THE HURRICANE–CLIMATE CONNECTION

BY KERRY EMANUEL

The Bernhard Haurwitz Memorial lecture delivered at the 2007 AMS Annual Meeting in San Antonio reviews progress in identifying and predicting the effects of climate change on tropical cyclone activity as well as possible feedbacks of changing storminess on climate itself. A recording of this lecture is also available in the meetings archives on the AMS Web site.

It has been understood for some time (e.g., Palmén 1948) that tropical cyclones respond to climate change on a variety of time scales. Empirical studies (e.g., Gray 1968) have established that tropical cyclone activity is sensitive to a variety of environmental conditions, including the magnitude of the shear of the horizontal wind through the depth of the troposphere, sea surface temperature (SST), low-level vorticity, and the humidity of the lower and middle troposphere. Theory has so far established only a bound on the intensity of tropical cyclones (Emanuel 1987), though empirically, this bound has been shown to provide the relevant scaling for the intensity of real storms (Emanuel 2000). This bound, referred to as the "potential intensity," has the units of velocity and is a function of the sea surface temperature and the profile of temperature through the troposphere and lower stratosphere (Bister and Emanuel 2002); it is a far more physically based quantity than SST.

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The abstract for this article can be found in the May issue, following the table of contents.
DOI:10.1175/BAMS-89-5-Emanuel

In final form 2 August 2007

While there has been some advance in the theory of tropical cyclone intensity, the question of frequency is more vexing. About 90 tropical cyclones develop each year around the globe, with a standard deviation of 10; at present, we lack a theory that predicts even the order of magnitude of this number. Although there has been little progress in developing a theory governing the rates of occurrence of tropical cyclones, a number of empirical indices have been developed, beginning with that of Gray (1979). Recently, Emanuel and Nolan (2004) incorporated potential intensity in an empirical index of the frequency of tropical cyclone genesis, called the genesis potential index (GPI):

$$\mathrm{GPI} \equiv \left|10^{5}\eta\right|^{3/2} \left(\frac{\mathcal{H}}{50}\right)^{3} \left(1+0.1V_{\mathrm{shear}}\right)^{-2}, \qquad (1)$$

where η is the absolute vorticity in s^{-1}, $\mathcal{H}$ is the relative humidity at 600 hPa in percent, V_{pot} is the potential intensity in $\mathrm{m\ s}^{-1}$, and V_{shear} is the magnitude of the vector shear from 850 to 250 hPa in $\mathrm{m\ s}^{-1}$.[1] This

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[1] The numerical factors in (1) are designed to yield values of the GPI of order unity, but the absolute magnitude of the GPI is regarded as arbitrary.

index was fitted to the annual cycle of genesis in each hemisphere, and to the spatial distributions of storms each month of the year, as described in some detail in Camargo et al. (2007), who also showed that the GPI captures some of the dependence of genesis rates on El Niño–Southern Oscillation (ENSO). The high power with which the potential intensity enters this empirical index suggests that it plays an important role in the frequency as well as intensity of tropical cyclones, but it must be stressed that a good theoretical understanding of the environmental control of storm frequency is lacking.

While theory is still deficient, there has been some progress in using climate models to simulate the effects of climate change on tropical cyclone activity, as reviewed in the "Attribution" section. At present, global models are too coarse to resolve the inner cores of intense tropical cyclones, and their ability to simulate the full intensity of such storms is therefore seriously compromised. Yet this approach is beginning to yield interesting and possibly useful insights into the effect of climate change on storm activity.

In this essay, I will review evidence from the instrumental record of changing tropical cyclone activity, including a discussion of various problems with the tropical cyclone data itself, and also briefly review the budding new field of paleotempestology. After describing the debate over attribution, I review the use of global models to deduce the effects of climate change on tropical cyclones, and present some results of a new method of deriving tropical cyclone climatology from global gridded data, such as contained in the output of global climate simulations. Finally, I argue that global tropical cyclone activity is responsible for some or perhaps most of the observed poleward heat transport by the oceans, thereby constituting an essential element of the global climate system. A summary is provided in the final section.

