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STRATOSPHERIC COOLING AND TROPICAL CYCLONES

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1. INTRODUCTION

Over the past 25 years or so, there has been a strong increase in almost every metric of tropical cyclone activity in the North Atlantic, and in some metrics of intense cyclones elsewhere around the globe (Elsner et al., 2008). These have exceeded expectations based on simple theory applied to observed increases in sea surface temperature over the period, leading to questions about whether the observed increase in activity is related to the observed warming of the ocean. In this study, we apply a recently developed tropical cyclone downscaling technique to several atmospheric general circulation models (AGCMs) driven only by observed sea surface temperatures over the last 100 years or so. We show that, for the tropical Atlantic, such downscalings capture the observed variability of tropical cyclones up to about 1970, but largely fail to capture the recent large upswing in activity. We show that this is directly attributable to the failure of the ACGMs to capture the large increase in potential intensity present in reanalyses, in spite of the fact the observed increase in sea surface temperature is prescribed in the models. Finally, we show that the large increase in potential intensity over the North Atlantic over the past quarter century is attributable both to the increase in sea surface temperature and to the decline in temperatures near the tropical tropopause, which together lead to a precipitous decline in hurricane outflow temperatures. The failure of AGCMs to capture the near-tropopause cooling is may be owing to their omission of ozone variability and to their relatively poor vertical resolution near the tropopause.

2. BRIEF REVIEW

Potential intensity (PI) theory (Bister and Emanuel, 1998) shows that tropical cyclone intensity is bounded above by a quantity that depends on the degree of thermodynamic disequilibrium between the atmosphere and ocean and the difference between the sea surface temperature (SST) and the entropy-weighted mean temperature of the outflow layer. In spite of this modestly complex dependence on the thermodynamic characteristics of the environment, PI is often discussed as though it depends only on sea surface temperature.

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The first calculations of the response of PI to changing greenhouse gases in the context of a single column model run into a state of radiative-convective equilibrium (Emanuel, 1987) showed modest increases of PI tabulated as a function of sea surface temperature, averaging around $4\,ms^{-1}K^{-1}$. Given the expected increase in tropical sea surface temperatures of around 2-3 C for a doubling of CO_2, this would yield increases of PI of around 10-15%, consistent with the findings of the most recent review of global warming effects on tropical cyclones (Knutson et al., 2010).

At the same time, comparison of observed tropical cyclone intensity in the North Atlantic region with sea surface temperatures suggests a more sensitive dependence on SST than elementary PI theory appears to predict. For example, the author (Emanuel, 2005) showed a high correlation between low-pass-filtered tropical cyclone power dissipation index (PDI) and storm season low-pass-filtered SST in the North Atlantic for the years 1949-2003. (PDI is defined as the integral over the life of each storm of its peak wind speed cubed.) Figure 1 compares low-pass-filtered[1] Atlantic PDI to Atlantic main development region (MDR) low-pass-filtered August-October SST using data from the period 1949-2009.

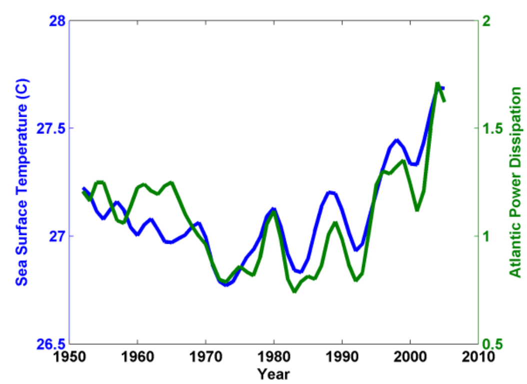

Figure 1: Low-pass-filtered August-October MDR SST (blue) and annual Atlantic PDI (green) based on data from the period 1949-2009.

It is notable that PDI has nearly tripled between about 1980 and 2009, despite an increase of SST of only about 0.8 C. As noted by Emanuel

[1] The filtered employed here is weighted 1-3-4-3-1.

(2005), cubing the velocity increases the expected dependence of PDI on SST to around 20% per degree C, while the correlation between storm duration (which contributes to PDI) and peak winds might yield another 7% K^{-1}, giving a net predicted increase of around 27% K^{-1}, or about 22% for the observed 0.8 C increase in SST, still far short of the observed change. At the same time, the actual number of Atlantic tropical cyclones has increased over the period in question, accounting for much if not all of the remaining increase in power dissipation. In fact, the actual intensity of Atlantic tropical cyclones has increased by about 20% (Emanuel, 2007), a great deal more than predicted by standard PI theory. The object of the present study is to understand this difference.

