Sea Surface Temperature and the Maximum Intensity of Atlantic Tropical Cyclones

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ABSTRACT

An empirical relationship between climatological sea surface temperature (SST) and the maximum intensity of tropical cyclones in the North Atlantic basin is developed from a 31-year sample (1962-1992). This relationship is compared with the theoretical results described by Emanuel. The theoretical results are in agreement with the observations over a wide range of SST, provided that the tropopause temperature is assumed to be a function of SST. Each storm is examined to determine how close the observed intensity comes to the maximum possible intensity (MPI). Results show that only about 20% of Atlantic tropical cyclones reach 80% or more of their MPI at the time when they are the most intense. On average, storms reach about 55% of their MPI. Storms that are farther west and farther north tend to reach a larger fraction of their MPI. Storms are also more likely to reach a larger fraction of their MPI in August-November than in June-July. There is considerable interannual variability in the yearly average of the ratio of the observed maximum intensity to the MPI.

1. Introduction

It has long been recognized that the sea surface temperature (SST) plays a significant role in the formation and intensification of tropical cyclones (Palmen 1948; Miller 1958). However, the SST by itself is not a very good indicator of whether an individual tropical cyclone will intensify. For example, during the 1992 Atlantic hurricane season, the maximum surface wind of Tropical Storm Earl reached 28 m s⁻¹ with an SST of about 27°C, while the maximum wind of Hurricane Bonnie reached 49 m s⁻¹ with an SST of about 25°C.

Although the SST is not a good predictor of intensity change, it does provide an upper bound on the storm intensity. Miller (1958) developed a relationship between the SST and the minimum sea level pressure that a storm can attain. This relationship is based upon the pressure reduction that results from lifting air moist adiabatically to the tropopause level in the eyewall and then subsiding the air dry adiabatically back to the surface in the eye. In this theory, some mixing with the eyewall air must be included during the subsidence to obtain realistic humidities and temperatures in the eye. Emanuel (1988) developed a theory for the minimum sea level pressure of tropical cyclones where a storm is treated as a Carnot heat engine. In this theory, the upper bound on intensity is a function of the SST, the relative humidity in the boundary layer, and the atmospheric temperature in the outflow layer of the storm. The results of Miller (1958) and Emanuel (1988) show that the maximum intensity increases rapidly for increasing SST. These two theories also indicate that the relationship between maximum intensity and SST is nonlinear, where the maximum intensity is more sensitive to SST variations at warmer SSTs.

Merrill (1987) examined the relationship between storm intensity and climatological SST for a 12-year sample of Atlantic tropical cyclones. His results show that the most intense storms occur over regions with the warmest SST, but the presence of warm water is not a sufficient condition for intensification. The results of Merrill (1987) again indicate that SST provides an upper bound on the intensity of tropical cyclones. Evans (1991) compared storm intensity in several ocean basins with the observed SST at the time when storms were at their maximum intensity. Her study confirmed the results of Merrill (1987) for Atlantic storms, although the relationship between SST and maximum intensity was not as obvious in other ocean basins.

The concept of an upper bound on intensity has been used in a number of applications. Darling (1991) used the theory of Emanuel (1988) to estimate probabilities of hurricane wind speeds. In his study, observed storm intensities were normalized by the upper-bound intensity to facilitate the comparison of cases in different SST regimes. Merrill (1987) and DeMaria and Kaplan (1994) used empirical relationships between the SST and storm intensity in statistical intensity forecast models. Their results show that the difference between the upper-bound intensity and the cur-

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rent storm intensity is a more reliable predictor of intensity change than the SST.

The primary purpose of this study is to further document the relationship between climatological SST and the intensity of Atlantic hurricanes. The work of Merrill (1987) is extended by using a 31-year sample (1962–1992) of Atlantic storms to develop an empirical relationship between SST and maximum storm intensity. This empirical relationship is then compared with the theoretical results of Emanuel (1988). It has often been noted that most storms do not reach their upper bound on intensity (e.g., Miller 1958; Emanuel 1991). The empirical relationship developed from the 31-year sample is used to quantitatively determine the probability that a given storm will reach a specified fraction of its maximum possible intensity. The tropical cyclone and SST data are described in section 2. The empirical relationship between SST and intensity and the comparison with theoretical results are presented in section 3. In section 4, the storms are examined at the time when their maximum intensity.