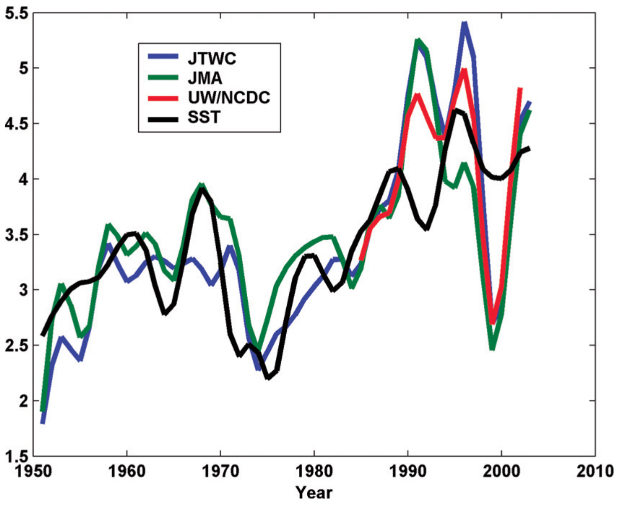

Fig. 1. Power dissipation (colored curves) in the western North Pacific according to (blue) data from the U.S. Navy Joint Typhoon Warning Center as adjusted by Emanuel 2005a, (green) unadjusted data from the Japanese Meteorological Agency, and (red) reanalyzed satellite data from Kossin et al. (2007). The black curve represents a scaled Jul–Oct SST in the tropical western North Pacific region. All quantities have been smoothed using a 1-3-4-3-1 filter.

TROPICAL CYCLONE VARIABILITY IN THE INSTRUMENTAL RECORD.

Beginning shortly after World War II (WWII), aircraft have surveyed tropical cyclones in the North Atlantic and western North Pacific, though aircraft reconnaissance in the latter basin ended in 1987. During the 1960s, earth-orbiting satellites began to image some tropical cyclones, and by about 1970 it can be safely assumed that hardly any events were missed. Before the aircraft reconnaissance era, tropical cyclone counts depended on observations from ships, islands, and coastal locations. Detection rates were reasonably high only in the North Atlantic, owing to dense shipping, but even here, the precise rate of detection remains controversial (Holland and Webster 2007; Landsea 2007). Estimates of the intensity of storms as measured, for example, by their maximum surface wind speeds are dubious prior to about 1958, and some would say prior to 1970 in the Atlantic and western North Pacific. Elsewhere, there are only very spotty estimates prior to the satellite era. Satellite-based estimates of intensity commenced in the 1970s and have improved along with the spatial resolution of satellite imagery, but the accuracy of such estimates is still debated. Intensity estimates based on aircraft measurements are prone to a variety of biases owing to changing instrumentation and means of inferring wind from central pressure, as described in the online supplement to Emanuel (2005a; available online at ftp://texmex.mit.edu/pub/emanuel/PAPERS/NATURE03906_suppl.pdf). Some indication of the nature of these problems is evident in Fig. 1, which shows a variety of estimates of tropical cyclone power dissipation in the western North Pacific since 1949. [The power dissipation is defined as the integral over the life of each storm of its maximum surface wind speed cubed, also accumulated over each year (see Emanuel 2005a).]

Note that the adjusted estimate from the Joint Typhoon Warning Center agrees well with the unadjusted estimate from the Japanese Meteorological Agency and that both are well correlated with sea surface temperature prior to the cessation of aircraft reconnaissance in 1987; after that time, there is much more divergence in the estimates and less correlation with SST. There is a general upward trend in SST and tropical cyclone power dissipation, but there are also prominent decadal fluctuations in both. The general upward trend in power dissipation was pointed out by the author (Emanuel 2005a) and is consistent with the finding by Webster et al. (2005) that the global incidence of intense tropical cyclones is generally trending upward.

In the North Atlantic, tropical cyclone records extend back to 1851 but are considered less reliable early in the period, and intensity estimates are increasingly dubious as one proceeds back in time from 1970. [A discussion of the sources or error may be found in the online supplement to Emanuel (2005a; see footnote 2) and in Emanuel (2007).] A vigorous debate has ensued over the quality of the wind data (Emanuel 2005b; Landsea 2005; Landsea et al. 2006), and even the annual frequency of storms is open to question prior to 1970 (Holland 2007; Holland and Webster 2007; Landsea 2007). Similar questions have been raised about the veracity and interpretation of the record of storms in the western North Pacific (Chan 2006).

Here, on the premise that storms were more likely to be detected near the time of their maximum intensity, we define a "storm maximum power dissipation" as the product of the storm lifetime maximum wind speed cubed and its duration, summed over all the storms in a given year. Figure 2 compares this quantity to the sea surface temperature of the tropical Atlantic in July through October, going back to 1870. Except for the period 1939–45, the correspondence between power dissipation and SST is remarkable, even early in the period. Since 1970, the r^2 between the two series is 0.86.

The very low power dissipation during WWII may reflect a dearth of observations owing to enforced radio silence on ships during the war. In the Atlantic, variations in the power dissipation reflect variations in numbers of storms to a large degree (Emanuel 2007). While some have argued that the number of Atlantic storms may have been grossly underestimated prior to the aircraft and/or satellite eras (Landsea 2007), statistical analyses of the likelihood of ships encountering storms suggest that the counts are good to one or two storms per year back to 1900 (Chang and Guo 2007; Vecchi and Knutson 2008), and it is also possible to overestimate storm counts owing to multiple counting of the same event encountered infrequently. In addition, Holland and Webster (2007) point out that the large increases during the 1930s and 1990s both occurred during periods when measurement techniques were relatively stable; the advent of aircraft reconnaissance in the 1940s and the introduction of satellites during the 1960s were not accompanied by obvious increases in reported activity. Even with fairly liberal estimates of storm undercounts in the early part of the Atlantic record, the correlation with tropical Atlantic SST remains high (Mann et al. 2007).

