and Emanuel (1988). However, the slope of the maximum storm intensity percentile in the 27°–29°C SST groups is reduced. DK argued that the reduced slope, in their case in terms of the maximum wind, is somewhat related to variations in the tropopause temperature. Another plausible explanation for the reduced slope, a limiting factor in intensity, is that evaporative cooling of blowing sea spray at extreme wind speeds may inhibit storm intensification (Lighthill *et al.*, 1994; Willoughby, 1995). This explanation is supported by Uang and Thorncroft (1996) who showed that the model storm intensity at the mature stage simulated with the parameterized sea spray effect is slightly less intense than that simulated without it when the sea spray parameterization is included from the very beginning of the time integration. The maximum storm intensity in the 29°C SST group is stronger than that in the 30°C SST group. This might be partly related to the small sample size in the 30°C SST group, which may not represent the group properly. Figure 2 shows the flattening in the maximum and 99th intensity percentiles in the 15°–23°C SST groups. This implies that in this range of SSTs, many storms attain their potential intensities by moving over cooler waters rather than by intensification. This situation is possible when a storm moves quickly over region of gradually decreasing SSTs so that the storm cannot have enough time to be adjusted in intensity to SST near the storm center.

To determine empirical functions that cap storm intensity over the western North Pacific, least squares fitting to exponential function and polynomial function with order of up to 5 was performed using the observed maximum and 99th intensity percentile data (Fig. 2). Among the tested functions, polynomial equation of order 5 fits best to both the maximum and 99th intensity percentiles. The regression curve for the maximum storm intensity percentile is given by

$$P_r = 0.004399994T^5 - 0.466817T^4 + 19.4898T^3 - 400.881T^2 + 4066.55T - 15329.0, \quad (1)$$

and the regression curve for the 99th storm intensity percentile is given by

$$P_r = -\ 0.0000762819T^5 + 0.0155213T^4 - 1.00220T^3 + 28.3003T^2 - 365.422T + 2727.63, \quad (2)$$

where P_r is the minimum surface pressure in hPa and T is the sea surface temperature in °C at a storm center.

To compare the present results with previous ones on the maximum possible intensity of tropical cyclones, the potential minimum surface pressures as a function of SST based upon the observational and theoretical studies are plotted in Fig. 3. This figure shows that for a given SST in the indicated SST range the minimum surface pressure revealed as the maximum intensity curve for western North Pacific storms is lower than the potential minimum surface pressures for North Atlantic storms analyzed by Merrill (1987, 1988) and DK. This is also true for the 99th intensity percentile curve except for the slightly lower minimum surface pressure in Merrill's curve in the 29°C SST group. For North Atlantic storms, DK's curve exhibits greater intensity than Merrill's when SST is less than about 28.3°C and weaker intensity at higher SST. At 27°C SST, the difference in the minimum surface pressure between the two is as large as 17 hPa. This suggests that data and analysis method can affect potential intensity calculation. Since the analysis period in DK is from 1962 to 1992, which includes that of 1974–1985 in Merrill (1987), DK's results might be more

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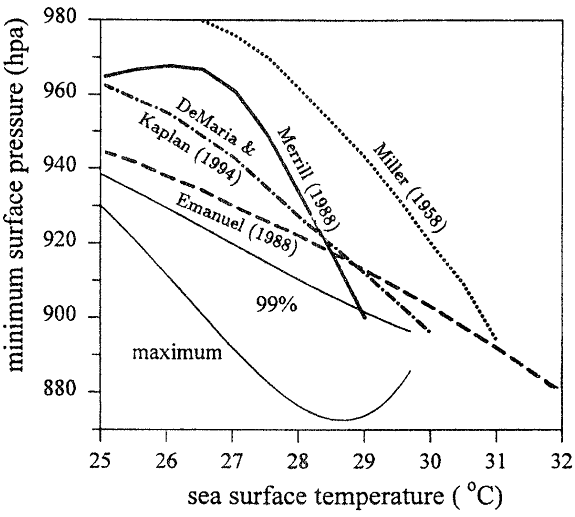

Fig. 3. The potential minimum surface pressures of tropical cyclones as a function of sea surface temperature. The results of Miller (1958), Emanuel (1988), Merrill (1988), and DeMaria and Kaplan (1994b) are reproduced from Willoughby (1995) and the regression curves for the maximum and 99th intensity percentiles in the present study are drawn. In the curve from Emanuel's result, mean outflow temperature of −75°C is used.