2. Tropical cyclone and SST data

An archive of the tracks and intensities of Atlantic tropical cyclones from 1876 to the present (Jarvinen et al. 1984) is available from the National Hurricane Center (NHC). As described by Merrill (1987), the most reliable portion of this data record is the period beginning in 1974 when the intensity estimates were determined from a systematic postanalysis of aircraft reconnaissance data and geostationary satellite imagery. By comparing data from the period prior to 1974 with those from 1974 to 1985, Merrill (1987) determined the reliability of the observations from the earlier period. This comparison showed that the data record beginning in 1962 was generally consistent with the period from 1974 to 1985. Prior to 1962, there was a fairly sharp increase in the number of inconsistent observations in the data record. For this reason, the period from 1962 to 1992 was chosen for the current study.

The NHC archive contains the positions and intensities of all Atlantic tropical cyclones at 6-h intervals. The intensity is represented by the maximum 1-min sustained surface wind or the minimum sea level pressure of the storm. The wind estimates are available for every 6-h interval but the sea level pressure estimates are often not reported. For this reason, the intensity is represented by the maximum sustained surface wind in this study. Merrill (1987) noted that in a few high-latitude cases, very high maximum wind speeds were reported for storms that were moving unusually fast. To correct this problem, the wind speeds are evaluated in storm-relative coordinates. (For each storm, the translational speed is subtracted from the maximum wind estimate.) For the total sample, the average translational speed is 6 m s\(^{-1}\).

Many high-latitude tropical cyclones eventually become extratropical systems. The NHC archive includes the extratropical portion of the storms, but these observations were eliminated from the current sample. The NHC archive also contains storms that are categorized as subtropical. These systems were also eliminated from the sample. The final dataset used in this study contains 710 observations of the positions and intensities of 257 tropical cyclones that reached at least tropical storm intensity (maximum winds > 17 m s\(^{-1}\)). About 63% of these tropical storms reached hurricane intensity (maximum winds > 33 m s\(^{-1}\)).

The SST fields for this study were obtained from the climatological analysis described by Levitus (1982). Monthly mean values of SST were available on a 1° lat–long grid. These values were linearly interpolated in space and time to the position and date of each tropical cyclone observation. For the time interpolation, it was assumed that analyses were valid in the middle of each month. Monthly mean values of SST for some individual years (1962–1988) were also available. The monthly mean SSTs for the individual years were produced by the Geophysical Fluid Dynamics Laboratory (GFDL) and were obtained on a 1° lat–long grid (Oort et al. 1987). The GFDL SSTs were linearly interpolated in space and time using the same method as for the climatological SSTs.

3. Tropical cyclone intensity distribution

To determine the relationship between intensity and SST, each observation was assigned to an SST group. The SST from 14.5°C to 30.5°C was divided into 16 evenly spaced groups (1°C intervals) as indicated in Table 1. The 15°C–20°C groups contain 3% of the cases, the 21°–25°C groups contain 15% of the cases, and the 26°–30°C groups contain 82% of the cases. Except for the 30°C group, the average SST of all the cases in each group is within 0.1°C of the SST midpoint. The

<table>
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<th>SST midpoint (°C)</th>
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average temperature of 29.7°C for the warmest SST group is probably a more representative value for the midpoint because there were no cases in the sample with an SST above 30.0°C. In all of the following results, a value of 29.7°C was used for the midpoint of the warmest SST group.

For each SST group, the maximum storm intensity and the 99th, 95th, 90th, and 50th intensity percentiles were determined. Figure 1 shows that the 50th percentile varies very little as a function of SST, but the other percentiles and group maxima are strong functions of SST. The variation of the maximum and 99th percentile is much more pronounced for SST > 26°C than for the colder temperatures. The results in Fig. 1 are consistent with the results of Merrill (1987), although the curves in Fig. 1 are slightly less noisy due to the larger sample size.