PALEOTEMPESTOLOGY. A number of remarkable efforts are underway to extend tropical cyclone climatology into the geological past by analyzing paleoproxies for strong wind storms. One technique looks at storm surge–generated overwash deposits in near-shore marshes and ponds; this was pioneered by Liu and Fearn (1993) and has been followed up with analyses of such deposits in various places around the western rim of the North Atlantic (Liu and Fearn 2000; Donnelly et al. 2001a,b; Donnelly 2005; Donnelly and Woodruff 2007). Another technique makes use of dunes of sand, shells, and other debris produced along beaches by storm surges (Nott and Hayne 2001; Nott 2003). Very recently, new techniques have been perfected that make use of the anomalous oxygen isotope content of hurricane rainfall (Lawrence and Gedzelman 1996) as recorded in tree rings (Miller et al. 2006) and speleothems (Frappier et al. 2007b). Collectively, these methods are beginning to reveal variability of tropical cyclone activity on centennial to millennial time scales. For example, the recent work of Donnelly and Woodruff (2007), analyzing overwash deposits near Puerto

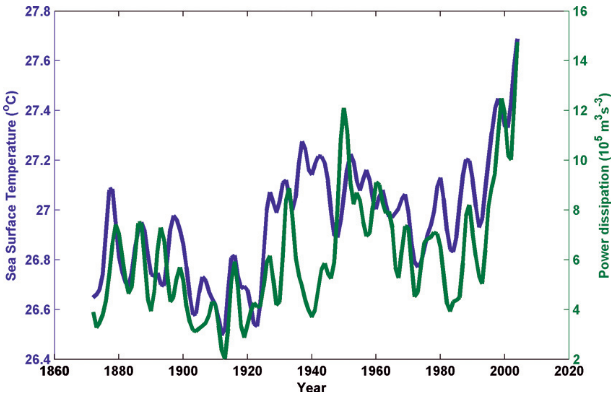

Fig. 2. Storm lifetime maximum power dissipation in the North Atlantic according to (green) data from the NOAA National Hurricane Center as adjusted by Emanuel 2005a. (blue) Aug–Oct SST in the tropical North Atlantic, from 20° to 60°W and from 6° to 18°N. Both quantities have been smoothed using a 1-3-4-3-1 filter. The SST is from the Met Office Hadley Centre SST dataset (HadSST1).

Rico, reveals centennial variability of Atlantic tropical cyclones that is highly correlated with proxies recording long-term variability of ENSO; the same record shows a pronounced upswing over the last century that may reflect a global warming signal. The interested reader is directed to reviews by Nott (2004), Liu (2007), and Frappier et al. (2007a).

ATTRIBUTION. The North Atlantic is the only basin with a reasonably long time series of tropical cyclone records, and it is clear from Fig. 2 that there is variability on a broad spectrum of time scales. A Fourier decomposition of the detrended, unfiltered time series of storm maximum power dissipation shows prominent spectral peaks at around 3, 5, 9, and 80 yr. Similar spectral peaks are evident in the detrended SST data. The first two of these are likely associated with ENSO, known to have a strong effect on Atlantic hurricanes (Gray 1984). The longest period spectral peak at 80 yr is of dubious significance, given that the time series is only ~130 yr long, but it is clear from inspection of Fig. 2 that both SST and tropical cyclone power have seesawed up and down on a multidecadal time scale over the past century or so.

Mestas-Nuñez and Enfield (1999) examined rotated empirical orthogonal functions (EOFs) of the detrended global SST and identified the first six of these with modes[2] of the ocean–atmosphere system. The first EOF had time scales of many decades and maximum amplitude in the North Atlantic; this was later identified as a prominent cause of both SST and Atlantic tropical cyclone variability on multidecadal time scales (Goldenberg et al. 2001) and christened the "Atlantic Multi-Decadal Oscillation" or "AMO" (Kerr 2000). What began as an EOF ended up as a mode, even though there are only two troughs and one peak in the time series. It is important to recognize that this EOF is global, and while it has large amplitude in the North Atlantic, its amplitude is almost as large in the North Pacific. Furthermore, it turns out that the time series of the amplitude of this first EOF is barely distinguishable from the detrended time series of August–October tropical North Atlantic SST, so that there is little advantage in referring to this EOF versus the raw SST. We can ask the somewhat more direct question: What caused the tropical North Atlantic SST (and tropical cyclone power) to seesaw as it did during the twentieth century, as evident in Fig. 2?