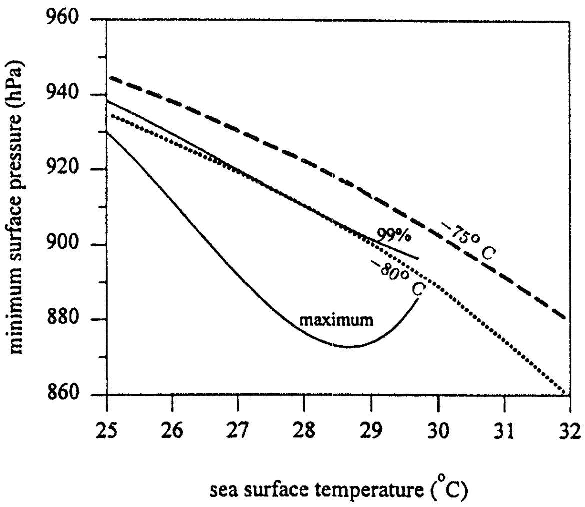

Fig. 4. The potential minimum surface pressures of tropical cyclones as a function of sea surface temperature for the Emanuel's result using mean outflow temperatures of −75°C (dashed curve) and −80°C (dotted curve) and the regression curves for the maximum and 99th intensity percentiles in the present study.

statistically stable, and hence more representative of Atlantic storms, because of the larger sample size. However, impact of data and analysis method on the calculation needs to be examined further. Based upon the observational results in Fig. 3, it is concluded that for a given SST western North Pacific storms tend to climatologically reach greater potential intensity than North Atlantic storms. Figure 3 also indicates that the potential minimum surface pressure estimated by Miller (1958) is highest over the overlapping SSTs in the six curves.

The 99th intensity percentile curve lies below Emanuel's curve evaluated at mean outflow temperature of −75°C and closely matches with Emanuel's curve at mean outflow temperature of −80°C (Fig. 4). Note that the mean temperature at 100 hPa averaged over 4°–8° area of typhoons is −78°C (Gray *et al.*, 1975). The climatological tropopause temperature, which might be thought to give a lower bound estimate of the outflow temperature (DK), increases toward the north away from the equator (higher tropopause height toward the equator). Accordingly, the warmer tropopause coincides with the lower SSTs and the potential storm intensity over cool waters should be weaker than the Emanuel's calculation with the fixed mean outflow temperature (Fig. 4). An examination of North Atlantic storms over the relatively cool waters in connection with the Emanuel's results (Figs. 3 and 4) does reveal this combined effect. However, Fig. 4 shows that over cool waters, the potential intensities of western North Pacific storms are still close to the theoretical estimates with the lower outflow temperature. It is speculated that this is to some extent because, in the western North Pacific, more storms with strong intensities developed or matured over warm waters tend to move over cool waters without greatly changing intensity.

To get a better idea of how close western North Pacific storms reach their maximum potential intensities, we calculated for each storm the relative intensity, which is defined in this study as

$$R = \frac{P_0 - P_c}{P_0 - P_r}. \tag{3}$$

Here, P_0 is the ambient surface pressure (specified as 1013 hPa), P_c the minimum surface pressure of a storm, and P_r the maximum potential intensity obtained from (1) or (2). For the P_c value, the minimum surface pressure at the time when the storm has its first maximum intensity during the life-time is used, and the regression curve is used to compute P_r at its corresponding location. The use of the 99th regression curve gave larger values of R than the maximum regression curve, especially for SSTs around 28°C, but overall analysis results were similar to each other in trend. Thus, results calculated using the maximum regression curve will be mainly presented below, but some results using the 99th regression curve will be briefly mentioned where nec-