The rather abrupt change in slope of the maximum (and the 99th and 95th percentiles) at 26°C in Fig. 1 might be a result of storms that move north across the Gulf Stream. The SST at the southern edge of the Gulf Stream in September is about 26°C. The timescale for tropical cyclone intensity change over the ocean is on the order of 12–24 h (Willoughby 1990). If a storm were moving north at 10 m s⁻¹, the SST at the storm center could decrease from 26°C to about 20°C in 12 h. If the storm intensity were close to its upper bound for a 26°C SST, a 12-h period might not be long enough for the storm to decay to the upper bound at the colder temperatures. For this reason, it might be expected that the observed intensities of the most intense storms would be similar for SSTs within a few degrees of 26°C. The intensity distribution was recalculated using the SST averaged over the previous 12 h (not shown). In this case, the maximum curve was similar to that shown in Fig. 1 but was somewhat smoother. The decrease in the maximum intensity for SSTs between 25°C and 26°C in Fig. 1 no longer occurred when the SSTs were averaged over the previous 12 h.

To smooth the results shown in Fig. 1, a function was fit to the maximum intensity observations. The shape of the maximum intensity and 99th percentile curves suggested an exponential function of the form

$$ V = A + B e^{C(T - T_0)}, \quad (1) $$

where $V$ is the maximum wind (m s⁻¹), $T$ is the SST (°C), $T_0$ is a specified reference temperature, and $A$, $B$, and $C$ are constants. Letting $T_0 = 30.0$ and using a least-squares fit, the constants in (1) are given by $A = 28.2$, $B = 55.8$, and $C = 0.1813$. Figure 2 shows the data and the fitted function. The average difference between the function and the data points in Fig. 2 is 2.3 m s⁻¹. In the remainder of this paper, the storm intensity determined from (1) will be referred to as the maximum possible intensity (MPI).

The choice of an exponential function of the form of (1) does not allow for the flattening (reduced slope) of the maximum curve in Fig. 1 for SSTs above 28°C. As will be described later, it is likely that this flattening is a real feature that is related to variations in the tropopause temperature. However, it is also possible that the sample size of the warmest SST group (Table 1) was too small to adequately determine the maximum
storm intensity. In addition, the 99th percentile in Fig. 1 does not have a reduced slope for SSTs above 28°C. Because there is some uncertainty in the shape of the maximum curve near the warm end of the SST interval, (1) should not be used for SSTs > 30°C.

To estimate the sensitivity of the maximum intensity relationship to the use of climatological SSTs, monthly averaged SST analyses for individual years were obtained for a subset of the cases (1962–1988). These analyses were produced by GFDL as described previously. Figure 3 shows the maximum intensity distribution for the Levitus climatology and the GFDL SST analyses for the 1962–1988 sample (6045 cases). This figure shows that the basic structure of the maximum intensity curve is very similar in each case, although the curve with the GFDL SST analyses is somewhat smoother than the Levitus curve. The GFDL SSTs are stored as deviations (anomalies) from the 1950–1979 mean fields for each month. The SST anomalies were evaluated at all of the tropical cyclone locations for the 1962–1988 sample. These results showed that the magnitude of the anomaly was <1°C in 98% of the cases, and the mean anomaly was very close to zero. These results indicate that the use of climatological SSTs to estimate the MPI of Atlantic tropical cyclones is not a severe approximation in most cases. Therefore, the Levitus climatology was used in all of the following results so that a consistent SST analysis could be used for the entire sample (1962–1992).

The theoretical results of Emanuel (1988) indicate that the maximum intensity of tropical cyclones is a function of SST, the temperature in the outflow layer of the storm, and the ambient boundary layer relative humidity. However, Emanuel (1987) argued that, to a first approximation, the maximum intensity is a function only of SST. This approximation is valid because the ambient relative humidity is fairly uniform over the tropical and subtropical oceans. Also, the tropopause temperature provides a lower-bound estimate of the outflow temperature, and the tropopause temperature over the tropical and subtropical oceans is strongly controlled by the SST (e.g., Reed and Gage 1981). Thus, if we assume a constant ambient relative humidity and specify a relationship between SST and tropopause temperature, we could use Emanuel's theory to determine the maximum intensity as a function of SST. These theoretical results could then be compared with (1). However, instead of specifying some relationship between SST and tropopause temperature, we determined the tropopause temperature as a function of SST (outflow temperature) from Emanuel's theory (Fig. 3a from Emanuel 1988), assuming that the ambient relative humidity was 80% and that (1) was exactly satisfied. For this purpose, we used the minimum sea level pressure to maximum wind conversion for Atlantic storms that was developed as part of the Dvorak satellite intensity classification scheme (Dvorak 1984) because the maximum intensity is represented by maximum wind in (1) and by minimum sea level pressure in Emanuel (1988). Although this pressure–wind relationship is not perfect, many of the intensity estimates included in the NHC archive were made with the Dvorak scheme.