Figure 3 provides one clue. This compares the 10-yr running averages of the August–October SST of the so-called main development region (MDR) of the tropical North Atlantic (between Africa and the eastern Caribbean) with the Northern Hemisphere mean surface temperature (including land). The excellent correspondence between the two time series would seem to imply that on decadal time scales, over the last 100 yr or so, the tropical North Atlantic is simply covarying with the rest of the Northern Hemisphere. Occam's razor would lead one to suspect that variations of the two series have a common cause, though it has been suggested that the North Atlantic might be forcing the rest of the hemisphere (Zhang et al. 2007).

The decadal variability in the Northern Hemispheric surface temperature has been addressed in a number of studies, as summarized in the most recent report of the Intergovernmental Panel on Climate Change (IPCC 2007). In contrast to Mestas-Nuñez and Enfield (1999), Goldenberg et al. (2001), and others, the IPCC report attributes most of the decadal variability to time-varying radiative forcing associated principally with varying solar radiation, major volcanic eruptions, and anthropogenic sulfate aerosols and greenhouse gases. This also helps explain the overall trend, which was disregarded in the EOF analyses. In particular, the warming of the last 30 yr or so is attributed mostly to increasing greenhouse gas concentrations, while the cooling from around 1950 to around 1980 is ascribed, in part, to increasing con-

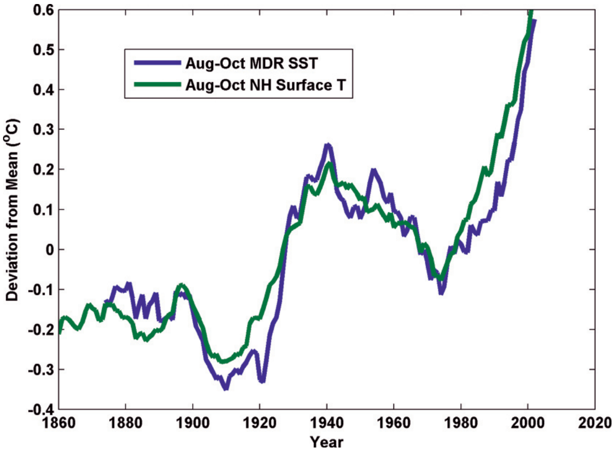

Fig. 3. Ten-year running averages of the (blue) Atlantic MDR SST and (green) the Northern Hemispheric surface temperature, both averaged over Aug–Oct. The long-term mean has been subtracted in both cases. The Met Office Hadley Centre supplies the SST data (HadSST1) and the Northern Hemispheric surface temperature [Hadley Centre Climate Research Unit (HadCRU)].

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[2] This is technically an incorrect term, as modes are not mathematically equivalent to EOFs.

centrations of anthropogenic sulfate aerosols. Mann and Emanuel (2006) pointed out that the cooling of the Northern Hemisphere relative to the globe from about 1955 to 1980, evident in Fig. 3, might very well be explained by the concentration of sulfate aerosols in the Northern Hemisphere. While there is still a great deal of uncertainty about the magnitude of the radiative forcing due to sulfate aerosols, the time series of sulfate concentration is strongly correlated with the difference between global and Northern Hemisphere surface temperature (Mann and Emanuel 2006). The important influence of anthropogenic effects in the time history of SST is also emphasized in the work of Hoyos et al. (2006), Trenberth and Shea (2006), Santer et al. (2006), Elsner (2006), and Elsner et al. (2006). Emanuel (2007) emphasizes that the thermodynamic control on tropical cyclone activity is exercised not through SST but through potential intensity, which in the North Atlantic has increased by 10% over the past 30 yr. This increase, which is greater than predicted by single-column models for the observed increase in SST, can be traced to increasing greenhouse gases, decreasing surface wind speed in the tropics, and also to decreasing lower-stratospheric temperature (Emanuel 2007).

Thus there are two schools of thought about the decadal variability of tropical North Atlantic SST and tropical cyclone activity. The first holds that the multidecadal variability is mostly attributable to natural oscillations of the ocean–atmosphere system (Goldenberg et al. 2001; Kossin and Vimont 2007), while the second attributes it to time-varying radiative forcing, some of which is natural. These two schools are not mutually exclusive, as the response to time-varying radiative forcing can be greatly modified by natural modes of the system.

