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Low frequency variability of tropical cyclone potential intensity 1. Interannual to interdecadal variability

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[1] Recent research suggests that anthropogenic global warming would be associated with an increase in the intensity of tropical cyclones. A recent statistical analysis of observed tropical cyclone intensity shows that its variability with location and season is strongly tied to the variability of the thermodynamic potential intensity (PI) of tropical cyclones, as calculated using a theory described in an earlier work by the authors. Thus it is of interest to look for possible trends in global measures of PI, which are far more stable than those of actual storm intensity. We estimate global trends of PI from 1958 to 1996, averaged over the region where it exceeds 40 m s^{-1}, using the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis and the NCEP Empirical Orthogonal Function (EOF) sea surface temperature (SST) analysis. We adjust the Reanalysis temperatures for a large, spurious temperature increase that occurred around 1979. We do this by subtracting from the Reanalysis the atmospheric temperature difference between pairs of years with similar tropical SST before and after 1979. The value of the global mean PI is very large for the SST of the corresponding region in the mid-1990s. Supported by a recent study on the effects of ozone decrease on tropospheric temperatures, we suggest that the ozone decrease might be one of the factors contributing to increase of PI during the 1990s. *INDEX TERMS:* 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 3374 Meteorology and Atmospheric Dynamics: Tropical meteorology; 1630 Global Change: Impact phenomena; *KEYWORDS:* tropical cyclone, maximum intensity, reanalysis, global change

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1. Introduction

[2] Recent studies suggest that climate change associated with a doubled CO_2 would result in 10–20% increase of PI [*Henderson-Sellers et al.*, 1998]. Specifically, it has been estimated that the PI should rise about 3.5 m s^{-1} for every degree of tropical SST increase [*Emanuel*, 1987]. *Knutson et al.* [1998] obtained an increase of wind of 3–7 m s^{-1} for an increase of SST of about 2.2°C. It is tempting to look for a signal in global records of tropical cyclone strength, given that there has been measurable warming of the tropical oceans in the last few decades. Unfortunately, the predicted rate of increase is undetectable, given that the observed warming so far has been a few tenths of a degree C and that tropical cyclone intensities are only reported to the nearest 2.5 m s^{-1}. In addition, natural interannual and interdecadal variability of tropical cyclones is too large to allow the detection of any global warming signal that may be present.

[3] Recent studies indicate that thermodynamical estimates of PI of tropical cyclones [*Emanuel*, 1986; *Holland*, 1997] agree well with the observed maximum intensities of the most severe storms [*Tonkin et al.*, 2000]. The importance of PIs in describing tropical cyclone climatology is not restricted to maximum intensities. *Emanuel* [2000] calculated cumulative distribution functions (CDFs) of storm lifetime maximum wind speed normalized by climatological potential intensity. The CDFs of storms whose lifetime maximum exceeded 32 m s^{-1} and which were not limited by declining PI are nearly linear. These CDFs appear to be universal. This means that there is an equal likelihood that any given tropical cyclone will achieve any given intensity up to its PI. There is also a uniform probability that a storm that has not achieved its lifetime maximum intensity will have an intensity that is any given fraction of its lifetime maximum intensity. The universality of the normalized distribution functions suggests that *any climatic change in PI would affect the intensity distribution of real tropical cyclones uniformly*. In this paper (Part 1) we study the potential intensity during 1958–1996. In a companion paper (Part 2) [*Bister and Emanuel*, 2002] we present a

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climatology for years 1982–1995 for which the existing data are homogeneous in quality (see Part 2 for more information about the data).

[4] Knowledge of the SST and atmospheric profiles of temperature, humidity and pressure allow the calculation of PI, which can then be used as a proxy for the global intensities of hurricanes, assuming that the CDFs of *Emanuel* [2000] are universal. With the NCEP/NCAR Reanalysis data [*Kalnay et al.*, 1996; *Kistler et al.*, 2001], it has become possible to calculate the PIs for the last 50 years. However, the quality of the data poses problems and the temperature trends calculated using the Reanalysis data are considered unreliable [e.g., *Ebisuzaki and Kistler*, 2000]. The introduction of the National Environmental Satellite, Data and Information Service (NESDIS)-derived temperature retrievals in the NCEP/NCAR Reanalysis (hereafter referred to as NCEP Reanalysis or Reanalysis) data assimilation was associated with a spurious warming of the upper tropospheric temperatures in the Reanalysis around 1978–1979 [*Santer et al.*, 1999; *Pawson and Fiorino*, 1999] and the Reanalysis has shown spurious cooling in the early 1990s [*Basist and Chelliah*, 1997], which may be related to the Reanalysis temperatures of the tropical tropopause being warmest compared to the radiosonde temperatures in 1989–1993 [*Randel et al.*, 2000]. The smaller amount of observations from the presatellite era suggests that reanalyses may then have had larger problems. On the other hand, false trends may result from the use of satellite data if these are biased.

[5] According to a recent study by *Trenberth et al.* [2002] on the quality of reanalyses in the Tropics, the European Centre for Medium Range Weather Forecasts (ECMWF) reanalysis (ERA [*Gibson et al.*, 1997]) temperature data shows large stepwise discontinuities in the 80s and in the 90s. NCEP Reanalysis and the latest Microwave Sounding Unit (MSU, version d) temperature data show close agreement. Trenberth et al. conclude that they may be closer to the truth especially since these data sets are to some extent independent. However, they note that it is possible that there are further biases present.

[6] In section 2, we discuss the data sets used in this study and studies on the reliability of the Reanalysis data. In section 3 the calculation of PI is described. In section 4, we describe a method to adjust the Reanalysis temperatures to estimate the effect of the large jump in upper tropospheric temperatures that occurred in the beginning of the satellite era. In section 5, results are shown and discussed. In section 6, conclusions are given.

2. Data

[7] For an accurate calculation of PI, accurate knowledge of SST and temperatures and humidities from the surface to at least the tropopause is required. The data that are used in the PI calculation are described in the following. We first discuss the Reanalysis data and some known problems in that data set. Then we discuss the SST data. We obtained the SST data from http://dss.ucar.edu/data sets/ds277.0/data/recon/data and the NCEP Reanalysis data from http://sgi62.wwb.noaa.gov:8080/reanlm//test.daily.prs/.

[8] NCEP Reanalysis has now been performed for the years 1948–2000. Use of the same data assimilation system for the whole period assures that there are no spurious trends associated with changes in the assimilation system. However, the types and amount of data used as input varies greatly in time. *Kistler and Kalnay* [2000] note that there are three major phases in the global upper air observing system a) the early period starting with the first upper-air rawinsonde observations and ending with the International Geophysical Year (IGY) of 1957–1958, b) the "modern" global rawinsonde network established during the IGY and used almost exclusively until 1978; and c) the advent of a global operational satellite observing system starting in 1979 until the present. *Onogi* [2000] studied the performance of the radiosonde observing system and notes that the quality of the observed values after the 1970s are significantly better than those before the 1960s. Based on the amount and quality of observations we decided to calculate the PI for years 1958–1996 only. We exclude from the analysis years after 1996 since changes in radiosonde types may have caused spurious changes in the relative humidity of the boundary layer (see section 5.1) [e.g., *Guichard et al.*, 2000].

[9] *Mo et al.* [1995] assessed the impact of the satellite data on analysis and forecasts using the NCEP Reanalysis System for the month of August 1985. Their experiments SAT and NOSAT differed only in their use or lack of use of satellite data. In the Tropics, temperatures at 100 hPa were up to 5°C larger in the SAT experiment, in which the NESDIS-derived temperature retrievals and cloud-tracked winds were used, than in the NOSAT experiment. Mo et al. also compared the NOSAT and SAT analyses to radiosonde data and concluded that the SAT analysis was significantly warmer than the radiosonde data in the Tropics.

[10] *Santer et al.* [1999] showed that there was an abrupt change in the mean and the character of the variability of the NCEP Reanalysis lower stratospheric temperatures (i.e., for Channel 4 with the maximum weighting function at 74 hPa). The transition occurred between 1978 and 1979 for the global average and the Northern hemisphere (NH). This is when NCEP began to assimilate satellite-derived temperature retrievals [*Kalnay et al.*, 1996]. In the Southern hemisphere (SH), where satellite data were used from 1977 onward, the transition occurred around 1977–1978. The abrupt increase of temperature was most pronounced in the Tropics and poleward of 30 S (see their Plates 1 and 3). The transition toward warmer temperatures as compared to the gridded radiosonde data set, compiled by *Parker et al.* [1997] at the Hadley Centre for Climate Prediction and Research, (HadRT1.1), is also apparent in the midtroposphere of the Reanalysis data, although it is much smaller than in the lower stratosphere. Note that owing to Santer et al.'s choice of a reference period, any warm bias relative to the radiosonde data in the satellite era must manifest itself in their figures as a cool bias in the presatellite era.

[11] *Pawson and Fiorino* [1999] compared several reanalyses and found that those reanalysis products which assimilated the NESDIS temperature retrievals were several degrees warmer than those which did not. Based on all these studies, the differences between the upper tropospheric and lower stratospheric temperature before and after 1979 seem mostly due to the use of satellite data.