3. RELATIVE SST

Swanson (2008) argued that hurricane activity is much more sensitive to the SST relative to the tropical mean SST than it is to the absolute value of the SST. This proposition is well supported by potential intensity theory, which holds that the square of the maximum sustainable wind speed is given by (e.g. Bister and Emanuel, 1998)

$$V_{max}^2 = \frac{C_k}{C_D}\frac{T_s - T_o}{T_o}\left(k_0^* - k_b\right), \qquad (1)$$

where C_k and C_D are nondimensional exchange coefficients for enthalpy and momentum, respectively, T_s and T_o are the surface temperature and entropy-weighted mean outflow temperature, respectively, and k_0^* and k_b are, respectively, the saturation enthalpy of the sea surface (a function of SST and surface pressure) and actual enthalpy of the boundary layer.

By adding the potential energy gz to the enthalpies in (1), we can express them instead as moist static energies, h_0^* and h_b. Moreover, regions that support tropical cyclones are nearly neutral to convection, so that the boundary layer moist static energy, h_b, is nearly equal to the saturation moist static energy of the troposphere, h^*. Thus we may re-write (1) as

$$V_{max}^2 = \frac{C_k}{C_D}\frac{T_s - T_o}{T_o}\left(h_0^* - h^*\right). \qquad (2)$$

Note that the tropospheric saturation moist static energy, h^*, is nearly constant with altitude: this is just the condition that the temperature lapse rate in the Tropics is nearly moist adiabatic. Owing to the inability of the weakly rotating tropical atmosphere to support large temperature gradients on time scales greater than those characterizing internal waves (Sobel and Bretherton, 2000), h^* is also nearly horizontally uniform within the Tropics. Thus h^* is a nearly universal function of time in the tropical troposphere.

Therefore, according to (2), potential intensity is proportional to the difference between the local SST, as reflected by the value of h_0^*, and the nearly spatially and vertically uniform value of the saturation moist static energy of the Tropics, h^*. For the relative SST concept to be consistent with theory, one would have to advance the additional postulate that h^* is proportional to the tropical mean SST. At first blush, it is not obvious that this should be so. For example, if one were to increase the mean surface wind speed in the Tropics, SST should decrease and h^* should increase, so under this exercise, the two quantities would be negatively correlated. Also, using the tropical mean SST ignores the possibly large influence of land on tropical mean atmospheric temperature. Changes in land surface albedo, for example, would be expected to influence the tropical mean atmospheric temperature. Nevertheless, empirically, there is a close correlation between tropical mean SST and h^*, as shown in Figure 2, though note that the saturation entropy of the sea surface appears to be increasing faster than that of the middle troposphere.

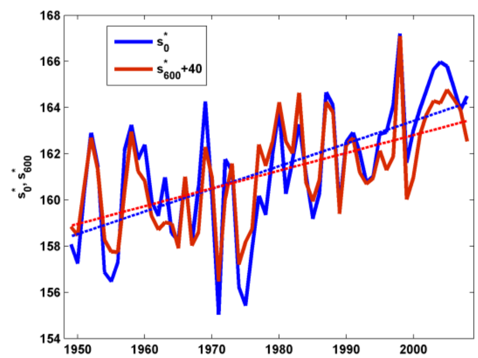

Figure 2: Annual average saturation entropy of the sea surface (blue) and annual average saturation entropy at 600 hPa (green), from 1949 to 2008. Both quantities have been averaged over all ocean grid points between 30° N and 30° S using NCAR/NCEP reanalysis data. Units are $J\,kg^{-1}K^{-1}$, and 40 of these units have been added to the 600 hPa saturation entropy for ease of comparison.