Those who attribute Atlantic SST and tropical cyclone variability to a putative AMO often refer to paleoproxy evidence for its existence. The most prominently cited among this evidence is the work of Gray et al. (2004), who looked at the first five principal components of variability in tree rings around the rim of the North Atlantic. They fitted these components to the instrumental record of North Atlantic SST, capturing the prominent variability apparent in Fig. 2, and then, using the same curve fit, inferred North Atlantic SST from tree rings back to the middle of the sixteenth century. Although the record obviously shows the observed multidecadal variability of the twentieth century, close inspection of the reconstructed SST prior to this shows that most of the variability was on somewhat longer time scales, casting doubt on the existence of a quasi-periodic mode. In fact, the only real evidence for the existence of such a mode comes from coupled climate models (e.g., Delworth and Mann 2000), and although many of them exhibit prominent quasi-periodic variability on time scales greater than a decade, the period of such oscillations varies greatly from model to model. It is left to the reader to judge whether the existence of such modes in freely run coupled climate models constitutes strong or weak evidence for such modes in nature.

SIMULATING GLOBAL WARMING EFFECTS ON TROPICAL CYCLONES.

Another approach to understanding how climate change might affect tropical cyclone activity is to simulate changing tropical cyclone activity using global climate models. Unfortunately, the horizontal resolution of today's generation of global models is nowhere near sufficient to resolve the intense inner cores of tropical cyclones, and numerical resolution experiments (e.g., Chen et al. 2007) suggest that grid spacing of no more than a few kilometers is necessary for convergence. Nevertheless, there are quite a few studies of the response of tropical cyclone activity to global warming using global models (e.g., Bengtsson et al. 1996; Sugi et al. 2002; Oouchi et al. 2006; Yoshimura et al. 2006; Bengtsson et al. 2007). A related approach involves embedding finer-resolution regional models within global climate models, so as to better simulate tropical cyclones (Knutson et al. 1998; Knutson and Tuleya 2004; Knutson et al. 2007). Although results can differ greatly from model to model, there is a general tendency for global warming to reduce the overall frequency of events, to increase the incidence of the most intense storms, and to increase tropical cyclone rainfall rates.

Another approach to downscaling global models to derive tropical cyclone climatologies was presented by Emanuel (2006) and Emanuel et al. (2006); this has recently been extended to account for varying genesis rates (Emanuel et al. 2008). Using certain key statistics from the output of climate models, this technique synthesizes very large numbers (~10^3) of tropical cyclones using a three-step process. In the first step, the climate state is "seeded" with a large number of candidate tropical cyclones, consisting of warm-core vortices whose maximum wind speed is only 12 m s^{-1}. These candidate storms then move according to a "beta-and-advection" model (Marks 1992), which postulates that tropical cyclones move with a weighted tropospheric mean large-scale flow in which they are embedded, plus a correction owing to gyres generated by the storm's advection of planetary vorticity; here the large-scale flow is taken as the climate model-simulated flow. Finally, in the third step, the storm's intensity evolution is simulated using a deterministic, coupled ocean–atmosphere tropical cyclone model phrased in angular momentum coordinates, which achieve very high spatial resolution in the critical central core region. In practice, most of the seeds die a natural death owing to small potential intensity, large wind shear, and/or low humidity in the middle troposphere. We show that the climatology of the survivors is in good accord with observed tropical cyclone climatology.

While details of this technique and the results of applying it to a suite of global climate models are presented in Emanuel et al. (2008), we here show one critical result of comparing tropical cyclone activity in the late twentieth century to that of the late twenty-second century as simulated by global climate models under IPCC scenario A1b, in which atmospheric CO_2 concentrations continue to increase to 720 ppm by 2100, after which they are held constant. Figure 4 shows the percentage increase in "power dissipation" in five ocean basins using seven climate models, deduced using 2,000 synthetic events in each model, in each basin, and for each of the twentieth-century and A1b simulations. Power dissipation is just the sum over each storm's lifetime and over a given year of the cube of its maximum wind speed and is a rough measure of potential destructiveness of tropical cyclones.

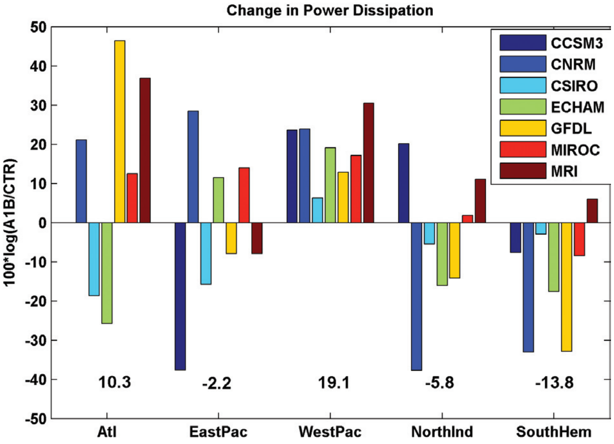

Fig. 4. Change in tropical cyclone power dissipation between the last 20 yr of the twentieth century and the last 20 yr of the twenty-second century, based on 2,000 synthetic storms in each of five ocean basins for each of seven global climate models. The change here is given as 100 multiplied by the logarithm of the ratio of the 22nd century and 20th century quantities, and the values of the changes averaged across all models are given by the numbers in black. The twenty-second-century statistics are taken from models forced according to IPCC scenario A1b. From Emanuel et al. (2008).