Figure 4 shows the outflow temperature obtained from Emanuel (1988) assuming that (1) is exactly satisfied. This figure shows that the required outflow temperature decreases from about −55°C to −75°C as the SST increases from about 24°C to 30°C. The climatological tropopause temperature in the North Atlantic
region during the summer and fall varies from about −75°C in the deep Tropics to about −55°C in mid-latitudes (e.g., Newell et al. 1972). Thus, the outflow temperature structure in Fig. 4 is quite reasonable. This agreement indicates that the theoretical results of Emanuel (1988) are consistent with the observations of the maximum intensity of Atlantic tropical cyclones.

Figure 4 provides an explanation of why the maximum intensity in Fig. 1 is so much more sensitive to SST variations between 26°C and 29°C than for colder temperatures. In Emanuel’s theory, the minimum sea level pressure is a function of SST and tropopause temperature. When the SST increases from 26°C to about 29°C, the tropopause temperature decreases from typical midlatitude values to tropical values. Both of these factors lead to decreases in the minimum sea level pressure (increases in the storm intensity). For SSTs below 26°C there is less variation in the tropopause temperature with SST, so the storm intensity is less sensitive to SST variations in this temperature range. For SSTs above 29°C, the tropopause temperature is probably also less sensitive to SST variations, so it might be expected that the MPI would be less sensitive to SST variations for very warm SSTs. As described previously, the maximum curve in Fig. 1 does appear to flatten out at very warm SSTs.

The outflow temperatures in Fig. 4 decrease slightly as the SST decreases from 22°C to 15°C. In the North Atlantic during the hurricane season, SSTs in this range occur only on the north side of the Gulf Stream. The decrease in outflow temperature with decreasing SST might be due to storms that are moving rapidly toward colder water. As described previously, these storms may not have had enough time for the intensity to decrease to a level consistent with the instantaneous value of SST. In the theoretical relationship, this “extra” intensity requires a colder outflow temperature. An alternate explanation is that tropical cyclones might exist only over cold water when the tropopause temperature is anomalously cold. Although the climatological tropopause temperature is fairly uniform at higher latitudes, significant variations are associated with midlatitude weather systems (e.g., Palmen and Newton 1969). Another possible explanation for the decreased outflow temperatures is that the ambient relative humidity is lower for these high-latitude cases. In Emanuel’s theory, the maximum intensity increases as the ambient humidity decreases. If the ambient humidity were assumed to decrease with latitude, a warmer tropopause temperature would have been necessary to give the same intensity. Actual values of ambient relative humidity and tropopause temperatures would be necessary to refine the comparison between the theory and the observations.

4. The relative intensity of tropical cyclones

A comparison of the 50th percentile and maximum curves in Fig. 1 shows that most storms are well below their MPI. This result can partially be explained by the fact that the entire life cycle of each storm is included in the intensity distribution. Most storms form from easterly waves or baroclinic systems. However, some storms continue to develop until their intensity is close to the MPI, while others stop intensifying well before reaching the maximum. To give a better idea of how close storms get to their MPI, each storm was examined to determine the time when it reached its maximum intensity. The SST and MPI were calculated at this time, and then the relative intensity was determined. The relative intensity is defined as the maximum intensity of the storm divided by its MPI as determined from (1).

Figure 5 shows the location of each storm at the time of maximum intensity. There is a tendency for more storms to reach their maximum intensity on the western side of the Atlantic. This is not surprising because easterly waves that move off Africa often require several days to intensify to tropical storm strength. Figure 5 also shows that very few storms reach their maximum intensity in the central Caribbean, even though some of the most intense Atlantic tropical cyclones on record have passed through this region (for example, Hurricane Gilbert in 1988 and Hurricane Allen in 1980). This lack of intense storms might be related to the fact that the climatological location of the tropical upper-tropospheric trough (TUTT) extends into the central Caribbean during the hurricane season (e.g., Newell et al. 1972). The increased vertical wind shear associated with the TUTT might prevent many storms from intensifying in this region. It is also noteworthy that a large number of storms reach their maximum intensity very close to the coast in the Gulf of Mexico. This clustering of storms might simply be due to the fact that most storms move to the west or north in the Gulf and stop intensifying because of the influence of land. Another possibility is that due to land–sea contrasts or topographical influences, the region within about 100 km of the Gulf coast is favorable for tropical cyclone intensification.

