Relationship between the potential and actual intensities of tropical cyclones on interannual time scales

A. A. Wing,[1] A. H. Sobel,[2] and S. J. Camargo[3]

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[1] The thermodynamic theory for the physics of a mature tropical cyclone (TC) tells us that the cyclone's intensity cannot exceed an upper bound, the potential intensity (PI). This combined with an empirical result due to Emanuel leads to a prediction of average TC intensity change, given the change in PI. The slope of the predicted relationship between percentagewise variations in PI and those in intensity can vary between 0.5 and 1, depending on the mean PI and on what threshold is applied to the intensity data. For the Atlantic and Pacific, typical values are around 0.65 when tropical storms are excluded and 0.8 when they are included. The authors use best track data for the North Atlantic and western North Pacific, combined with PI computed from reanalysis data sets, to test these predictions. The results show that observed interannual variations of maximum TC intensity are consistent with the predictions of PI theory. Modest fractions of the variance in actual intensity are explained by PI variations. Much of the interannual variation in PI experienced by the storms comes from variation in TC tracks, so that the storms in different years are more or less likely to sample regions of high PI, rather than from variations in PI at a fixed location.
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1. Introduction

[2] Recent analyses of archived "best track" data sets of tropical cyclone intensity show a trend in the last 30 years toward more frequent, longer-lived, and more intense storms, and suggest that this is related to large-scale increases in sea surface temperature (SST) [*Emanuel*, 2005a; *Webster et al.*, 2005; *Emanuel*, 2007]. While it may be debated whether these increases represent secular trends or oscillatory cycles, particularly in the Atlantic [*Landsea*, 2005; *Goldenberg et al.*, 2001], given the uncertainties it is clear that anthropogenic climate change has at least the potential to influence the frequency and intensity of tropical cyclones. For this and other reasons, it is of interest to obtain a better understanding of the way in which changes in large-scale environmental variables, of which SST is an important but not the only important one, lead to changes in the statistics of tropical cyclones.

[3] A thermodynamic theory exists for the maximum, or potential intensity (PI) of steady state tropical cyclones, given SST and a few other environmental variables [*Emanuel*, 1986, 1988; *Holland*, 1997; *Bister and Emanuel*, 1998]. Most storms do not reach their PI, and no comparable theory exists for the actual intensity a given storm will reach. *Emanuel* [2000] performed an empirical study in which he examined cumulative distribution functions of storm intensity in the best track data base. Emanuel sampled the data in various ways, in some cases excluding tropical storms, and in all cases either excluding or modifying the maximum intensity of any storm whose maximum intensity exceeded its PI at the time of maximum intensity. Actual intensity can exceed PI when a storm moves quickly from a region of higher to lower PI (since the time for a storm's intensity to adjust to its environment is finite) or due to errors in the PI theory itself, or in the data used to compute PI.

[4] In all cases Emanuel found that the cumulative distribution function of observed maximum wind speed, V, normalized by PI, is approximately linear between a lower bound and 1. This implies a uniform probability distribution function (PDF) over that interval. The lower bound corresponded approximately, for a typical value of PI, to hurricane intensity. A uniform PDF indicates an equal likelihood that any given tropical cyclone will achieve any given intensity up to its PI. *Emanuel* [2000] stated that this implies that a given percentagewise climatic change in PI will lead to an equal percentagewise change in the average actual intensity of tropical cyclones, corresponding to a slope of 1 in the relationship between PI and intensity when both are normalized by their climatological means. This is true if the lower and upper bounds on V/P in the sample (where P is the PI) remain constant as PI varies. The predicted change in intensity is different if a constant lower bound on intensity itself is imposed, such as is the case if only storms reaching at least hurricane intensity, or tropical storm intensity, are included in the sample. In this case, as shown in the Appendix, the normalized slope is $\overline{P}/(\delta + \overline{P})$, where $\overline{P}$ is the mean PI and δ is the lower bound on intensity.

[5] *Emanuel* [2007] evaluated the intensity trend over the last 30 years in the North Atlantic basin, which is the best-observed basin, including all storms, and found that it was approximately equal in percentage terms to that in PI, roughly consistent with theoretical expectations. Our purpose here is to test whether the predictions of PI theory continue to hold when a broader spectrum of variability, including in particular interannual fluctuations, is considered. We analyze the relationship between PI and actual

[1]Department of Earth and Atmospheric Sciences, Cornell University, Ithaca, New York, USA.

[2]Department of Applied Physics and Applied Mathematics and Department of Earth and Environmental Sciences, Columbia University, New York, New York, USA.

[3]International Research Institute for Climate and Society, The Earth Institute at Columbia University, Palisades, New York, USA.