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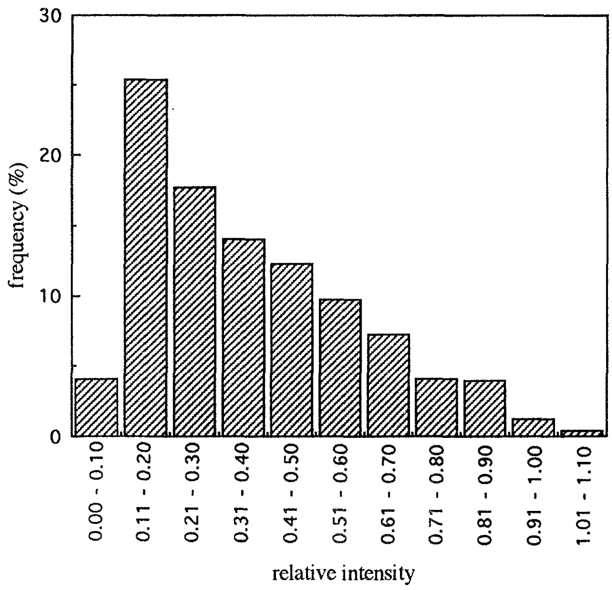

Fig. 5. The histogram of the frequency distribution of the relative intensity at the time of storm's first maximum intensity. The relative intensity is calculated using the regression curve for the maximum intensity percentile.

essary. The relative intensity defined in (3) is a ratio of the actual pressure drop to the maximum potential pressure drop. If R is close to one, it means in general that the storm has a life-time maximum intensity close to the maximum potential intensity predicted by SST.

To examine the frequency distribution of the relative intensity, relative intensities are grouped in a 0.1 interval (Fig. 5). The highest frequency (25 %) is observed in the interval between $R = 0.1$ and 0.2. The average relative intensity of the total sample is 37 % (47 %) when the regression curve for the maximum (99th) intensity percentile is used to compute the relative intensity. Over the North Atlantic, the average relative intensity of the total sample is 58 % when the land cases are removed (DK). As in North Atlantic storms, western North Pacific storms attain their maximum intensities well below maximum potential intensities. Figure 5 indicates that environmental influences appear to be more important than SST in determining maximum intensity of western North Pacific storms. It is well known that vertical wind shear in the storm environment and SST reduction through oceanic mixing and upwelling associated with storm circulation can reduce tropical cyclone intensity (*e.g.*, DeMaria *et al.*, 1993; DeMaria and Kaplan, 1994a).

It should be remembered that in DK the relative intensity is defined in terms of the wind speed, but in this study it is defined in terms of the pressure. Therefore, a direct comparison in magnitude of relative intensity, but not in trend, is not reasonable, although both the defined relative intensities represent a measure of how close a storm reaches its maximum potential intensity.

Table 2. The average relative intensity for each month of June–November. The relative intensity is calculated using the regression curve for the maximum intensity percentile.

month	average relative intensity (%)
June	28.5
July	36.4
August	35.4
September	39.4
October	42.2
November	42.8

The histogram of average relative intensity for each month of June–November is shown in Table 2. The average relative intensity of late-season storms tends to be larger than that of early season or mid-season storms (this is also true when the 99th regression curve is used). Further analysis for each month at the time and location of storm's first maximum intensity indicated that relatively lower SST and higher storm intensity in November result in the largest relative intensity (42.8 %).

Figure 6 shows the geographical distribution of average relative intensity together with the number of storms in a 10° latitude by 10° longitude box. Among 857 storms, 5 storms at their maximum lifetime intensities are located outside the domain. Between 10°–30°N band, the number of storms that reach their maximum life-time intensities tends to become larger when approaching farther west except for the western boundary of the domain. In the region between 10°–40°N and 120°–150°E, the average relative intensity decreases with increasing latitude. This might be to some extent due to the increasing baroclinic effect, such as increased vertical wind shear, *etc.* with latitude. The highest frequency in the number of storms (136 storms, 16 %) is observed in the region between 10°–20°N and 110°–120°E, but in this region the average relative intensity is only 24 %. A possible reason for the lower average relative intensity is that storms moving westward in this region often cannot have enough time for further intensification before landfalls because of the landmass influence.