Figure 5 shows the location of each storm when it reached its maximum intensity. The location of each
storm at the time when it reached its maximum relative intensity was also determined (not shown), and the
results were quite similar to those in Fig. 5. In a few
cases, storms reached their maximum relative intensity
at a slightly higher latitude than where they reached
their maximum intensity. This increase in latitude is
possible for storms that move rapidly north toward
colder water where the MPI decreases more rapidly
than the storm intensity so that the relative intensity
increases. If a storm moved rapidly enough over cold
water, it is possible that the actual intensity could ex-
ceed the MPI for a short period of time, so that the
relative intensity can be larger than unity.

Figure 6 shows the relative intensity distribution for
the 31-year sample. This figure shows that most storms
reach a maximum intensity that is well below the max-
imum possible intensity. (The average relative intensity
of the total sample is 55%.) One explanation for the
lack of intensification is the influence of land. Storms
with a relative intensity < 60% were examined in detail,
and it was found that in 41 cases (16% of the total
sample of storms) intensification ceased due to move-
ment over land. The relative intensity distribution with
the land cases removed is also shown in Fig. 6. The
average relative intensity of this sample (58%) is slightly
larger than that of the total sample.

In Fig. 6 there are maxima in the distribution for
relative intensities near 30% and 70% with a minimum
near 60%. This bimodal structure might be a result of
storms that stop intensifying for different reasons. For
storms that stop intensifying well before reaching their
MPI, unfavorable synoptic conditions such as the ver-
tical shear of the horizontal wind might be important.
For storms that stop intensifying after reaching larger
relative intensities, the response of the ocean to the
tropical cyclone circulation might be important. The
possible effects of the ocean response will be discussed
in more detail later in this section.

Figure 7 shows the cumulative distribution of relative
intensity. For the total sample with (without) the land
cases, 51% (38%) of the storms reached 50% of their
MPI, but only 16% (19%) reached 80% of their MPI.

To gain further insight into factors affecting the rel-
ative intensity, the sample was stratified by latitude,
longitude, month, and year. The average relative in-
tensity is defined as the mean of the relative intensities
(one value per storm) in each subset of the sample. As
described previously, the relative intensity distribution
(Fig. 6) is bimodal. Thus, the average relative intensity
does not describe all of the characteristics of the relative
intensity distributions of the subsets. In general, the
subsets also tended to have a bimodal structure. For
subsets with larger average relative intensities, the per-
centage of cases with higher relative intensities in-
creased, and the number of cases with the lower relative
intensities decreased, although the distributions for the
subsets were more noisy due to the smaller sample sizes.

A standard t statistic for the case with independent
samples with unequal variances (Steel and Torrie 1980)
was used to determine whether the average relative in-
tensities from different subsets were significantly dif-
ferent from each other. For this test, the 95% level was
used to determine statistical significance. The results
from the significance tests will be discussed where ap-
propriate.

Table 2 shows that the average relative intensity
generally increases with latitude. This result is some-
what surprising because the climatological vertical

![Figure 6](image1.png)

**Fig. 6.** The number of cases per 5% interval of relative intensity.

![Figure 7](image2.png)

**Fig. 7.** The cumulative distribution of relative intensity.
shear generally increases with latitude. When the vertical shear is large, storms are less likely to reach their MPI (DeMaria et al. 1993). The increase in relative intensity with latitude might be due to the fact that the northward component of the storm motion usually increases with latitude. Storms that are below their MPI and intensifying, but moving north, will eventually stop intensifying when they move over cold water, even if the synoptic environment is favorable for intensification. These types of storms should reach a large fraction of their MPI.