Results vary greatly from model to model, reflecting the general uncertainties remaining in the field of climate modeling. There is a great tendency for the frequency of events (not shown) to decline, as is the case with direct simulations using global models; this is offset by a tendency for increased intensity. The decline in frequency of events in these simulations is owing to the increase in the magnitude of an important nondimensional parameter in the intensity model. This parameter, χ_m, is defined

$$\chi_m \equiv \frac{s_m - s_m^*}{s_0^* - s_b}, \qquad (2)$$

where s_m is the moist entropy of the middle troposphere (near the level where it attains a minimum value), s_m^* is its saturation value, s_0^* is the moist entropy of air saturated at sea surface temperature and pressure, and s_b is the moist entropy of the boundary layer. This quantity is nonpositive, and its magnitude measures the degree of thermodynamic inhibition to tropical cyclone formation. It is easy to show that at constant relative humidity, the numerator of (2) scales with the saturation specific humidity, as dictated by Clausius–Clapeyron. On the other hand, the denominator measures the air–sea thermodynamic disequilibrium, which, at constant surface wind speed, is proportional to the surface turbulent energy flux into the atmosphere. This, in turn, rises only slowly with global warming, since surface evaporation is constrained to balance the net surface radiative flux, which changes only slowly, once the surface temperature becomes fairly large. Thus global warming has the effect of decreasing tropical cyclone frequency. At the same time, potential intensity generally increases with warming, so that some increase in the frequency of the most intense events is to be expected.

EFFECT OF TROPICAL CYCLONES ON CLIMATE.

Discussions of tropical cyclones and climate almost always assume that any changes in tropical cyclone activity are passive; that is, there is little or no feedback of tropical cyclones on the climate system. Globally, tropical cyclones contribute only a few percent of the total precipitation (and thus latent heat release) in the tropics; on the other hand, their precipitation efficiency is anomalously high, so that they may serve to dehydrate the tropical atmosphere to some degree. This might serve to cool the tropics, owing to the decline of the greenhouse effect of water vapor. Because of the very high specific entropy

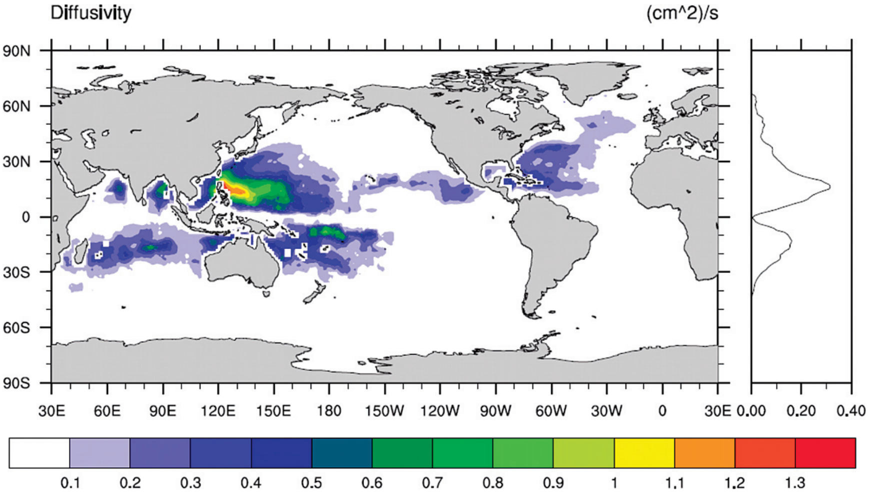

Fig. 5. Vertical diffusivity induced by tropical cyclones, estimated from European Centre for Medium-Range Weather Forecasts reanalyses, reproduced from Sriver and Huber (2007). The panel at right shows the zonal average.

content of the tropical cyclone eyewall, they can extend farther into the lower stratosphere than most convection; so it is possible that they play a role in the regulation of stratospheric water vapor. But perhaps their greatest influence on climate is exerted through the oceans.