[12] The Reanalysis uses the Global Sea Ice and Sea Surface Temperatures (GISST2.2) from the U.K. Meteorological Office until late 1981 and the NCEP Optimal

Interpolation (OI) weekly SST analysis since late 1981 as boundary conditions. The GISST2.2 over the period 1949–1981 is based on reconstructions using EOFs [*Hurrell and Trenberth*, 1999]. The OI SST analysis technique [*Reynolds and Smith*, 1994] combines in situ and satellite-derived SST data. As we want to minimize the possibility of artificial trends in the calculation of PIs during 1958–1996, we opted to use the same SST analysis for the whole period. We chose the NCEP EOF analyses [*Smith et al.*, 1996]. This reconstructed analysis uses 12 year OI SST anomalies to calculate EOF spatial basis functions. The dominant EOF modes are fit to detrended monthly median SST anomaly statistics from the Comprehensive Ocean–Atmosphere Data Set (COADS [*Woodruff et al*, 1987]). After reconstruction of monthly SST anomalies, the smoothed long-term trend is restored. The analysis has a southern limit of 45°S and a northern limit of 69°N because of lack of data. See *Hurrell and Trenberth* [1999] for a more detailed description of these three different SST analyses.

[13] *Hurrell and Trenberth* [1999] have calculated differences of monthly anomalies among the three data sets averaged over the Tropics (20°S–20°N) and extratropics. Similarly averaged differences of the total SST values are not available. However, their Figure 1 shows that in most of the Tropics the differences between the 30-year climatologies of GISST and NCEP OI are within 0.25°C. The time-mean difference in the SST analyses can affect the time-mean value of PI. In addition, the change of SST analysis used in Reanalysis from GISST2.2 to NCEP OI in 1981 can lead to a jump in the mean value of PI. However, such a change is accounted for by our method of adjusting the Reanalysis temperatures, as will be discussed in sections 4 and 5.

[14] The NCEP EOF SST monthly anomalies show relatively small differences from the NCEP OI monthly anomalies. However, differences between the NCEP EOF and GISST monthly anomalies for 1958–1982 are larger. Therefore we can expect larger errors in the PI calculation before 1982 associated with differences between the SST analysis used by the Reanalysis and that used in our calculation of PI. Differences among SST data sets will be discussed in more detail in section 5.

3. Calculation of Potential Intensity

[15] The technique represents a generalization of that presented by *Emanuel* [1995] to the case of an open-cycle, irreversible heat engine. The description presented by *Emanuel* [1995] has been further modified to account for dissipative heating, which had been neglected in earlier treatments. These modifications are described by *Bister and Emanuel* [1998]. The input to the scheme at each grid point is the sea surface temperature and vertical profiles of pressure, temperature, and mixing ratio from the Reanalysis data. For the purposes of this calculation, the work done against friction in the hurricane outflow is ignored. If the size of the storm is known, this work can be estimated, but it is usually very small unless the storm has an exceptionally large diameter [*Emanuel*, 1986]. We also assume that at maximum intensity, the anticyclone at the storm top is fully developed (zero absolute vorticity) and that the gradient wind may be approximated by the cyclostrophic wind at the radius of maximum winds. When these conditions are satisfied, the thermal wind equation can be combined with an equation governing entropy in the boundary layer in regions near the eyewall where little entrainment of dry air through the top of the boundary layer occurs, resulting in:

$$V_m^2 = c_p(T_s - T_0)\frac{T_s}{T_0}\frac{C_k}{C_D}(ln\theta_e^* - ln\theta_e)|_m \quad (1)$$

where V_m is the maximum gradient wind speed, c_p is the heat capacity at constant pressure, T_s is the ocean temperature, T_0 the mean outflow temperature, C_k the exchange coefficient for enthalpy, C_D the drag coefficient, θ_e^* the saturation equivalent potential temperature at the ocean surface, and θ_e the boundary layer equivalent potential temperature. The last factor in (1) is evaluated at the radius of maximum winds. This expression is the dimensional equivalent of equation (13) of *Emanuel* [1995], and can also be written:

$$V_m^2 = \frac{T_s}{T_0}\frac{C_k}{C_D}\oint Tds \quad (2)$$

where s is the entropy ($=c_p ln\theta_e$) and the closed cycle is for a parcel beginning at the ambient boundary layer value of θ_e and winding up at saturation. Entropy is acquired at temperature T_s and exported at temperature T_0. It is important to note that (2) does not imply that air parcels actually become saturated at sea level under the eyewall.

[16] The integral in (2) is also equal to the convective available potential energy [see *Emanuel*, 1994, equation 6.4.2]. Thus (1) can be written:

$$V_m^2 = \frac{T_s}{T_0}\frac{C_k}{C_D}[CAPE^* - CAPE]|_m \quad (3)$$

where *CAPE** is the convective available potential energy of air lifted from saturation at sea level in reference to the environmental sounding, and *CAPE* is that of boundary layer air. Both quantities are evaluated near the radius of maximum wind. Note that the effect of dissipative heating comes in term T_s/T_0. With dissipative heating, the efficiency of the tropical cyclone is $(T_s - T_0)/T_0$ instead of $(T_s - T_0)/T_s$. The latter of these efficiencies is for a reversible Carnot engine which does work on an external body.

[17] To evaluate (3) it is first necessary to determine the surface pressure at the radius of maximum winds, needed to calculate the saturation mixing ratio necessary for *CAPE**. To do so, we use a combination of gradient wind balance and thermal wind balance in the outer region of the hurricane. This is equation (6) of *Emanuel* [1995], whose dimensional equivalent (assuming cyclostrophic balance at the radius of maximum winds) is:

$$c_p T_s ln\frac{p_0}{p_m} = \frac{1}{2}V_m^2 + CAPE|_m \quad (4)$$

where p_0 is the ambient surface pressure and p_m is the surface pressure at the radius of maximum winds.

[18] To calculate $CAPE|_m$, the mixing ratio and temperature of the boundary layer under the eyewall is needed. Following *Emanuel* [1995], we assume that the surface

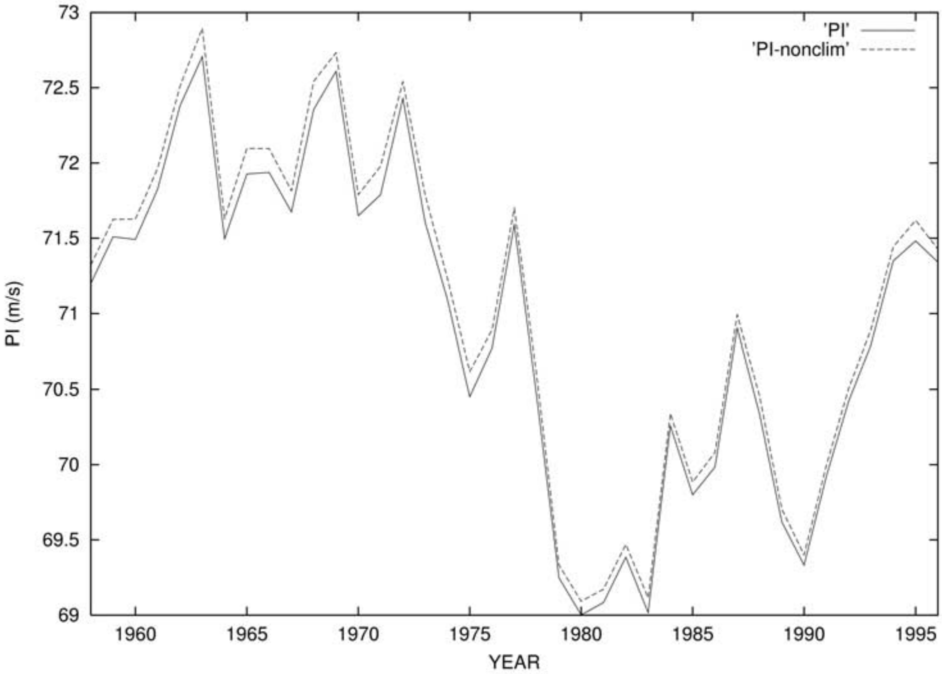

Figure 1. Annual average of daily PI values that exceed 40 m s^{-1} over the whole globe (solid) and excluding those regions where the SST data consists of climatology only, i.e., outside 45°S–69°N (dashed).

temperature is T_s and that the relative humidity is constant from the outer region to the outer edge of the eyewall. Since the pressure under the eyewall is lower than ambient, this entails a small inward increase in mixing ratio. (Experiments assuming constant mixing ratio rather than constant relative humidity yield only slightly different results.) For this reason, $CAPE|_m$ is a little larger than the $CAPE$ of ambient boundary layer air. The assumption of constant relative humidity in the outer region boundary layer is well supported by numerical simulations.

[19] Given a value of C_k/C_D, the sea surface temperature, and an ambient profile of virtual temperature, (3) and (4) constitute closed relations for V_m and p_m. $CAPE$ is calculated by a reversible adiabatic parcel lifting algorithm. Owing to the pressure dependence of $CAPE$* and $CAPE|_m$, (3) and (4) must be solved iteratively. This converges for all input we use from the Reanalysis data.