Table 3 shows that for the sample with the land cases removed, the relative intensity tends to be larger for storms that are farther west. This increase with west longitude might be due to synoptic influences. There are numerous cases of storms that reach tropical storm intensity after moving off Africa as tropical waves but then dissipate in the eastern or central Atlantic (Neumann et al. 1987). The decay of these storms is often associated with the influence of upper-level westerlies (increased vertical shear) that penetrates to low latitudes. As described previously, the climatological location of the TUTT is near 60°–70°W during the Atlantic hurricane season. In some cases, storms move west of this climatologically unfavorable environment and then are able to attain a larger fraction of their MPI. It is also possible that the tropopause is at a lower altitude in the vicinity of the TUTT, so that the tropopause temperature is warmer. Thus, the actual MPI near the TUTT might be lower than the MPI determined from (1), which allows for only a very smooth variation of tropopause temperature with SST.

Table 4 shows the average relative intensity for each month of the hurricane season. The peak of the Atlantic hurricane season as measured by the number of storms occurs in early September (Neumann et al. 1987). This peak is consistent with the sample sizes in Table 4. Table 4 shows that there is little difference in the average relative intensity from August to November. However, the relative intensity is lower in June–July. Thus, early-season storms are less likely to approach their MPI than mid- or late-season storms. It is somewhat surprising that early-season storms are less likely to approach their MPI than late-season storms. Climatologically, the upper-level westerlies over the tropical and subtropical Atlantic are considerably stronger in October than in July. One possible explanation for the lower relative intensities is that the maximum temperatures over land tend to occur in July, while the maximum SSTs in the tropical and subtropical Atlantic tend to occur in September. Thus, there might be increased large-scale subsidence over the Atlantic in July. Climatologically, the Bermuda–Azores high has its maximum pressure in July, which is also an indication of increased subsidence early in the Atlantic hurricane season.

The statistical tests were applied to the subsets of the total sample in Tables 2–4 and all of the relationships described above were significant. Specifically, the average relative intensity of the lowest latitude subset in Table 2 was significantly different from the average of each of the other subsets, most of the subset averages for the sample with the land cases removed in Table 3 were significantly different, and the average for July in Table 4 was significantly different from the averages for the later months.

Figure 8a shows the average relative intensity as a function of year. This figure shows that there is considerable interannual variability in the average relative intensity. For the total sample, the average annual relative intensity ranged from a low of 40.4% in 1987 to a high of 69.8% in 1992. With the land cases removed, it ranged from 40.9% in 1987 to 74.4% in 1992. Figure 8b shows the spectra of the average relative intensity time series. This figure shows that the interannual variability occurs on a variety of timescales. There appear to be three primary spectral peaks in Fig. 8b near periods of about 2 years (frequency = 14–15), 3–4 years (frequency = 8–12), and 15 years (frequency = 2).
Fig. 8. (a) The average relative intensity for each year of the sample and (b) the spectral amplitudes of the discrete Fourier transform of the average relative intensity time series.

There are well-established relationships between the interannual variability of Atlantic tropical cyclone activity and large-scale circulation features. These large-scale features affect the number of storms per year, the intensity of the storms, and, to some extent, the locations of the storms as described below. The number of tropical cyclones and the frequency of intense storms tend to increase when the stratospheric quasi-biennial oscillation (QBO) is in a more westerly phase (e.g., Gray 1984; Shapiro 1989). Atlantic hurricane activity (especially low-latitude storms) is usually suppressed during El Niño years (e.g., Gray 1984). Atlantic hurricane activity is also correlated with the rainfall in the Sahel region of West Africa (e.g., Landsea and Gray 1992). During years with above-normal precipitation (wet years), the number of intense tropical cyclones tends to increase.

To give an idea of how the above circulation features affect the relative intensity, the phase of the QBO (east, west, or transition), the presence or absence of El Niño, and the precipitation anomaly (dry or wet) of the Sahel region of West Africa were determined for each year of the 31-year sample. In 14 years, the QBO was in a westerly phase, in 13 years it was in an easterly phase, and in 4 years it was in a transition between phases, according to the classification described by Shapiro (1989) updated through 1992. According to the criteria described by Gray (1984) updated through 1992, there are 7 El Niño years in our sample (1965, 1972, 1976, 1982, 1983, 1986, and 1991). Using the data in Landsea and Gray (1992) updated through 1992, there are 9 wet years in our sample (1962, 1964-1967, 1969, 1975, 1988, and 1989).