Thus equating relative SST to $h_0^* - h^*$ as it appears in (2) can capture short-term variations quite well, but will miss long-term trends.
It is worth noting that local variability of SST on time scales longer than those associated with thermal equilibration of the ocean mixed layer (around 2 years) is strongly constrained by the surface energy balance, which may be written

$$C_k \rho |\mathbf{V}_s| \left(k_0^* - k_b\right) = F_{rad} - d\nabla \bullet \mathbf{F}_{\mathbf{ocean}}, \qquad (3)$$

where ρ is the air density, $|\mathbf{V}_s|$ is the local mean surface wind speed, F_{rad} is the local net downward surface radiative flux, d is the depth of the ocean mixed layer, and F_{ocean} is the net heat flux in the ocean mixed layer. Replacing the enthalpies by moist static energies, and using the assumption of convective neutrality, we can re-write (3) as

$$h_0^* = h^* + \frac{F_{rad} - d\nabla \bullet \mathbf{F}_{\mathbf{ocean}}}{C_k \rho |\mathbf{V}_s|}. \qquad (4)$$

This shows that on time scales longer than about two years, the SST, as represented here by h_0^*, responds to changing atmospheric temperature (h^*), surface radiation, ocean het flux, and surface wind speed.

If we substitute (3) into (1) we get

$$V_{max}^2 = \frac{T_s - T_o}{T_o} \frac{F_{rad} - d\nabla \bullet \mathbf{F}_{\mathbf{ocean}}}{C_D \rho |\mathbf{V}_s|}. \qquad (5)$$

Thus on time scales longer than about two years, potential intensity depends on surface temperature, outflow temperature, surface radiation, ocean heat flux, and surface wind speed. The idea that SST can vary independently of atmospheric temperature, inherent in some interpretations of relative SST, can only work on relatively short times scales (<~ 2 years). This is perhaps why relative SST is no better correlated with hurricane power than is absolute SST when both temperatures are smoothed over multiple years. In any case, by its definition, relative SST cannot explain global increases in potential intensity. Figure 3 shows that potential intensity has been increasing during the past quarter century in all of the tropical ocean basins, at least according to NCAR/NCEP reanalysis data. Moreover, it has been increasing at a rate far in excess of the $4\, ms^{-1}K^{-1}$ prediction from single-column models. But, looking at (5), it is clear that changing radiative forcing is only one way of changing potential intensity. The other major influences are

- Changing mean surface wind speed
- Changing outflow temperature

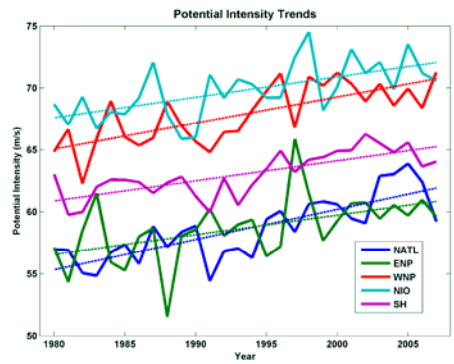

Figure 3: Annual values of potential intensity, from 1980 to 2007, in five regions of the tropical oceans: North Atlantic (blue), eastern North Pacific (green), western North Pacific (red), north Indian (aqua), and all of the Southern Hemisphere (violet). Linear trends shown by dashed lines. From NCAR/NCEP reanalysis data.

Examination of the monthly mean surface wind speed in NCAR/NCEP reanalysis shows modest decreases in the eastern North Pacific and northern Indian Ocean regions, but no trend or increases elsewhere. Thus in the North Atlantic region, for example, surface wind trends cannot explain the upward trend in potential intensity. Thus we focus on trends in outflow temperature in the following discussion.

4. OUTFLOW TEMPERATURE AND STRATOSPHERIC COOLING

Figure 4 shows the annual values and trends of outflow temperature derived from NCAR/NCEP reanalysis data for the period 1980-2008.

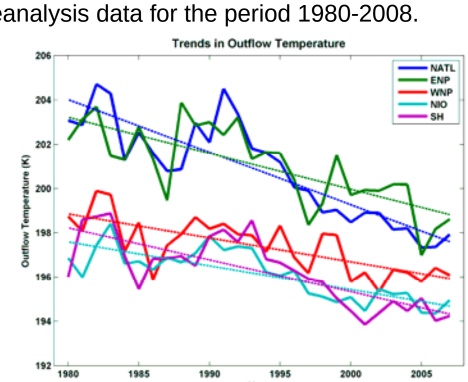

Figure 4: As in Figure 3, but showing the outflow temperature, T_o.