[20] The NCEP EOF SST data set consists of monthly SST analyses with a resolution of 2° in both latitudinal and longitudinal directions with data outside 45°S–69°N consisting of climatology only. The Reanalysis daily data have a resolution of 2.5°in both directions. Temperature and pressure from the Reanalysis from 13 vertical levels between 1000 hPa and 70 hPa is used in the calculation of PI. The 1000 hPa height is used to calculate the surface pressure. Specific humidity is available for the lowest 8 levels only. At higher levels, the relative humidity is assumed to be 50%. This assumption has a negligible effect on the value of PI.

[21] The value of PI in terms of maximum wind speed is calculated for each atmospheric data location. First, the closest SST point is checked to see whether it is land or ocean. For land points and if the SST is less than −1.78°C PI is not calculated. To estimate the SST at the atmospheric point, a one-over-distance weighted average of SST values over the four closest points is calculated. If the latitude or longitude of two of the SST data locations is the same as that for the atmospheric data then only these two SST values are used in the calculation. If any of the four closest SST data points happens to be located on land or has a SST value of less than −1.78°C, then that value is not used in calculating the average. We also tried using bilinear interpolation to get the SST value to use in the PI calculation; this gave similar results.

4. Adjustment of Reanalysis Temperatures

[22] PI is calculated for each day and grid point of the Reanalysis data. Those values that exceed 40 m s^{-1} are averaged over the globe and over each year to get the annual mean global PI. The annual average, global mean values are shown in Figure 1. They should be compared to the annual average SST between 27°S and 27°N, where the mean SST is about 26°C. The annual mean SST was calculated for each grid point and the values between 27°S and 27°N were then averaged. The result, referred to as "tropical SST" in the following, can be seen in Figure 2. There is a large drop in PI both in an absolute sense and as compared to the tropical SST, starting in the late seventies. Exclusion of data in the region where SST is purely climatological, i.e., outside 45°S–69°N, does not affect the result (Figure 1). A drop in the PI similar to that in the global PI is also found in different ocean regions (not shown).

[23] As an increase of the upper tropospheric temperature not accompanied by an increase in the SST would result in a decrease of PI, the drop in the PI could result from the

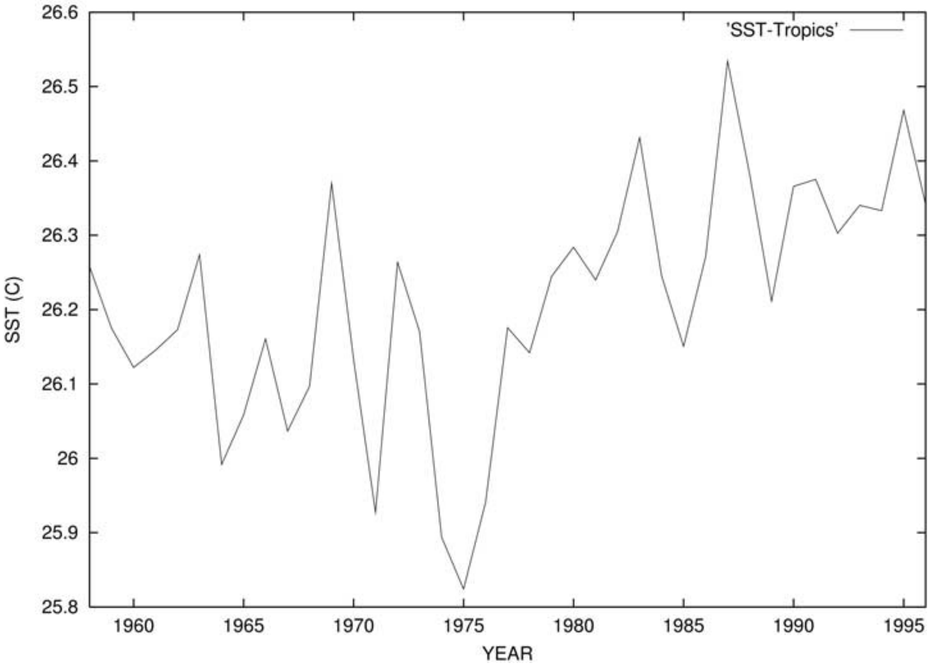

Figure 2. Tropical SST. Average of annual mean SST values over grid points between 27°S and 27°N.

spurious increase in the Reanalysis temperature around 1979, discussed in section 2. Ten-year averages of temperature between 27°S and 27°N were calculated to see how the temperature changed between the years before and after the PI drop. The increase of the upper tropospheric and lower stratospheric temperature in the Reanalysis data in the late seventies can clearly be seen in the ten-year averages (Figure 3). Also the annual mean temperature averaged between 27°*S* and 27°N shows a large jump in 1979 (Figure 4). Magnitude of the temperature jump between year pairs with similar tropical SST, to be discussed shortly, reaches its maximum, about 5 K, in the equatorial mid-Pacific (not shown).

[24] To estimate the effect of the spurious temperature increase on PI, the increase should be subtracted from the temperatures in the satellite era. First, we calculated 9-year

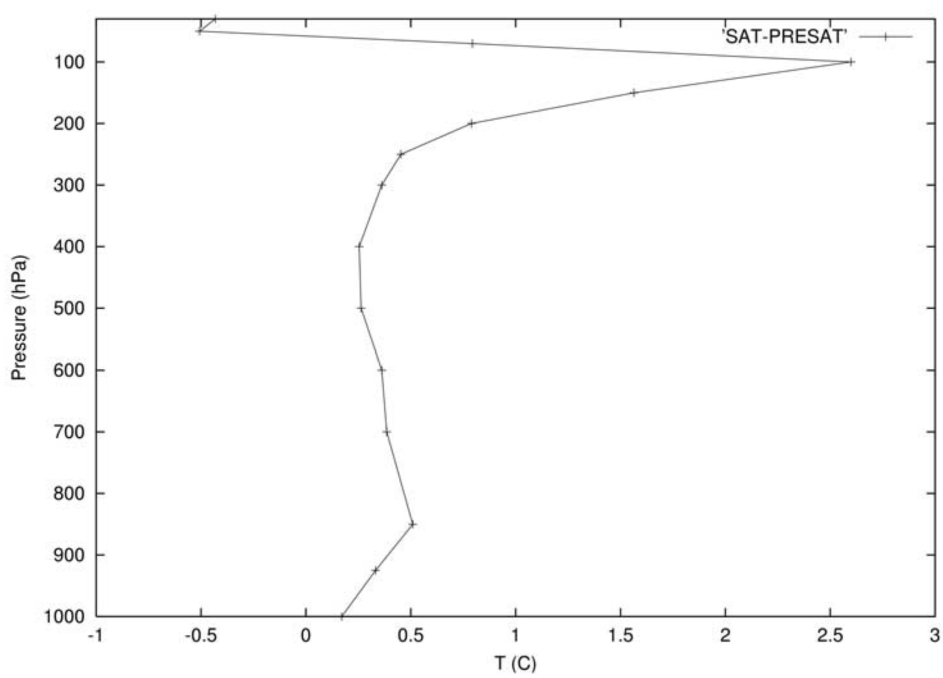

Figure 3. Difference of ten-year averages of temperature between years 1980–1989 and 1965–1974 in the Tropics (27°S–27°N).

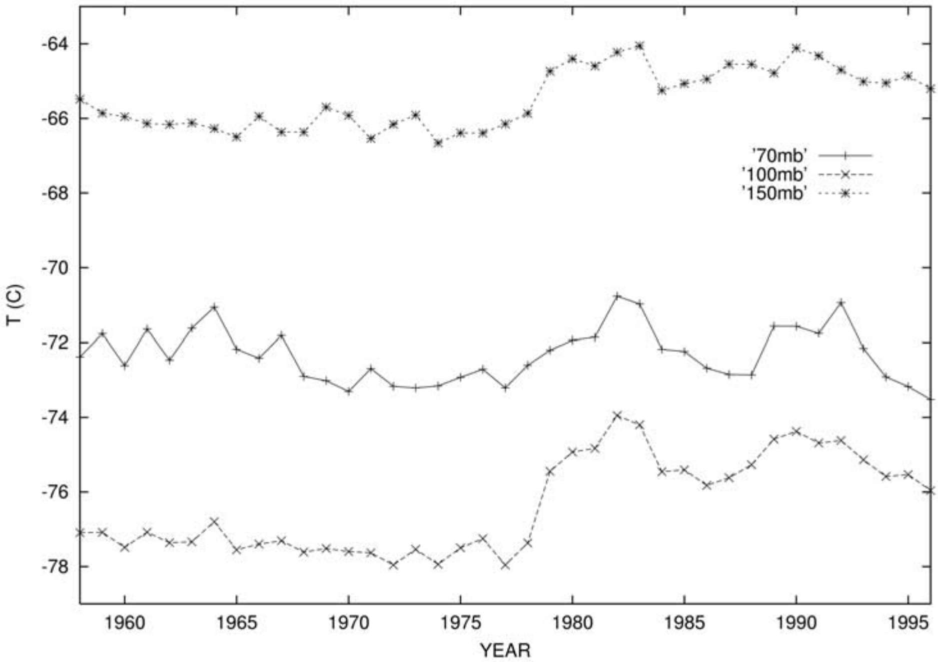

Figure 4. Annual average of temperature over grid points between 27°S and 27°N at the pressure of 70 (pluses), 100 (crosses), and 150 hPa (asterisks).

averages of temperature before and after 1979 to get the difference to be subtracted from temperatures after 1979. This adjustment increased the global PI by about 3 m s^{-1}. However, the tropical SST has increased after 1976 [*Hurrell and Trenberth*, 1999], which can also be seen in Figure 2. Part of the change of the temperature between the 9-year averages can thus be real and associated with the increased SST. Therefore, we opt to use the change of the annual mean temperature between years with similar tropical SST (Figure 2) to get an estimate for the spurious temperature increase.