Table 5 shows the characteristics of the stratified sample. Consistent with previous studies, there are more storms per season and the average annual maximum storm intensity is greater in wet QBO years, in years without El Niño, and in wet years. The El Niño primarily affects the number of storms per year, while the precipitation anomaly is better correlated with the average storm intensity. The average relative intensity appears to be larger in west QBO and wet years (especially when the land cases are removed) but is about the same in years with and without El Niño. The statistical tests showed that there was no significant difference in the average relative intensity in the samples with and without El Niño, and that the difference between the average relative intensities for the wet and dry years (with the land cases removed) was significant. The difference between the east and west QBO samples was not significant at the 95% level, but was significant at the 80% level. Thus, the average relative intensity tends to be larger in wet years, with some suggestion that it is higher in west QBO years.

Table 5 shows that the storms in El Niño years reach their maximum intensity at higher latitudes and thus have a smaller average MPI. Thus, the difference in the average storm intensity during El Niño years might primarily be due to the differences in storm tracks. In contrast, the average MPI is about the same in east and west QBO years and in dry and wet years, but the average relative intensity is larger in west QBO and wet years. Thus, storms tend to reach a larger fraction of their MPI during west QBO and wet years.

The above results suggest that normalizing by the MPI may be useful in studies of the interannual variability of tropical cyclones. This normalization may help distinguish between intensity changes due to track
variations (and thus SST variations) and intensity changes due to other factors.

Figure 7 shows that only about half of Atlantic tropical cyclones reach ~60% or more of their MPI, even when the land cases are removed from the sample. A natural question to ask is, Why don’t most storms over the ocean continue to intensify until they reach a larger fraction of their MPI? One possibility is that unfavorable environmental conditions (such as the vertical shear of the horizontal wind) limit the growth of most storms (e.g., Merrill 1988; DeMaria et al. 1993). Another possibility is that in some cases, the ocean responds to the tropical cyclone in a way that reduces the SST in the vicinity of the storm (e.g., Emanuel 1991). Shay et al. (1992) have shown that the SST cooled by ~4°C in the wake of Hurricane Gilbert as it moved through the southern Gulf of Mexico, although the SST directly under the eyewall cooled by only 1°–2°C.

To provide some insight into whether the intensification is limited by the ocean temperatures, the amount of cooling necessary to increase the relative intensity of each storm (at the time of maximum intensity) to 100% was calculated. For this purpose, \( I \) was used to determine the SST, given the observed maximum intensity of each storm. This SST was then subtracted from the actual SST to determine the required cooling. Figure 9 shows the distribution of the required cooling, where the cooling was divided into groups separated by 2°C intervals. As the SST decreases, the maximum possible intensity \( [V \text{ in Eq. (1)}] \) asymptotically approaches the constant \( A \) (28.2 m s\(^{-1}\)).

Because \( V > A \) for finite SST, no amount of cooling will give a relative intensity of 100% for storms with observed maximum winds \( < A \). All of these storms were included in the category with required cooling > 14°C.

Figure 9 shows that for the total sample, the required cooling is ~4°C in about 25% of the cases, for the total sample and in about 30% of the cases for the sample with the land cases removed. However, in the majority of the cases, the required cooling is >4°C, suggesting that intensification is not limited by the SST. This result suggests that the storm environment is probably important in limiting the storm intensification in most cases.

The results in Fig. 9 cannot be applied to estimate the impact of the ocean response on intensification because the equation for MPI implicitly includes a relationship between SST and outflow temperature. The SST reduction due to the storm occurs over a relatively small area and probably does not influence the outflow temperature in the same way as the large-scale SST.

5. Concluding remarks

An empirical relationship between climatological SST and the maximum intensity of Atlantic tropical cyclones was developed from a 31-year sample (1962–1992). This relationship was compared with the theoretical results described by Emanuel (1988). The theoretical results are in agreement with the observations over a wide range of SST, providing the tropopause temperature is assumed to be a function of SST. The implied relationship between tropopause temperature

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**Table 5.** Stratification of the total sample by factors affecting interannual variability. The average relative intensity for the sample with the land cases removed is in parentheses.