The outflow temperature is estimated by first saturating an air parcel at sea surface temperature and at the pressure under the eyewall as calculated by the potential intensity algorithm (Bister and Emanuel, 2002). This parcel is then lifted under a reversible moist adiabatic process until its density temperature matches the density temperature of its environment. Thus the outflow temperature can change owing both to changing SST and to changing temperature near the level of neutral buoyancy for the ascent of a parcel saturated at SST under the eyewall; this is generally in the lower stratosphere. (Remember that the outflow level thus defined is for a hypothetical storm that has reached its potential intensity.) Generally, increasing SST will yield decreasing outflow temperature, since parcels can ascend to higher levels. It should also be noted that the effects of increasing SST and decreasing lower stratospheric temperature are not, in general, linearly additive.

The decreases in outflow temperature shown in Figure 4 are impressive. Over the 29 years from 1980 to 2008, the outflow temperature over the tropical North Atlantic decreased by about 6 C, contributing to about a 5% increase in potential intensity. This is about half the observed increase of potential intensity shown in Figure 3; the remainder comes from increasing values of $h_0^* - h^*$. Thus the cooling of the stratosphere is as important as increasing SST (or relative SST) in explaining the nearly 10% increase of potential intensity in the tropical North Atlantic from 1980 to 2008.

But is it real? There is considerable controversy about the quality of various estimates of temperature in the upper troposphere and lower stratosphere (Free and Seidel, 2007). The two principal methods of estimating temperature, radiosondes thermistors and passive microwave retrievals from space, each suffer from various problems.

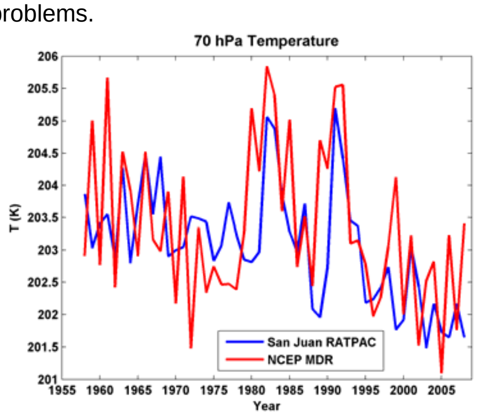

Figure 5: August-October mean temperature at 70 hPa, from 1957-2008, for San Juan (blue) and averaged over the tropical Atlantic MDR (6-18 N, 20-60W; red).

Figure 5 compares a time series of August-October mean temperature at 70 hPa measured by radiosondes at San Juan, Puerto Rico, with the temperature averaged over the same months and over the area defined by 6-18N and 20-60W averaged over the same months. The San Juan data have been corrected for inhomgeneities by Free et al. (2005). The radius of deformation is very large in the tropical lower stratosphere, owing to the combination of large stratification and small Coriolis parameter, so one might expect a good correlation between the temperature at San Juan and the temperature over the Atlantic Main Development Region (MDR). Figure 5 shows quite good agreement. Bear in mind that the NCAR/NCEP reanalysis assimilates the San Juan sounding, so the two time series are not independently derived, but it also assimilates passive microwave satellite measurements after 1979, which do not always agree with the radiosonde data (Free and Seidel, 2007), so agreement is by no mean guaranteed. Note in particular the high temperature spikes associated with the volcanic eruptions of El Chicón in 1982 and Pinatubo in 1991.

Thus the cooling of the lower stratosphere over the past quarter century appears to be real, and as we have shown, it is responsible for a substantial fraction of the increase of potential intensity of the Tropics over this period. We defer a discussion of the possible causes of this cooling to the next section.

5. FAILURE OF GCMS TO SIMULATE LOWER STRATOSPHERIC COOLING

Here we show that contemporary general circulation models (GCMs) forced by specified sea surface temperatures, sea ice distributions, and atmospheric composition (including aerosols) largely fail to capture the cooling of the lower stratosphere over the past quarter century. For this purpose, we examine two AGCMs: The ECHAM5 model (Roeckner et al., 2006) as configured by a group at ETH Zurich (Wild Martin and Doris Folini, personal communication) and the NOAA/CIRES 20th Century Reanalysis, Version 2 (Compo et al., 2006).

The ECHAM 5 model was run at T42 resolution for the period 1870-2005, forced by prescribed, time-varying Hadley Center SSTs and sea ice (HADSST2). The model also has time-varying anthropogenic and natural aerosols, but it is important to note that the concentration of atmospheric greenhouse gases remains constant in these simulations.