The time series of yearly-averaged relative intensity is shown in Fig. 7. Both the regression curves for the maximum and 99th intensity percentiles are used to compute the relative intensity. For a comparison, the time series of yearly-averaged relative intensity

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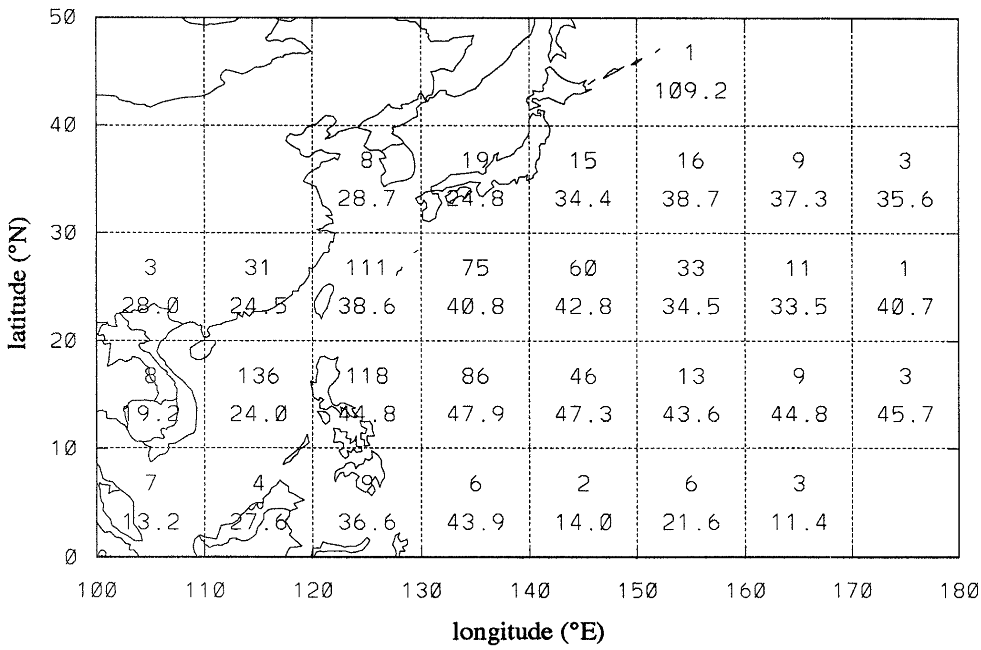

Fig. 6. The geographical distribution of the number of storms (upper value) and average relative intensity (lower value) in a 10° latitude by 10° longitude box. The relative intensity is calculated using the regression curve for the maximum intensity percentile.

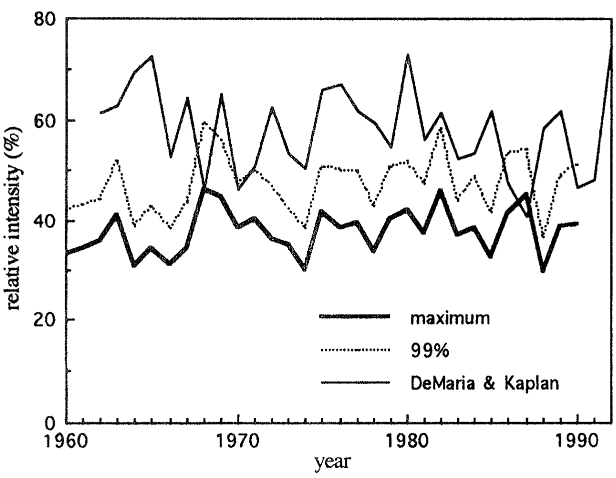

Fig. 7. The time series of yearly-averaged relative intensity. The relative intensity is calculated using the regression curves for the maximum and 99th intensity percentiles. Also, the time series of yearly-averaged relative intensity of North Atlantic storms (land cases removed) (DeMaria and Kaplan, 1994b) is shown for a comparison.

of North Atlantic storms for 1962–1992 (DK) is also shown. This figure shows to some extent interannual variability in the average relative intensity. The average annual relative intensity calculated using the maximum percentile regression curve ranges from the lowest value of 30.5 % in 1964 to the highest value of 46.5 % in 1968.