[25] Pairs of years with similar tropical SST are listed in Table 1. The table also shows the change of tropical SST, SST averaged over the PI region (referred to as "SST_{PI}" in the following), the Southern Oscillation Index (SOI), and the SST change's absolute value averaged over 27°S–27°N. The SST_{PI} represents an average over all locations and times for which the PI exceeded 40 m s^{-1}. It should be noted that SST_{PI} is therefore an average over a region that changes in time and this averaging region depends on the same factors that affect the PI. As an example, an error in the Reanalysis temperatures would affect the SST_{PI} as well as the PI, as will be discussed later in this section. However, SST_{PI} is a useful measure of SST since it is calculated using exactly the same averaging region as for the PI. Therefore, PI and SST_{PI} can compared to find occurrences of PI variations that do not seem to be related to SST variations in the same region.

[26] The magnitude of the tropical SST change is less than or equal to 0.04°C for all pairs of years. The magnitude of the change of the SST_{PI} is less than 0.2 for all other pairs except 72–80 for which it is 0.3°C. The magnitude of the change of SOI is less than 0.5 except for 69–88 for which it is 1.4. Year pair 69–88 also shows the largest average of the absolute value of the SST change.

[27] New estimates of PI were calculated for the satellite era in the following way. First, the change of the annual mean temperature from the earlier to the later year was calculated for each year pair and for each grid point. This change was subtracted from the daily Reanalysis temperature values in 1980–1999. PIs were then recalculated using the adjusted temperatures.

[28] The results are shown in Figure 5. It can be seen that PIs vary greatly depending on which pair of years is used in the calculation of the temperature change. While the curve with the lowest values shows a PI that is only 1.25 m s^{-1} larger than the nonadjusted PI for 1980, the one with the highest values shows a PI that is 3 m s^{-1} larger than the nonadjusted PI for the same year. We will shortly discuss the possible reasons for the differences among these curves.

[29] With no adjustment of the Reanalysis temperatures, the change of SST_{PI} (Table 1) is positive for all cases other than 73–89 and the average decrease of PI from the earlier to the later year of each year pair is 2.5 m s^{-1} (Figure 1). Both changes are consistent with a spurious increase of atmospheric temperature. Namely, assuming that the true SST is the same for the two years, an increase in atmospheric temperatures, whether spurious or real, would result in a decrease in PI. Then the mean SST for the region in

Table 1. Changes Between Earlier and Later Years of Year Pairs

Years	Tropical SST, °C	SST_{PI}, °C	SOI × 10	|SST|, °C
72–80	0.020	0.301	4.6	0.305
69–90	−0.004	0.139	3.0	0.246
72–86	0.007	0.141	4.4	0.298
73–89	0.040	−0.031	−0.1	0.278
70–85	0.018	0.052	−2.3	0.260
69–88	0.012	0.156	14.1	0.408

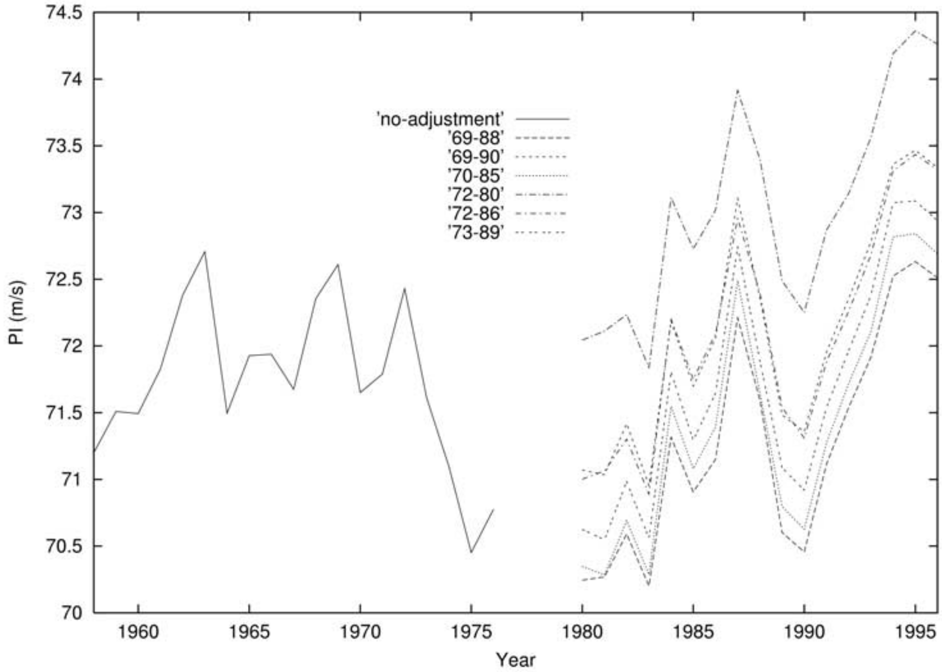

Figure 5. Annual average of daily PI values that exceed 40 m s^{-1} over the whole globe with no adjustment of Reanalysis temperatures (solid), and with adjustments using different year pairs as explained in the text. From top to bottom year pairs are 72–80, 69–90, 72–86, 73–89, 70–85, and 69–88.

which the PI exceeds 40 m s^{-1} should have increased, as has happened.

[30] Even with the adjustment of temperatures, PI decreases from the earlier to the later year of each year pair (Figure 5). This decrease can be real or spurious. Our method of using annual averaged temperatures of years with similar mean tropical SST may not properly account for the spurious temperature increase. On the other hand, years with similar tropical SST do not need to show similar mean PI. Indeed, if there is an EOF mode in the SST field with an abrupt change toward a warmer tropical Eastern Pacific and a colder extratropical central North Pacific in 1976–1977, as suggested by *Zhang et al.* [1997] (see also *Mantua et al.* [1997] for the variation of the Pacific Decadal Oscillation [PDO] index representing the long-term variation in the Pacific SST pattern) then the mean PI may have changed even though the mean tropical SST has stayed the same. For these reasons, we stress that *the adjustment of temperatures using year pairs is only a way to estimate the sensitivity to the temperature adjustment and not a way to correct the Reanalysis data*. Therefore also, *the time series of PI cannot be used to estimate trends over years 1958–1996. This pertains to all figures showing PI*, *i.e*., *Figures 5 and 7–10*.

[31] The change in the SOI (Table 1, values obtained from http://tao.atmos.washington.edu/pacs/additional_analyses/soi.html) was so dramatic between years 1969 and 1988 that the temperature change between these two years may not be representative of the increase which occurred around 1979. Namely, a change in the SOI implies changes in the geographical patterns of SST and the larger the change in the SOI is the larger the associated local changes in SST are. This is supported by the fact that the largest value of the mean absolute change of the SST occurred for year pair 69–88 (Table 1). Since the atmospheric temperatures are affected by the change in the SST, the temperature change between years 1969 and 1988 should probably not be used in the PI calculation.

[32] Of interest is year pair 73–89 for which the SST_{PI} has decreased and year pair 72–80 for which it has increased much more than for other pairs of years. From Figure 2 it can be seen that the local maximum in the tropical SST occurs in 1972 and SST is smaller in 1973. However, the SST_{PI} (Figure 6) reaches a maximum only in 1973. Thus in the region with hurricane potential, the maximum SST may occur in 1973 and not in 1972, which is possible if the SST change does not occur simultaneously everywhere. In any case, the SST_{PI} probably also affects the atmospheric temperatures in the same region so that the atmospheric temperatures may have warmed between 1972 and 1973 in the PI region whereas they should have cooled if we only look at the tropical SST (Figure 2). Therefore, subtracting the atmospheric temperatures of 1972 from the temperatures of 1980 probably results in a larger temperature change in the PI region than subtracting the atmospheric temperatures of 1973 from the temperatures of 1989. This could partly explain why using year pair 72–80 results in much larger PIs after 1980 than using year pair 73–89. Similarly, this difference in the behavior of tropical SST and SST_{PI} during 1972–1973 is probably also behind the decrease of the SST_{PI} between years 1973 and 1989.