Version 2 of the NOAA/CIRES 20th century analysis assimilates Hadley Center SSTs and sea ice (HADSST1.1) into a version of the NCEP operational Climate System Forecast Model (Saha and coauthors, 2006), run at T63 resolution with 28 levels in the vertical. The model also assimilates surface pressure observations, which allows it to simulate actual day-to-day variations in the state of the atmosphere. In addition, the model includes time-varying CO_2, solar output, and volcanic aerosols, and ozone varies according to the internal dynamics and chemistry of the model. The simulation extends from 1891 through 2008.

Figure 6 compares the MDR August-October mean outflow temperatures produced by the ECHAM5 and NOAA/CIRES simulations to each other and to the standard NCAR/NCEP reanalysis-derived outflow temperature.

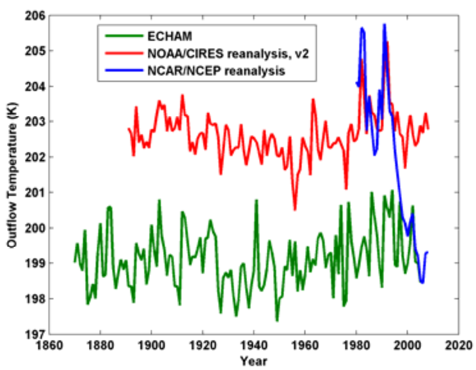

Figure 6: August-October outflow temperatures averaged over the Atlantic MDR from the ECHAM 5 simulation (green), the NOAA/CIRES 20th Century reanalysis, version 2 (red) and the NCAR/NCEP reanalysis (blue).

The NCAR/NCEP reanalysis assimilates atmospheric data with serious inhomogeneities, particularly around 1979, when assimilation of satellite-derived radiances was begun; thus we use only the post-1980 reanalysis for comparison. It can be seen that the NOAA/CIRES reanalysis captures some of the spikes in outflow temperature related to the eruptions of El Chicón in 1982 and Pinatubo in 1991, but fails to simulate the sharp decline in outflow temperature that began after Pinatubo. The ECHAM5 simulation shows fairly flat outflow temperature through the period, but with some noticeable random variability.

The important conclusion here is that *neither simulation captures the sharp decline in outflow temperature after 1991*. This has strong consequences for the models' simulation of potential intensity, and this in turn strongly affects the models' downscaled tropical cyclone activity.

The August-October MDR mean potential intensities from ECHAM5 and the two reanalyses are shown in Figure 7. The large post-1991 increase in potential intensity evident in the NCAR/NCEP reanalysis is largely absent from ECHAM 5 and the NOAA/CIRES 20th century reanalysis. Although the NCAR/NCEP reanalysis uses the Reynolds analyses of SST, comparison with the two Hadley Center SST products shows very slight differences over the post-1980 August-October MDR. We believe that the large difference in post-1991 trend is largely owing to the failure of the NOAA/CIRES and ECHAM5 models to capture the cooling of the lower stratosphere, which is the main cause of the post-1991 decline in outflow temperatures evident in Figure 6.

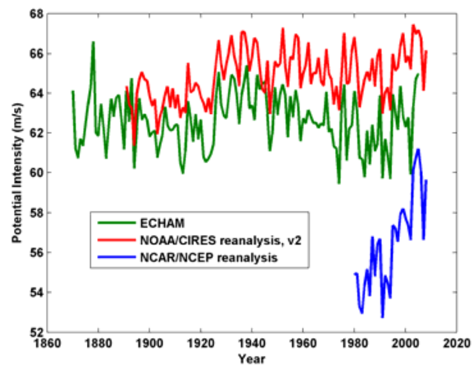

Figure 7: As in Figure 6, but for potential intensity.

The lack of cooling in the models is consistent with other studies (Cordero and Forster, 2006) that shows a general failure of GCMs to capture the recent cooling of the lower stratosphere.