To examine whether there are some significant connections between the well-known large-scale circulation features, quasi-biennial oscillation (QBO) and El Niño, and the interannual variability of western North Pacific storms in the average relative intensity, the yearly-averaged relative intensity time series data were stratified according to the classifications by Gray (1984), Shapiro (1989) and DK. The stratification of the yearly-averaged relative intensity into west QBO years and east QBO years showed no significant difference in the average relative intensities between west QBO years and east QBO years. The stratification of the yearly-averaged relative intensity into El Niño years and non-El Niño years also showed no significant difference in the average relative intensities between El Niño years and non-El Niño years. In both cases, the difference is less than 1.5 %. DK showed that Atlantic storms tend to reach a larger fraction of their maximum potential intensities during west QBO years than during east QBO years, but exhibit about the same average relative intensity in the samples with and without El Niño.

The problem of whether global climate change

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due to an increase in atmospheric CO_2 concentration is likely to increase the maximum intensity of tropical cyclones surely deserves to be examined. The well-established climate model predicts modest SST increase in the tropical oceans by about 1°C with its variation limited within 0°–2°C (Carson, 1992). This SST increase can lead to a substantial increase in the maximum potential intensity of tropical cyclones, provided that other environmental conditions work very favorably for tropical cyclone formation and intensification. However, since SST appears to play, as already mentioned, a less important role in determining the maximum intensity of tropical cyclones than other environmental influences, there is no reason to expect that the maximum intensity become stronger with increasing SST in a climatological point of view. Based upon a careful objective assessment on global climate change and tropical cyclones, Lighthill *et al.* (1994) concluded that all of the other causes of variability in tropical cyclone frequency and intensity will swamp any changes associated with the modest increases in tropical ocean temperatures that are predicted to emerge from a doubling of atmospheric CO_2. These studies suggest that when studying tropical cyclone climatology in a future environment of global warming, other possible environmental conditions besides SST that influence tropical cyclone formation and intensification should be investigated carefully.

4. Conclusion

Following the analysis method by DeMaria and Kaplan (1994b), an empirical relationship between SST and the maximum intensity of western North Pacific tropical cyclones was determined and used to calculate the relative intensity. A detailed comparison of results with observations over the North Atlantic showed some differences as well as some similarities.

A considerable effort has been undertaken and progress has been made in forecasting tropical cyclone track accurately using numerical models, but much less progress has been made so far in forecasting tropical cyclone intensity reliably because of many problems associated with the model resolution, cloud representation, initial data, computing capability, *etc.* A statistical tropical cyclone intensity prediction model (*e.g.*, Nyoumura and Yamashita, 1984; Merrill 1987; DeMaria and Kaplan 1994a), hence, might be helpful at least until a reliable numerical model for forecasting tropical cyclone intensity becomes available. There are many factors suggested to be included in a statistical tropical cyclone intensity prediction model. One of them is a represention of how close a storm reaches the maximum potential intensity for a given SST (Merrill, 1987; DeMaria and Kaplan, 1994a). The empirical relationship between SST and the maximum intensity of tropical cyclones determined in this study may be used for certain information for a statistical tropical cyclone intensity prediction model over the western North Pacific. This study also indicated that environmental factors other than SST play more significant role in determining tropical cyclone intensity over the western North Pacific.

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北西太平洋上の海面水温と台風最大強度の気候値について

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北西太平洋における 31 年間（1960–1990）の熱帯低気圧と月平均海面水温（SST）とのデータを用い、SST と北西太平洋ストーム最大強度との経験的関係を求めた。さらにこの関係から、最大ポテンシャル強度を基準にした相対強度を計算した。解析手法は、DeMaria and Kaplan（1994b）にならい、結果を北大西洋の観測結果および理論値と比較した。

これまでの研究と同様に、ある SST に対するストーム強度の上限を求めた。大西洋上のストームに比べ、太平洋上のストームはより暖かい海面上に観測される割合が高く、最大ポテンシャル強度も大西洋上のストームや理論的に求められた値より大きかった。個々の台風の一生の最大強度について最大（99 %）ポテンシャル強度の回帰曲線を用いて相対強度を計算すると、平均で 37 %（47 %）であった。このことは、太平洋上のストームの最大強度を決める要素として、SST 以外の他の環境要因がより重要であることを示している。平均相対強度は年間のシーズン後半の方がシーズン前半よりも大きい傾向があり、この点は大西洋と同様であった。年平均相対強度は経年変動を示すが、準 2 年振動やエルニーニョとの相関は認められなかった。

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