[33] Note, however, that if the larger increase of SST_{PI} between 1972 and 1980 than between other years (Table 1) was due solely to the reason just discussed, then the increase

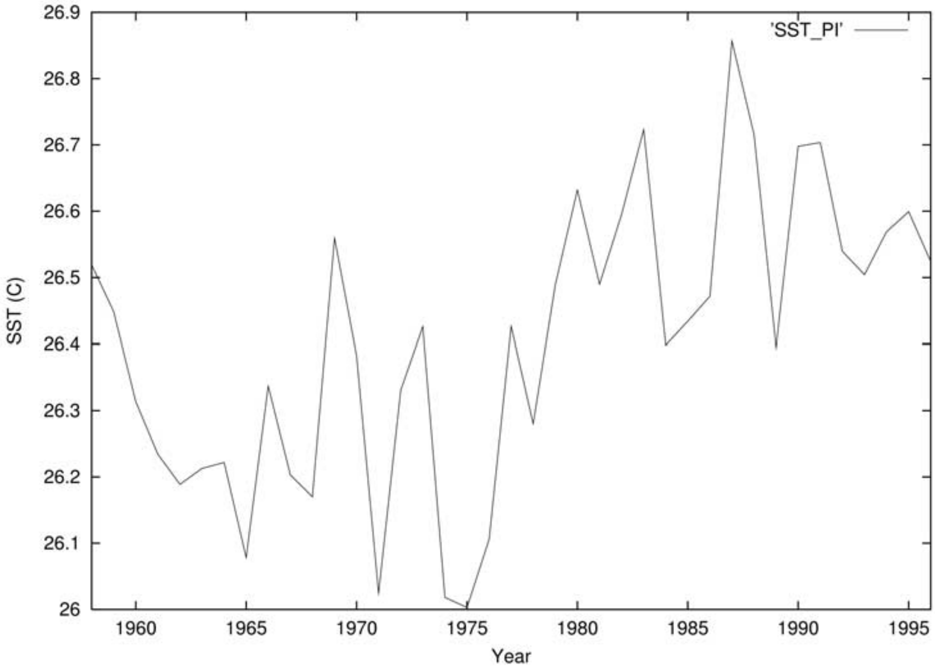

Figure 6. SST_{PI}. Annual average of daily SST values of those grid points where PI, calculated with no adjustment of the Reanalysis temperatures, exceeds 40 m s^{-1}.

of SST_{PI} should also be larger between years 1972 and 1986. However, it is only half of that for years 1972 and 1980. A larger positive bias in the Reanalysis temperature in 1980 than in 1986 could be a reason for this difference and such a bias could also partly explain why PIs obtained with year pair 72–80 are so much higher than those obtained with other year pairs. Yet, the authors have found no indication in the literature that the jump to warmer temperatures that occurred in 1978–1979 would have reversed in the early 1980s. However, *Randel et al.* [2000] showed that during the satellite era the tropical tropopause in the Reanalysis was coldest in 1986–1987 when compared to radiosonde temperatures. Using year 1986 as the latter year of a year pair would then lead to underestimates of PI in the satellite era if there were no other problems. Note, that there is also a possibility of real cooling in the 1980s and 1990s associated with ozone decrease, which will be discussed in section 5. It is also possible that the change of the SST analysis used in the Reanalysis would contribute. However, its role is likely to be small since, as will be shown in section 5, there is no sudden large change in the correspondence of PI and SST_{PI} in 1981.

[34] We choose one year pair to study the PI behavior more closely using the adjusted temperatures. Year pair 69–88 is not appropriate due to the large increase in the SOI. Year pairs 72–80, 72–86, and 73–89 may all suffer from the abovementioned problem with tropical SST peaking in 1972 and SST_{PI} peaking only in 1973. This leaves two year pairs: 69–90 and 70–85. We choose year pair 69–90 since it shows the smallest mean value of the absolute change of the SST in the region between 27°S and 27°N and also the smallest mean change of the SST in the same region (Table 1). Note also that the Pacific Decadal Oscillation index [*Mantua et al.*, 1997] is about the same for years 1969 and 1990. However, there may be a problem with using 1990 as the latter year. *Randel et al.* [2000] showed that the tropical tropopause in the Reanalysis was warmest in 1989–1993 when compared to radiosonde temperatures. This may indicate that using year pair 1969–1990 could lead to overestimates of PIs in the satellite era (see section 5.2 for a discussion of the representativeness of PIs obtained with year pair 1969–1990).

[35] The averaged global PI for year 1980 using all pairs of years except 69–88, is 71.0 m s^{-1} which is 2 m s^{-1} larger than the unadjusted PI for the same year. If also 72–80 is dropped as an outlier, the average is 70.76 m s^{-1}. For 69–90 the PI for 1980 is 71.07 m s^{-1}, slightly more than the average for the five year pairs, omitting 69–88 only.

5. Results

5.1. Sensitivity to Errors in Data

[36] When it comes to inherent biases, there is no single error estimate for the Reanalysis data that would cover the years 1958–1999. The quality of the Reanalysis changes as the amount and quality of input data sources change. In addition to problems with satellite data, radiosonde data also contains time-dependent biases. *Gaffen* [1994] and *Gaffen et al.* [2000] showed that temperatures measured by radiosondes can have errors of 1 K or even more. Moreover, *Angell* [2000] showed that exclusion of three Indian and four French and French ex-colonial stations with anomalous standard deviations changed the sign of the tropical 300–100 hPa temperature trend for 1959–1998. However, at least the Indian stations have been excluded from the Reanalysis, at least in one particular year [*Santer et al.*, 1999]. Errors in the radiosonde data vary greatly spatially and temporally and unfortunately corrections are not easy to

Table 2. Sensitivity of PI and SST_{PI} on Some Variables[a]

Change of Certain Variables	Change in PI, m s^{-1}	Change in SST_{PI}, °C
Δ SST = −0.1°C	−0.7	−0.03
Δ T(925–200 hPa) = −0.25*C*	0.02	−0.1
Δ T(925–100 hPa) = −0.25*C*	0.3	−0.1
Δ T(925–70 hPa) = −0.25*C*	0.3	−0.1
Δ RH(bound.layer) = −(5–10)%	1.7	−0.01

[a]Relative humidity was decreased by 10% at 1000 hPa and by 5% at 925 hPa.

make. However, since instrument (and other) changes have not occurred simultaneously at all radiosonde stations, we can expect that the assimilation of the Reanalysis tends to smooth out partially these errors especially in the Tropics, where the Rossby radius of deformation is large.

[37] *Santer et al.* [1999] have calculated root mean square (RMS) differences between the HadRT1.1 sounding data set and both full-coverage NCEP Reanalysis data and the same masked with the HadRT1.1 coverage. For the equivalent Channel 4 temperature (with maximum weighting at 74 hPa) in the satellite era, the global RMS difference of seasonal mean anomalies is 0.13°C using the masking with the Reanalysis data and 0.26°C with no masking. For the presatellite era, Santer et al's results are affected by the temperature jump occurring around 1979 due to their choice of the reference period being the satellite era.

[38] *Hurrell and Trenberth* [1999] showed that in the Tropics, GISST and NCEP EOF monthly SST anomalies differ by up to about 0.2°C in the presatellite era. After 1982, the differences between NCEP EOF and NCEP OI and, on the other hand, GISST and NCEP OI monthly anomalies are mostly below 0.1°C.

[39] The authors have found no error estimate for the boundary layer relative humidity in the Reanalysis. However, *Ross and Gaffen* [1998] showed evidence for the observed drying of the tropical troposphere from 1979 to 1995 [*Schroeder and McGuirk*, 1998] to be spurious and related to introduction of faster-response sensors and change from VIZ sondes to Vaisala sondes. In 1997–1998, Vaisala's RS80H sondes were introduced in U.S. stations. These sondes have a dry bias of about 6% at relative humidities of about 80% if the sonde has been stored for as long as six months. Although correction algorithms now exist for the humidity data from Vaisala's sondes [*Wang et al.*, 2002], the errors can affect the Reanalysis trends of relative humidity. For this reason, we decided to exclude years after 1996 from our analysis.

[40] We estimate the sensitivity of PI and SST_{PI} for decreases in SST, the atmospheric temperature, and the boundary layer humidity by applying these changes separately and calculating the PI and SST_{PI} anew for years 1990–1999. The results are shown in Table 2. Decreasing SST by 0.1°C decreases the mean PI by 0.7 m s^{-1}. A relatively large decrease in the boundary layer humidity (10% at 1000 hPa and 5% at 925 hPa) increases the mean PI by 1.7 m s^{-1}. The mean PI is not sensitive to a temperature change in a layer from 925 to 200 hPa, but it is sensitive to a temperature change at 100 hPa. We will use these sensitivities in the following to estimate the effect of several suspected time-dependent biases in the data and the effect of other time dependent factors that affect upper tropospheric temperatures.

5.2. Global Potential Intensity

[41] The level of PIs in the satellite era is very sensitive to which year pair is used. The PIs shown by the highest curve

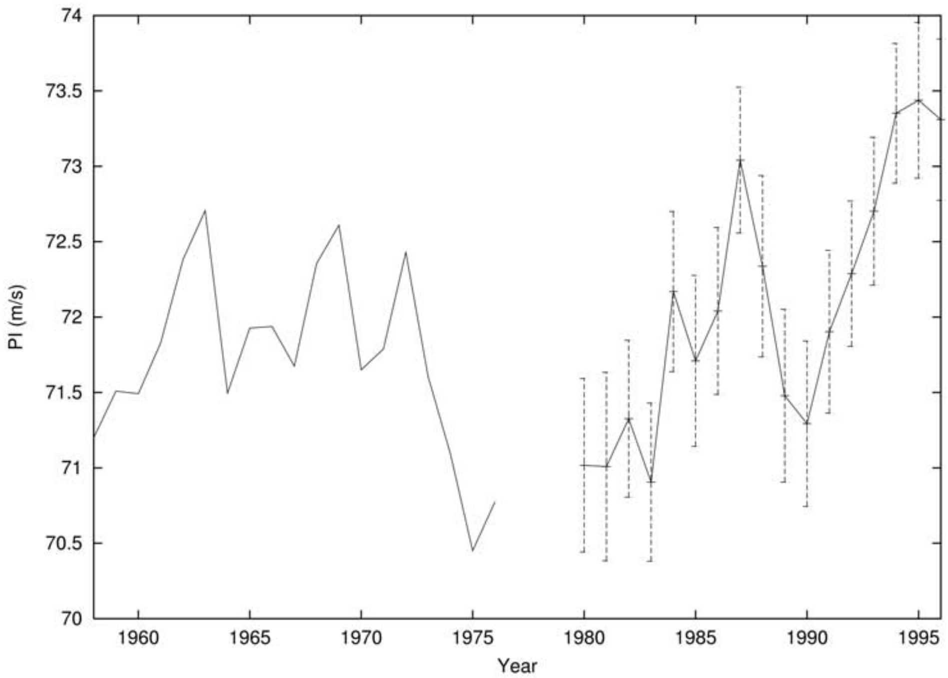

Figure 7. Same as Figure 5 except that for years after 1979, the mean and standard deviation of PI obtained using other year pairs than 69–88 are shown. The error bars show the standard deviation and reflect the uncertainty associated solely with the adjustment of the Reanalysis temperatures.