6. PHYSICAL CAUSES OF LOWER STRATOSPHERIC COOLING

The physical causes of the recent decline in lower stratospheric temperature are disputed. Increasing CO_2 content does lead to a cooling of the stratosphere, but this is most pronounced in the middle and upper stratosphere with little effect at and below about 70 hPa. Depletion of ozone is a more plausible explanation for at least some of the cooling. Examining ten GCMs run in support of the most recent IPCC report, Cordero and Forster (2006) found that those models that include variable ozone have trends in lower stratospheric temperature that are closer to observations than those that use fixed ozone. Thompson and Solomon (2005) also find that tropical stratospheric temperatures have been declining and note that decreasing ozone concentrations cannot explain all of the observed cooling. They speculate that the remaining cooling might be explained by increasing water vapor concentrations and/or increased upwelling in the lower tropical stratosphere. Fu et al. (2006), examining MSU observations, show a strong negative correlation between spatial trends in upper tropospheric and lower stratospheric temperatures, suggesting that the dynamical response of the lower stratosphere to changing temperature distributions in the troposphere (Holloway and Neelin, 2007) may make an important contribution to the lower stratospheric cooling, through the direct effect of upwelling on equilibrium temperature, and possibly also by the dilution of lower stratospheric ozone by injection of ozone-poor tropospheric air.

7. ANAMALOUS STRATOSPHERIC COOLING AND TROPICAL CYCLONES

Recently, the author and colleagues developed a method of downscaling tropical cyclone activity from the coarse-grained output of reanalyses and GCMs (Emanuel et al., 2008). The technique begins by randomly seeding ocean basins with proto-tropical cyclones, and then using a coupled atmospheric-ocean tropical cyclone model to determine the subsequent development of the initial disturbances, whose movement is determined using a beta-and-advection model applied to the GCM or reanalysis winds. Many of the initial disturbances die; the survivors are regarded as constituting the tropical cyclone climatology of the model state in question. When applied to the NCAR/NCEP reanalyses for the period 1980-2006, the method explains about 65% of the observed interannual variability in the frequency of North Atlantic tropical cyclones, and also captures the large upward trend over that period.

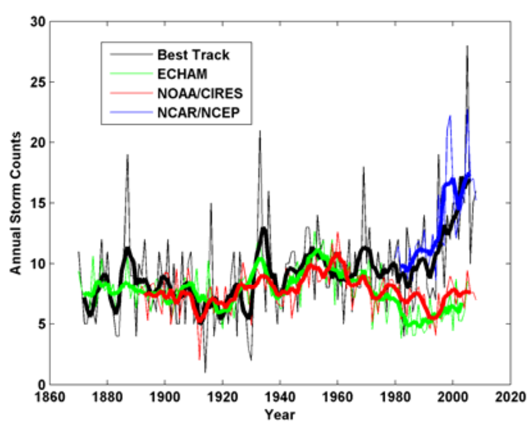

Figure 8: Annual Atlantic tropical cyclone counts: Unadjusted best-track data (black); and downscaled from the NCAR/NCEP reanalysis, 1980-2008 (blue), the ECHAM 5 simulation, 1870-2005 (green), and the NOAA/CIRES reanalysis, 1891-2008 (red). Thin lines show annual values, thick lines show 5-year running means.

Here we apply that method to the ECHAM 5 simulation and the NOAA/CIRES reanalysis described above, and compare the results to those derived by downscaling the NCAR/NCEP reanalyses and to historical tropical cyclone data in the North Atlantic region.

Figure 8 shows the hindcast and observed Atlantic tropical cyclone counts over the years covered by the respective simulations. The only hindcast that captures the large upward swing in tropical cyclone counts since 1990 is that derived from the NCAR/NCEP reanalysis, which is also the only product that captures the cooling of the lower stratosphere during that period. The downscaled tropical cyclone counts from the ECHAM 5 simulation and from the NOAA/CIRES reanalysis agree well with each other over the whole period, and with the historical data up through about 1965; thereafter, the two simulations dramatically underpredict the upward trend in storm counts. Since all of the simulations, including the NCAR/NCEP reanalysis, use SSTs that are very similar after 1980, we conclude that the failure of the downscaling derived from the ECHAM 5 simulation and the NOAA/CIRES reanalysis is owing mostly to the failure of the those simulations to capture the cooling of the lower stratosphere after 1980. It follows that the observed increase in Atlantic tropical cyclone activity in recent decades may be mostly attributable to the cooling of the lower stratosphere during that time.