Tropical cyclones are observed to vigorously mix the upper ocean (Leipper 1967). The mechanism for doing this is somewhat indirect. Because of their horizontal scale and translation speeds, tropical cyclones are particularly efficient in exciting near-inertial oscillations in the upper ocean (Price 1981). Vertical shear of ocean currents across the base of the mixed layer is almost invariably unstable, resulting in small-scale turbulence that mixes colder thermocline waters into the mixed layer, thereby cooling it and warming the upper thermocline (Price 1981). This mixing occurs on time scales of 6–24 h associated with the passage of storms and the near-inertial response to the time-varying wind stress they produce. The mixing itself does not change the column-integrated enthalpy; enthalpy is merely redistributed in the vertical. However, the cold anomaly produced at the surface is observed to recover over a period of about 10 days (Nelson 1996), owing to a reduction in the turbulent enthalpy flux to the atmosphere. This wake recovery is associated with a net, column-integrated enthalpy increase in the ocean. Assuming that all of the cold anomaly recovers, Emanuel (2001) estimated that global tropical cyclone activity results in an average net heat input rate of 1.4 ± 0.7 PW to the tropical oceans, a number comparable to the total poleward heat transport by the oceans. Recently, a more conservative estimate of around 0.3 PW was made by Sriver and Huber (2007), who used European Centre reanalyses to estimate cold wake recovery.[3] Figure 5, reproduced from Sriver and Huber (2007), shows the estimated vertical diffusivity induced by global tropical cyclone activity.

Experiments with ocean models show that spatially and temporally isolated mixing events are as effective as broadly distributed mixing in inducing a poleward heat transport in the ocean (Scott and Marotzke 2002; Boos et al. 2004), so that much of the upper ocean heat uptake induced by tropical cyclone mixing is exported toward higher latitudes, though some may return to the atmosphere locally in the tropics in the subsequent cool season.

It is possible that the cold wake may recover only through a very shallow depth, leaving a dipole temperature anomaly (or "heton") in the upper ocean, with very little change in the column-integrated enthalpy. In principle, the total heat uptake during wake recovery should be reflected in an elevation of the sea surface, which is detectable using satellite-based sea surface altimetry. Using the hydrostatic equation, the change in column enthalpy content, Δk, should be related to the change in sea surface elevation, Δz, by

$$\Delta k = \frac{\rho c_l}{\alpha}\Delta z, \tag{3}$$

where ρ is the density of seawater, c_l is its heat capacity, and α is its coefficient of thermal expansion. Figure 5 of Emanuel (2001), reproduced here as Fig. 6, shows sea surface elevation as a function of time and cross-track distance during the wake recovery of Atlantic Hurricane Edouard in 1996.

One observes that the sea surface rises mostly to the right of the storm track, where the largest near-

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[3] Cold wakes were assumed to penetrate only to 50-m depth, and SSTs are updated as infrequently as 7 days in the reanalyses, leading to underestimation of cold wake magnitude, so that this estimate is conservative.

inertial response and cooling occurs, and that the surface rises by about 5 cm. According to (3), this gives a heat uptake of about 8×10^8 J m^{-2}, which, when integrated over the approximately 800-km width and 3,000-km length of the wake, yields a total heat uptake of around 2×10^{21} J. If there were 15 such events globally each year, the average rate of induced heat uptake would be about 1×10^{15} W, consistent with the earlier estimate by Emanuel (2001). In particular, the magnitude of the sea surface height response evident in Fig. 6 suggests that wake recovery was deep in this case.

The implications of this for climate dynamics should not be understated. As pointed out by Emanuel (2001), increased tropical cyclone activity in a warmer climate would result in increased tropical heat export by the oceans, mitigating tropical warming but amplifying the warming of higher latitudes. This inference is supported by recent numerical simulations using a coupled climate model in which upper ocean mixing is related to a proxy for tropical cyclone activity (Korty et al. 2008). This effect offers a potential explanation for the equable nature of very warm climates, such as that of the early Eocene; high levels of tropical cyclone activity in such warm climates could drive a strong poleward heat flux in the ocean, even in the face of relatively weak pole-to-equator temperature gradients, thus helping to keep such gradients weak. (Today's coupled climate models are notoriously bad at reproducing such weak temperature gradients, perhaps because they have no representation of tropical cyclone–induced ocean mixing.) It may also help explain why most of the observed heat uptake by the oceans over the past 50 yr has been in the subtropics and middle latitudes (Levitus et al. 2005), whereas coupled models typically show most of the heat uptake occurring in subpolar regions (e.g. Manabe et al. 1991).

Clearly, tropical cyclone–induced heat uptake may be an important element of climate dynamics and should remain an active research topic for the next few years at least.