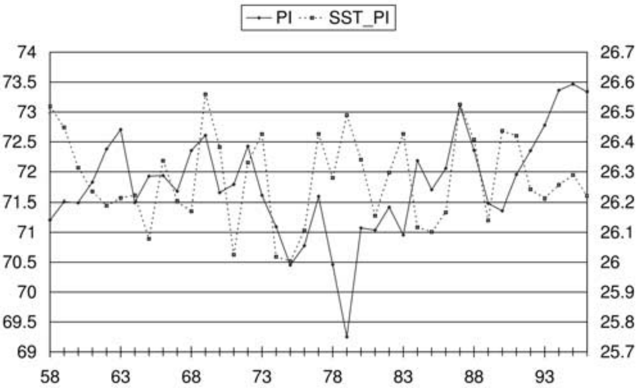

Figure 8. PI and SST_{PI} for 1958–1996 obtained by adjusting temperatures for 1980–1996 using year pair 69–90. Scale on the left is for PI (m s^{-1}) and on the right for SST (°C). Since some satellite data were assimilated starting in 1977, values for 1977–1979 should be disregarded.

in Figure 5 are about 1.8 m s^{-1} larger than those shown by the lowest curve. An estimate of the uncertainty associated with the method of using year pairs to adjust the Reanalysis temperatures can be obtained by calculating the mean and standard deviation of PIs obtained with different year pairs. Figure 7 is similar to Figure 5 except that for the satellite era, the mean value and standard deviation of PIs from other year pairs than 69–88 are shown; year pair 69–88 was omitted owing to the large difference of the SOI between these two years. The error bars, therefore, reflect the uncertainty associated solely with the adjustment of the Reanalysis temperatures and do not contain any information regarding other data problems, e.g., biases in radiosonde temperatures or humidities.

[42] Note that the mean PI follows rather closely the PI obtained using year pair 1969–1990. As discussed in section 4, it is possible that using this year pair results in values that are too large. As the only year pair that was not associated with any known problems, namely 1970–1985, also shows values of PI that are about 0.6 m s^{-1} smaller than the mean PI in Figure 7, the mean value of PI may overestimate the real PI in the satellite era. On the other hand, if there is cooling related to ozone decrease in the 1980s (see the end of this section for more information about the ozone related cooling), use of year pairs with the latter year from after 1980 would result in too low values of PI in the satellite era.

[43] In Figure 8, we have overlaid the adjusted PI and SST_{PI} obtained using year pair 69–90. In Figure 9, the same quantities are shown for the Tropics only (30°S–30°N). In overlaying the two curves, we tried to make the correspondence of PI and SST_{PI} as good as possible for the time period before 1979. The scale for the SST_{PI} was made 0.2 times the scale for the PI in Figures 8 and 9. Note that since assimilation of selected satellite data was started already in 1977 in the Southern hemisphere, the values for PI and SST_{PI} are unreliable for the period 1977–1979.

[44] Figure 8 shows that the year-to-year variations in PI can be as large as 1 m s^{-1}. Of particular interest is the large increase of PI occurring in the 1990s. PI increased from 71.4 in 1990 to 73.4 m s^{-1} in only four years. The corresponding increase in the Tropics (Figure 9) was even larger. The largest values of PI in the satellite era occur in 1987 and 1994–1996. In the rest of this section, we will discuss known biases in the Reanalysis and SST data as well as changes in the atmospheric temperature not related to SST and their possible effect on the calculated PI.

[45] Unfortunately, not much is known about the errors in the Reanalysis temperatures in the presatellite era. However, it is clear that use of different SST data sets in the Reanalysis and in our calculation of PI, can cause errors in PI. Moreover, errors can be introduced by the change of the SST analysis used in the Reanalysis from GISST2.2 to NCEP OI in 1981. However, for all year pairs, except 72–80, the later year is from after 1981 and thus the effect of the change of SST analysis on the atmospheric temperatures of the Reanalysis is at least crudely accounted for by our method of adjusting the Reanalysis temperatures.

[46] Figure 7 in the work of *Hurrell and Trenberth* [1999] shows the difference between the monthly SSTs from GISST and NCEP EOF. By comparing the difference in the Tropics with the correspondence of the PI and SST of the PI region, we can see that for 1958–1960, 1969, 1973, 1980 the GISST SST is larger than the NCEP EOF SST. For the same years, the PI values would seem to be too weak when compared to the SST_{PI}. The GISST SST is smaller than the NCEP EOF for years 1962, 1964–1965, 1971, 1976, and except for 1964 and 1976 which show small differences overall, the PI values would seem to be too large when compared to the SST_{PI}. Thus it seems that a bad correspondence between the PI and SST_{PI} can often be related to the problems in the SST analysis in the presatellite era. Why would larger GISST than NCEP EOF SST imply too weak PIs and vice versa? Since the SST analyses can be expected to affect the Reanalysis atmospheric temperatures, the larger GISST SSTs are probably associated with larger atmospheric temperatures as well. In other words, when the GISST SST is larger than the NCEP EOF SST used in the PI calculation, the atmospheric temperatures used in the PI calculation are too large compared to the NCEP EOF SST values. The sensitivity experiments in section 5.1 showed that an error in SST of 0.1°C, similar in magnitude to differences in the two SST analyses, can change the PI by 0.7 m s^{-1}.

[47] *Hurrell and Trenberth* [1999] show also the difference of the monthly anomalies of SST between NCEP OI

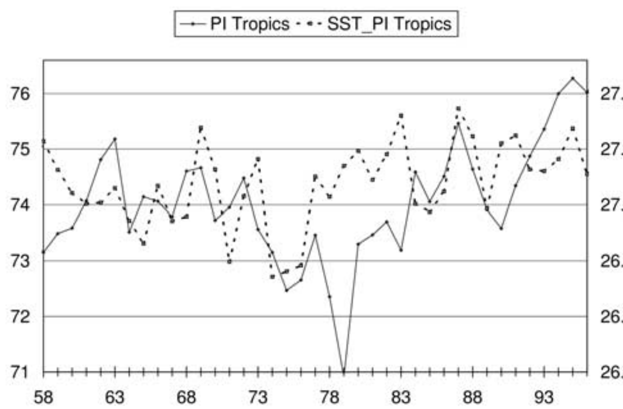

Figure 9. Same as Figure 8 but for the Tropics (30°S–30°N).

and NCEP EOF for years after 1981. The difference between these two analyses is small, less than 0.1°C. However, Hurrell and Trenberth note that there is a warming of GISST analysis relative to both NCEP analyses starting in mid- to late 1980s. It is argued that because processing of in situ data (in case of NCEP EOF) and problems with the satellite data (in case of NCEP OI) the EOF and OI are getting too cold as compared to GISST. The magnitude of the relative cooling is on the order of 0.1°C [*Hurrell and Trenberth*, 1999, Figure 7]. The resulting error in the PI depends on how much this cooling affects the tropospheric temperatures of the Reanalysis. Were there no effect on the tropospheric temperatures, the resulting error in PI would be maximal and as given by our sensitivity experiment in section 5.1. The qualitative effect is that a negative error in SST results in an underestimate of PI, especially in the 1990s for which the difference of the GISST and NCEP analyses is larger than for the 1980s.

[48] When it comes to Reanalysis temperatures, *Basist and Chelliah* [1997] observed a downward drift in the temperatures as compared to the MSU Channel 2 temperature (with maximum weighting at the pressure of 595 hPa) starting in 1991. The magnitude of the difference was about 0.2°C. Also *Hurrell and Trenberth* [1998] noted that the NCEP temperatures have been too cold since 1991–1992. According to Basist and Chelliah, the cold trend in the Reanalysis seems to start reversing in 1995. It is not clear what has happened to the cold trend since 1995. Moreover, if this downward drift is restricted to lower troposphere, it may not much affect the calculated PI (see section 5.1). If, on the other hand, the drift occurs as high as at 100 hPa, then a spurious cooling of 0.25°C could lead to an overestimate of PI of 0.3 m s^{-1} and an underestimate of SST_{PI} of 0.1°C. Such an error in the Reanalysis temperatures could then partly explain why the PIs are getting more intense relative to the SST_{PI} during the early 1990s.