8. SUMMARY

Observations suggest that the lower tropical stratosphere has been cooling over the last few decades. While the NCAR/NCEP reanalysis captures this cooling over the Atlantic during hurricane season, GCMs largely fail to simulate it. The cooling, especially when coupled to increasing SSTs, results in a large reduction of outflow temperature, an important component of potential intensity. The potential intensity, in turn, not only governs the intensity of tropical cyclones, but is an important component in setting the frequency of storms, as suggested both by downscaling studies and contemporary genesis indices. Atlantic tropical cyclone activity downscaled from NCAR/NCEP reanalyses after 1980 show remarkably good agreement with observed tropical cyclone activity during that period, but when the same technique is applied to two AGCMs forced by observed SSTs and sea ice, the substantial increase in Atlantic tropical cyclone activity after 1990 is almost completely absent, even though the simulations agree well with the historical data before about 1965. Since all the simulations use very similar SSTs, this suggests that the upswing in Atlantic activity since about 1990 is largely owing to the cooling of the stratosphere, which the GCMs fail to simulate.

Whatever the cause of the observed stratospheric cooling, the fact that climate models do not simulate it and that it is apparently an important influence on tropical cyclones together warrant lower confidence in recent projections of the response of tropical cyclones to global warming.

Acknowledgements

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5. REFERENCES

Bister, M., and K. A. Emanuel, 1998: Dissipative heating and hurricane intensity. *Meteor. Atmos. Physics*, **50,** 233-240.

——, 2002: Low frequency variability of tropical cyclone potential intensity, 1: Interannual to interdecadel variability. *J. Geophys. Res.*, **107,** doi:10.1029/2001JD000776.

Compo, G. P., J. S. Whitaker, and P. D. Sardeshmukh, 2006: Feasibility of a 100 year reanalysis using only surface pressure data. *Bull. Amer. Meteor. Soc.*, **87,** 175-190.

Cordero, E. C., and P. M. Forster, 2006: Stratospheric variability and trends in models used for the IPCC AR4. *Atmos. Chem. Phys.*, **6,** 5369-5380.

Elsner, J. B., J. P. Kossin, and T. H. Jagger, 2008: The increasing intensity of the strongest tropical cyclones. *Nature* **455,** 92-95.

Emanuel, K., 2007: Environmental factors affecting tropical cyclone power dissipation. *J. Climate*, **20,** 5497-5509.

Emanuel, K., R. Sundararajan, and J. Williams, 2008: Hurricanes and global warming: Results from downscaling IPCC AR4 simulations. *Bull. Amer. Meteor. Soc.*, **89,** 347-367.

Emanuel, K. A., 1987: The dependence of hurricane intensity on climate. *Nature*, **326,** 483-485.

——, 2005: Increasing destructiveness of tropical cyclones over the past 30 years. *Nature*, **436,** 686-688.

Free, M., and D. J. Seidel, 2007: Comments on “Biases in stratospheric and tropospheric temperature trends derived from historical radiosonde data”. *J. Climate*, **20,** 3074-3709.

Free, M., D. J. Seidel, J. K. Angell, J. Lanzante, I. Durre, and T. C. Peterson, 2005: Radiosonde Atmospheric Temperature Products for Assessing Climate (RATPAC): A new dataset of large-area anomaly time series. *J. Geophys. Res.*, **110,** doi:10.1029/2005JD006169.

Fu, Q., C. M. Johanson, J. M. Wallace, and T. Reichler, 2006: Enhanced mid-latitude tropospheric warming in satellite measurements. *Science*, **312,** 1179.

Holloway, C. E., and D. J. Neelin, 2007: The convective cold top and quasi equilibrium. *J. Atmos. Sci.*, **64,** 1467-1487.

Knutson, T. R., and Coauthors, 2010: Tropical cyclones and climate change. *Nature Geosci.*, **3,** 157-163.

Roeckner, E., R. , and Coauthors, 2006: Sensitivity of simulated climate to horizontal and vertical resolution in the ECHAM5 atmosphere model. *J. Climate*, **19,** 3771-3791.

Saha, S., and coauthors, 2006: The NCEP Climate Forecast System. *J. Climate*, **19,** 3488-3517.

Sobel, A. H., and C. S. Bretherton, 2000: Modeling tropical precipitation in a single column. *J. of Climate*, **13,** 4378-4392.

Swanson, K., 2008: Nonlocality of Atlantic tropical cyclone intensities. *Geochem. Geophys. Geosys.***,** doi:10.1029/2007GC001844.

Thompson, D., and S. Solomon, 2005: Recent stratospheric climate trends as evidenced in radiosonde data: Global structure and tropospheric linkages. *J. Climate*, **18,** 4785-4795.