SUMMARY. Tropical cyclones respond to climate change in a number of ways. Their level of activity appears to be controlled primarily by four factors: potential intensity, vertical shear of the horizontal environmental wind, low-level vorticity, and the parameter χ_m defined by (2) and measuring the specific humidity deficit of the middle troposphere. Records of tropical cyclones are best and longest in the North Atlantic, are somewhat less reliable in the western North Pacific, and are dubious elsewhere, particularly before the satellite era. In the North Atlantic region, tropical cyclone power dissipation is highly correlated with tropical sea surface temperature during hurricane season on time scales of a few years and longer. The tropical North Atlantic sea surface temperature is in turn highly correlated with Northern Hemisphere surface temperature, at least during hurricane season, on time scales of a decade and longer. The weight of available evidence suggests that multidecadal variability of hurricane season tropical Atlantic SST and Northern Hemispheric surface temperature, evident in Fig. 3, is controlled mostly by time-varying radiative forcing owing to solar variability, major volcanic eruptions, and

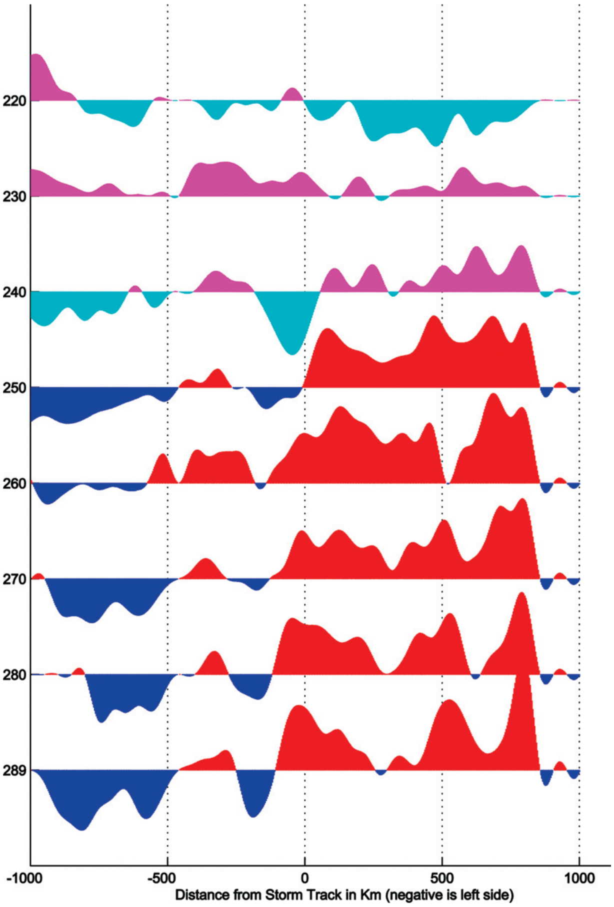

Fig. 6. Cross-track sections of the sea surface height anomaly from TOPEX/Posiden at 10-day intervals in Aug–Sep 1996. Hurricane Edouard passed this transect on Julian day 239. The height anomaly corresponding to the vertical separation between the transects is 20 cm. The transect is centered at 19.2°N, 56°W and the time of the transect is indicated at left by Julian day. The anomalies represent differences from the sea surface height averaged over the month preceding Julian day 220. (Analysis and figure courtesy of Peter Huybers.)

anthropogenic sulfate aerosols and greenhouse gases, though the response to this forcing may be modulated by natural modes of variability. The increase in potential intensity of about 10% in the North Atlantic over the last 30 yr was driven by increasing greenhouse gas forcing, declining lower-stratospheric temperature, and decreasing surface wind speed (Emanuel 2007); this increase is consistent with the ~60% increase in tropical cyclone power dissipation during this time.

Explicit simulations of tropical cyclones using global climate models as well as a variety of downscaling techniques all show a general tendency toward decreasing tropical cyclone frequency and increasing intensity and rainfall rates, although there is much variability from model to model and from ocean basin to ocean basin. The increased intensity is related to increasing potential intensity as the climate warms, while the increased rainfall rate is a straightforward consequence of increased atmospheric humidity, according to Clasius–Clapeyron. The decreasing frequency of tropical cyclones appears to be owing to an increase in the magnitude of the thermodynamic inhibition to genesis, as given by the parameter χ_m defined by (2); this is a predictable consequence of global (as opposed to local) warming.

Tropical cyclones may affect climate through drying of the troposphere and especially by mixing the upper tropical oceans. Available evidence suggests that global tropical cyclone activity may be an important or even dominant mechanism in maintaining poleward heat flux by the oceans. Since tropical cyclones both respond to and affect climate change, their existence modifies climate dynamics in a way that may help explain both the pattern of recent heat uptake by the oceans and the peculiar features of very warm climates, such as that of the early Eocene. Further research needs to be undertaken to explore these ideas.

ACKNOWLEDGMENTS. I thank the modeling groups for making their simulations available for analysis, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) for collecting and archiving the CMIP3 model output, and the WCRP's Working Group on Coupled Modeling (WGCM) for organizing the model data analysis activity. The WCRP CMIP3 multimodel dataset is supported by the Office of Science, U.S. Department of Energy. I was supported by Grant ATM-0432090 from the National Science Foundation.

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