[49] As discussed in section 4, in 1989–1993 the Reanalysis temperatures of the tropical tropopause were anomalously warm by 1 K [*Randel et al.*, 2000] as compared to radiosonde temperatures. If the 100 hPa temperatures were also anomalously warm by the same amount, PI would be too weak by about 1.2 m s^{-1} (Table 2) and SST_{PI} would be too high during these years. Hence, this supposedly spurious temperature anomaly could explain the low values of PI and high values of SST_{PI} around 1990–1992.

[50] As satellite humidity measurements are not used in the Reanalysis (W. Ebisuzaki, NCEP, personal communication, 2002), the data source type does not vary as much as in case of temperature. Still, changes in the availability and type of humidity measurements can cause errors in the resulting humidity fields. As discussed in section 5.1, spurious drying in the radiosonde measurements has occurred in 1979–1995. However, it is not clear to what extent the Reanalysis is affected by this drying.

[51] PI can also be affected by variations in atmospheric temperature that are not tied to SST. Such changes can result in differences in the variation of PI and SST_{PI}. Volcanic eruptions cool the troposphere and could therefore increase PI for the same SST. Since 1979 there have been two large eruptions of volcanoes: the 1982 eruption of El Chichon and the 1991 eruption of Mt. Pinatubo. The Mt. Pinatubo eruption was very efficient in sending aerosols into the stratosphere. As can be seen in Figure 8, after 1991 the PI increased as compared to SST_{PI}, which is qualitatively consistent with a larger relative cooling of the upper troposphere than SST. However, the PI remained large as compared to SST_{PI} until 1996, long after the cooling associated with Mt. Pinatubo had subsided. Therefore, Mt. Pinatubo cannot be the only reason for the divergence of the curves for PI and SST_{PI} in 1990s.

[52] *Onogi* [2000] notes that radiosondes show that temperature at 100 hPa in the 1990s is decreasing steeply in many countries and in the Tropics, radiosonde data show cooling of about 2 K in the layer from 50 to 100 hPa from 1979 to 1998 [*Angell*, 2000]. *Bengtsson et al.* [1999] have recently conducted experiments with the Max-Planck-Institut model, ECHAM4, to study effects of the decrease of the stratospheric ozone on tropospheric temperatures. Their Plate 2 shows that the decrease of the stratospheric ozone also cools the upper troposphere in the Tropics, although much less than at higher latitudes. The cooling from 1979 to 1997 in the model associated with the ozone decrease is around 0.2–0.4°C per decade between 100 and 200 hPa in the Tropics. Note, however, that Bengtsson et al. used an observed trend of ozone distribution for 1979 to 1993, linearized it and then extended it until 1997. Qualitatively similar results have been obtained by *Langematz* [2000] with the Berlin Climate Middle Atmosphere Model. Note that the sensitivity experiments in section 5.1 suggest that temperature in the upper troposphere affects PI much more than in the lower troposphere. This cooling can therefore cause an increase in PI and decrease in SST_{PI} that might partially explain the different behavior of PI and SST_{PI} in the mid-1990s.

[53] When comparing Figures 8 and 9, it should be noted that certain problems with the data may actually be quite large in the extratropics in the region where PI often exceeds 40 m s^{-1}, i.e., over the Gulf Stream and the Kuroshio current. Namely, studies by *Andersson et al.* [1991] and *Kelly et al.* [1991] show that both the statistical retrieval method used by the NESDIS until 1988 and the physical retrieval method used since 1988 have large problems over these two currents. Moreover, there is an enhanced bias of the Reanalysis temperatures in the 1990s as compared to the MSU temperatures over the Kuroshio current [*Basist and Chelliah*, 1997, Figure 15]. And still, the comparison of the monthly SST anomalies in three SST analyses by *Hurrell and Trenberth* [1999] shows that the correspondence is much worse outside the Tropics (20°S–20°N) than in the Tropics, especially in the early to mid-1990s. These error sources outside the Tropics may all contribute to the bad correspondence of the global PI and SST_{PI} in the 1990s. On the other hand, a natural reason for the worse correspondence of PI and SST_{PI} outside the Tropics is that atmospheric temperatures cannot be expected to follow local SSTs there as much as in the Tropics.

[54] In section 4, we discussed reasons why year pair 72–80 gives so much larger values for PI than other year pairs. One suggested reason was that there may have been spurious or real cooling of Reanalysis temperatures in the early 1980s not associated with changes in SST. Such a cooling could also be a reason for the relatively bad correspondence of PI and SST_{PI} in 1980–1983, especially in the Tropics (Figure 9), where one would expect a better

correspondence between the two than in the extratropics. For example, ozone related cooling of 0.6 K per two decades [*Bengtsson et al.*, 1999, Plate 2] could have decreased SST_{PI} by 0.26°C and increased PI by 0.7 m s^{-1}. This could explain why the curve for PI is mostly below the curve for SST_{PI} in the early 1980s and vice versa in the late 90s. If there has been real cooling in the early 1980s then year pairs other than 72–80 would give PIs that are too low, since for them the later years are from 1985 or after. In principle, however, it is also possible that the relative increase of PI as compared to SST_{PI} in Figures 8 and 9 is related to relative humidity and not temperature.

5.3. Potential Intensity in A Few Ocean Regions

[55] How representative is the global PI behavior of that in the different hurricane belts? Although, PI can easily be calculated for any ocean region, smaller averaging regions can be more influenced by inaccuracies in, for example, radiosonde data. *Gaffen* [1994] studied temporal inhomogeneities in radiosonde data due to changes in sensors, changes in solar radiation corrections to the data, and changes in the length of the train between the balloon and the instrument package. She concluded that the changes can cause discontinuities in the temperature records of several tenths to several degrees Celsius. She found that at least 27 of 63 stations in the network used by *Angell* [1988, 1991] were affected by inhomogeneities which were concentrated in the Tropics and Southern Hemisphere. As an example, bias corrections to data from radiosonde stations in Australia and New Zealand after 1979 have reduced the magnitude of stratospheric cooling by as much as 3°C over the years 1979–1996 [*Parker et al.*, 1997] in the corresponding region. Such bias corrections were not made in the radiosonde data assimilated by NCEP [*Santer et al.*, 1999]. An error in temperature of 3 K could cause an error in PI of more than 3 m s^{-1} if it extends to 100 hPa level. However, *Parker et al.* [1997] note that the error is much greater at 50 hPa than at 100 hPa.

[56] Another problem with smaller averaging areas are the relatively large local changes in SST in some regions. Area-averaged values and absolute values of the SST change from 1969 to 1990 were calculated for five ocean regions. Since these changes were somewhat larger for the North Atlantic and the Northeast Pacific than for the other ocean regions, different year pairs were chosen for these two ocean regions. For the North Atlantic and Northeast Pacific we use year pairs 72–86 and 67–81, respectively. The ocean regions and the SST changes between the former and latter year of corresponding year pairs are given in Table 3.

[57] The PIs for the five ocean regions are shown in Figure 10. The latitude and longitude ranges for the North Indian Ocean, Northwest Pacific, North Atlantic, and Northeast Pacific are the same as given in Table 3, but for the Southern hemisphere the whole hemisphere is included in the calculation. The same cutoff value of 40 m s^{-1} has been used as for the global PI. We stress that results for the presatellite and satellite era should be considered separately as in case of global PI.

[58] Figure 10 shows that the qualitative behavior of PI in the Southern Hemisphere and North Atlantic is rather similar to the behavior of the global PI. It is interesting that even in the 1960s when there were few observations from the Southern Hemisphere, the correspondence of the global PI and that of the Southern Hemisphere is rather good. Perhaps a sparser observation network is sufficient in the Tropics than in the middle latitudes due to a larger Rossby radius of deformation in the Tropics. In the Northeast Pacific the interannual variation of PI is much larger than that of the global PI and the signal of the El Niño-Southern Oscillation is clearly noticeable, with larger PIs occurring during El Niño years. The mean PI for the North Indian Ocean and Northwest Pacific is more than 10 m s^{-1} higher than for the whole globe.

Table 3. Ocean Regions and |SST| Changes Between Earlier and Later Years

Region	Year Pair	Lattitude Range	Longitude Range	|SST|, °C
N. Indian Ocean	69–90	0–30N	40–110E	0.11
N.W. Pacific	69–90	0–30N	100–180E	0.19
S. Hemisphere	69–90	0–30S	all	0.26
N. Atlantic	72–86	0–40N	100–20W	0.31
N.E. Pacific	67–81	0–30N	150–100W	0.16

[59] The maximum PI in the presatellite era occurs either in 1962 or 1963 in all ocean regions except for the Southern Hemisphere and Northeast Pacific for which it occurs in 1972. The maximum PI during the satellite era occurs in 1995–1996 except in the Northeast Pacific and North Indian Ocean where it occurs in 1987.

6. Conclusions

[60] As recent studies suggest that any climatic change in PI would affect the intensity distribution of real tropical cyclones uniformly, PI can be considered as a proxy for real tropical cyclone intensities. We have calculated the global PI for years 1958–1996 using the NCEP Reanalysis data and the NCEP EOF SST analysis to study the interannual to interdecadal variation of PI. Owing to a large spurious temperature increase in 1979, we adjusted the Reanalysis temperatures by finding years with similar mean tropical SST before and after year 1979. At each grid point, the annual mean temperature change from the earlier to the later year was subtracted from Reanalysis temperatures in 1980–1996 and PIs were recalculated using the adjusted temperatures.