<table>
<thead>
<tr>
<th>Factor</th>
<th>No. of years</th>
<th>Average storms per year</th>
<th>Average lat</th>
<th>Average long</th>
<th>Average maximum intensity</th>
<th>Average MPI</th>
<th>Average relative intensity</th>
</tr>
</thead>
<tbody>
<tr>
<td>El Niño</td>
<td>7</td>
<td>5.4</td>
<td>30.8</td>
<td>65.3</td>
<td>31.4</td>
<td>58.7</td>
<td>55.5 (58.6)</td>
</tr>
<tr>
<td>No El Niño</td>
<td>24</td>
<td>9.1</td>
<td>26.8</td>
<td>66.6</td>
<td>32.9</td>
<td>61.0</td>
<td>54.4 (57.8)</td>
</tr>
<tr>
<td>East QBO</td>
<td>14</td>
<td>7.1</td>
<td>27.8</td>
<td>69.2</td>
<td>31.5</td>
<td>61.5</td>
<td>52.3 (54.8)</td>
</tr>
<tr>
<td>West QBO</td>
<td>13</td>
<td>9.1</td>
<td>26.8</td>
<td>66.3</td>
<td>33.3</td>
<td>60.6</td>
<td>55.3 (59.5)</td>
</tr>
<tr>
<td>Dry</td>
<td>22</td>
<td>8.0</td>
<td>27.4</td>
<td>66.2</td>
<td>31.5</td>
<td>60.7</td>
<td>52.8 (55.7)</td>
</tr>
<tr>
<td>Wet</td>
<td>9</td>
<td>9.0</td>
<td>27.4</td>
<td>66.9</td>
<td>35.3</td>
<td>60.6</td>
<td>58.2 (62.3)</td>
</tr>
</tbody>
</table>
and SST is consistent with the climatological temperature structure in the North Atlantic region.

The relative intensity was defined as the observed maximum intensity of a given storm divided by the (empirical) maximum possible intensity (MPI) determined from the SST. The distribution of relative intensity showed that only about 16% of Atlantic tropical cyclones reach 80% of their MPI. In some cases the lack of intensification was caused by the influence of land. However, even when the land cases were removed, only about 19% of the storms reached 80% of their MPI.

The 31-year sample was stratified by latitude, longitude, month of the year, and year. It was shown that the average relative intensity tends to increase with latitude and longitude and that storms tend to reach a larger fraction of their MPI in August–November than in June or July. It was also shown that there is considerable interannual variability in the yearly averaged relative intensity. Some of this variability is related to the phase of the QBO and the precipitation anomalies over the Sahel region of Africa, as might be expected from previous studies of the interannual variability of Atlantic tropical cyclone activity. These results also suggested that normalizing the storm intensity by the MPI may be useful for studies of the interannual variability of tropical cyclone activity.

To provide some insight into whether the storm intensification is limited by the SST, the amount of cooling necessary to reduce the MPI to the observed storm intensity (resulting in a relative intensity of 100%) was calculated for each of the 257 storms in the sample at the time when each storm was at its maximum intensity. In about 70% of the cases, a cooling of greater than 4°C would be required to increase the relative intensity to 100%. These results suggest that the storm intensification is not limited by the large-scale SST. Other factors such as the vertical shear are probably more important in most cases. The response of the ocean to the storm (which acts to locally reduce the SST near the storm center) might also be important in some cases.

In this study, climatological SST analyses were used for simplicity. It was shown that for a subset of the sample, the use of monthly averaged SST analyses for individual years had only a minor effect on the relationship between maximum intensity and SST. The results in this study could be refined by using SST analyses averaged over shorter time periods (perhaps weekly analyses). In addition, the SST is only one measure of the state of the upper ocean. Numerous studies have shown that other properties of the upper ocean, such as mixed layer depth and Richardson number, have an effect on the interaction between the ocean and tropical cyclones (e.g., Perlofth 1969; Shay et al. 1992). The results of this study could be generalized by determining the relationship between additional ocean parameters and the maximum intensity of tropical cyclones.

The empirical formula for MPI developed in this study implicitly includes a relationship between SST and outflow temperature. It is possible that this implicit relationship has seasonal and interannual variability. The results of this study could also be generalized by stratifying the sample by season or year (for example, into QBO west versus QBO east years) and then determining the effect on the intensity distribution. However, this stratification would probably require the inclusion of cases prior to 1962. More careful quality control would be necessary to include these additional cases.

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