[61] A factor contributing most to the error in the calculated PI is the method of adjusting for the temperature jump in 1979. If all six year pairs are accounted for, the largest PIs, obtained with year pair 72–80, are 1.8 m s^{-1} larger than the smallest PIs, obtained with year pair 69–88 (Figure 5). Therefore, time series for the presatellite and satellite era should be treated as separate. Still, for each of the five different year pairs, excluding only year pair 1969–1988, the maximum global PI, averaged over regions where it exceeds 40 m s^{-1}, occurred in the mid-1990s. The year-to-year variations in PI can be as large as 1 m s^{-1}. All year pairs showed a large increase of PI occurring in the 1990s. For 69–90, PI increased from 71.4 in 1990 to 73.4 m s^{-1} in only four years. The corresponding increase in the Tropics (Figure 9) was even larger.

[62] Sensitivity experiments showed that an error in SST of 0.1°C can lead to an error in PI of 0.7 m s^{-1} and an error

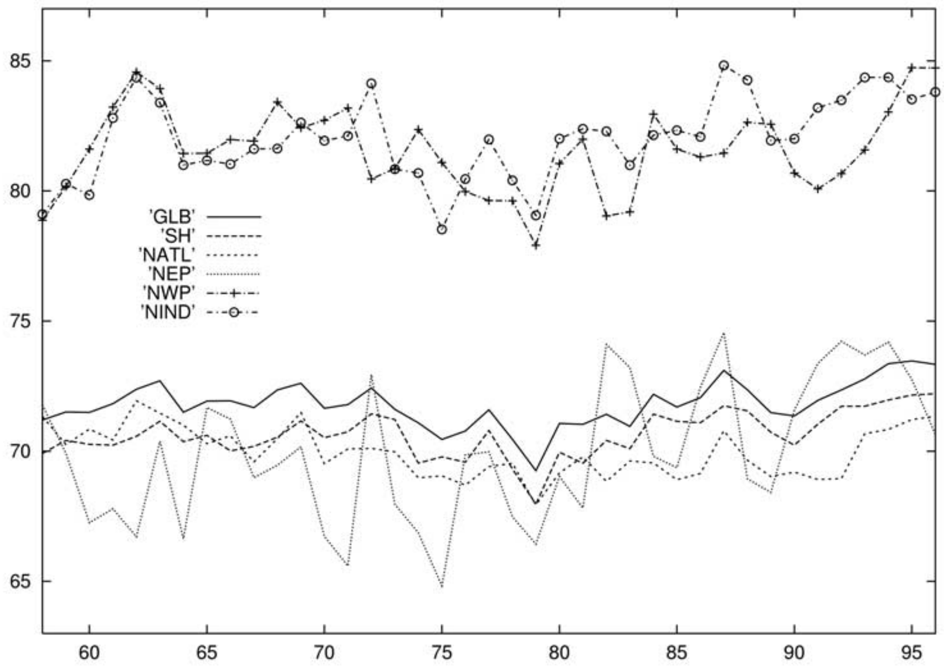

Figure 10. Annual average of daily PI values that exceed 40 m s^{-1} in the following ocean regions: global (GLB), Southern Hemisphere (SH), North Atlantic (NATL), Northeast Pacific (NEP), Northwest Pacific (NWP), and North Indian Ocean (NIND). Year pairs used in adjusting the Reanalysis temperatures are given in Table 2. Values for 1977–1979 should be disregarded as in Figure 8.

in tropospheric temperature of 0.25°C extending to 100 hPa would lead to an error in PI of 0.3 m s^{-1}. For the satellite era, depending on the method of comparison, the RMS difference of seasonal mean temperature anomalies between the HadTR1.1 sounding data set and the Reanalysis is 0.13°C or 0.26°C for the equivalent Channel 4 temperature [*Santer et al.*, 1999]. *Hurrell and Trenberth* [1999] have shown that in the Tropics, differences in the monthly anomalies among three SST data sets are within about 0.1°C for after 1981. In the presatellite era, the differences are somewhat larger and the lack of correspondence between PI and SST_{PI} is, indeed, often associated with known differences in the SST data set used by the Reanalysis and that used in the calculation of PI.

[63] When it comes to the reliability of the interdecadal variability of the calculated PI (apart from what happens in 1979), *Hurrell and Trenberth* [1999] have shown that the NCEP SST analyses may be getting spuriously cooler in the satellite era, especially in the 1990s. The spurious cooling is of the magnitude of 0.1°C and could lead to calculated values of PI that are too small. The effect of the possible spurious cooling of SST on PIs is, however, uncertain, since if the Reanalysis temperatures are affected by the spurious cooling of SSTs, then the effect on PIs would be smaller than suggested by our sensitivity experiment. There has also been spurious cooling of Reanalysis temperatures in the early 1990s until about 1995, as shown by *Basist and Chelliah* [1997]. If the spurious cooling extends up to 100 hPa, which is uncertain, it may be associated with values of PI that are too large. However, this cooling is rather small and causes an error in PI of at most 0.3 m s^{-1} or so in the early 1990s. It is also possible that the spurious cooling is related to the anomalously warm tropical tropopause in the Reanalysis as compared to radiosonde temperatures in 1989–1993 [*Randel et al.*, 2000]. A too warm upper troposphere in the Reanalysis could at least partially explain the relatively low values of PI when compared to SST_{PI} in 1990–1991.

[64] Suggested spurious drying of radiosonde humidities in late 1980s and early 1990s could cause an increase in values of PI as compared to SST_{PI}. However, PI is most sensitive to 1000 hPa relative humidity and at this level ship reports are more important than radiosonde measurements as input data to the Reanalysis in terms of number of observations (W. Ebisuzaki, personal communication, 2002). Therefore, it is likely that the Reanalysis is not affected as much by the spurious drying as the radiosonde data are. Moreover, for the regions of tropical oceans where there are no observations, the model's impact is very important (W. Ebisuzaki, personal communication, 2002).

[65] *Bengtsson et al.* [1999] have used The European Center/Hamburg ECHAM4 model to show cooling of the troposphere associated with the ozone decrease. They obtained 0.2–0.4°C cooling per decade between 100 and 200 hPa in the Tropics. Such a cooling might be a factor in the increase of PI as compared to SST_{PI} from the early 1980s to the mid-1990s. Decreasing the atmospheric temperatures while keeping the SSTs the same would increase PI and decrease the SST_{PI}, as shown by our sensitivity experiments. It is also possible that any ozone-related cooling may result in adjustments of Reanalysis temperatures using the year pair 1969–1990 that are too small, and therefore, PIs in the satellite era that are likewise too small. On the other hand, a possible spurious warm anomaly in the

Reanalysis temperatures [*Randel et al.*, 2000] in 1990 would have an opposite effect for the PIs obtained using year pair 1969–1990.

[66] PIs were also calculated for the Southern Hemisphere, North Atlantic, Northeast Pacific, North Indian Ocean, and the Northwest Pacific. The behavior in the Southern Hemisphere was rather similar to that for the whole globe. The mean PI for the North Indian Ocean and the Northwest Pacific was about 10 m s^{-1} larger than for the whole globe. For all of the ocean regions studied, the maximum PI during the presatellite era occurred in 1962–1963 or 1977. During the satellite era, the maximum PI occurred in 1995–1996 except in the Northeast Pacific and North Indian Ocean where it occurred in 1987.

[67] Estimates of the decadal variations of PI with smaller margins of error must await the direct assimilation of radiances from satellites in the next generation NCEP/NCAR Reanalysis, which will hopefully lead to a more consistent temperature analysis. Also, further improvements in the SST data are likely to yield significant improvements in the PI calculation. Removal of biases in the radiosonde data and using the corrected data in future reanalyses has a potential for greatly improving the usefulness of reanalyses for PI calculations as PI is sensitive to upper tropospheric temperatures and boundary layer humidities.

[68] Finally, we want to stress that some known errors in the Reanalysis, discussed in this paper, have a large effect on the global, and probably even larger effect on the regional PI trends both in the presatellite and satellite eras. Moreover, their net effect is difficult to estimate as the magnitude, the vertical extent, and even the time of appearance and disappearance of many errors are unknown. There may also be many errors that are still unknown and that significantly affect the calculated PI. Calculating PI with future reanalyses will hopefully provide us with more information about currently known and unknown errors. A good sign in the results presented in this paper is that the behavior of global PI and that averaged over the Southern hemisphere are rather similar (Figure 10). Comparison of our results with local PI behavior calculated using data from selected radiosonde stations would provide us with additional information. Work is underway to do such a comparison (M. Free, ARL NOAA, personal communication, 2002). Preliminary results show that different stations show very different long-term behavior. Some stations show increasing, some decreasing, and some practically no trends of PI over the same time period we have studied in this paper. A thorough comparison will hopefully lead to better understanding of whether the station-to-station variation of trends just indicates local variability or large magnitude errors in results presented in this paper. In that respect, it is interesting that Figure 10 show rough similarities of the long-term variability in many of the ocean basins.

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