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tropical cyclone intensities since 1950 for the Atlantic and western North Pacific basins.

2. Data and Methods

[6] We used the best track data from the North Atlantic and western North Pacific ocean basins to define both the tracks and intensities (measured by the maximum wind speeds) of tropical cyclones. We removed tropical depressions from the data set, so that this study considers only storms reaching at least tropical storm intensity. The Atlantic best track data, which cover the period 1950–2005, were produced by the National Oceanographic and Atmospheric Administration's National Hurricane Center/Tropical Prediction Center. The western North Pacific best track data, which cover the period 1950–2004, were produced by the US Navy's Joint Typhoon Warning Center. Intensity trends computed from these data sets can be expected to have systematic biases due to changes in observing practices over the years. This is one motivation to focus on interannual variability, as artifacts due to these changes in observing practices are not likely to have as large an influence on estimates of interannual fluctuations as on those of the trends. *Emanuel* [2005a] developed adjustments to the earlier portions of the best-track intensity records in order to make the relationship between reported minimum pressure and maximum wind consistent over the entire time period. These adjustments have been debated [*Landsea*, 2005; *Emanuel*, 2005b]; we do not apply them in the calculations shown below, but do so in auxiliary materials[1] as a sensitivity test. These adjustments have a small negative effect on the results.

[7] We computed PI according to *Emanuel* [1988, 1995], with modifications described by *Bister and Emanuel* [1998]. Atmospheric temperature and humidity profiles, as well as SST, are needed to compute PI. We used the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis data [*Kalnay et al.*, 1996], available for 1950 to 2005. To correct for spurious discontinuities due to the introduction of satellite data around 1979, we apply the fix of *Emanuel* [2007], subtracting 1.9 m s^{-1} from all PI values prior to 1979. This makes only small differences in the statistics shown below. For comparison, results were also computed using the European Centre for Medium-Range Forecasting's ERA-40 data set [*Uppala et al.*, 2005]; these results (not shown) were very similar to those computed from the NCEP/NCAR data.

[8] Both the best track intensities and the computed PI data were sampled and averaged in several ways to create time series of annual average values that could be compared to each other. One pair of PI time series was computed by averaging the PI over the Atlantic main development region (MDR), defined as 10°–20°N, 80°–20°W, from August through October, and the area from 5°–25°N and 100°–160°E from July through October for the Western North Pacific. Comparing these basin-averaged PI time series to the time series of actual storm intensity does not constitute a true test of the PI theory; to do that, the two should be compared along the actual tracks, as we describe below. However, basin averages are of interest because they are all that one can compute in a seasonal or longer-term forecast situation, when the storm tracks are not yet known but a forecast of the PI field may be available, e.g., from a general circulation model. We also computed a PI time series along the track of each specific storm in the best track data base. These PI time series and the actual intensity time series were both sampled at the first point at which each storm's maximum intensity occurred, resulting in one single value for each storm's intensity and a corresponding PI value. (Because intensity is estimated only in 5 *kt* increments, often the maximum intensity is reached at more than one point along the track, where each point represents a 6-hour period; averaging together all points at which maximum intensity was reached gives similar results (not shown) to those obtained using only the first such point.) All such values occurring in each single year were averaged together to create time series of PI and intensity with one point per year. This analysis was performed for all storms (reaching at least tropical storm intensity), and then for the subset of all storms that reached at least hurricane intensity. In each case we compare the PI and intensity time series, computing their correlations and the slopes of the regression lines obtained from least-squares fits. The significance of correlations is computed using a threshold of 95% and assuming the number of degrees of freedom is equal to half the number of years in the sample, which approximately accounts for the autocorrelation in the time series. 95% confidence limits on regression slopes are computed, and mentioned below, but not shown. The confidence limits computed assuming each year is independent are already quite generous, in that agreement between the computed regression slopes and the theoretical slopes tends to be much closer than the confidence limits. Accounting for autocorrelation will broaden the confidence limits further.

[9] Storms exceeding their PIs at the time of maximum intensity are included in our analysis, in contrast to that of *Emanuel* [2000]. Such storms are arguably outside the expected domain of validity of PI theory, but the theory we are testing is partially built on empirical results in any case, and it seems reasonable to ask whether it continues to hold when these storms are included. When tropical storms are included, excluding storms for which actual intensity exceeds PI at the time of maximum intensity has a significant effect on the results, improving the agreement with the theoretical prediction significantly, as described below. When tropical storms are excluded, excluding in addition those storms with $V > P$ at the time of maximum V reduces the sample size to the point that only one or two storms are left in some years. We judged this to be too small a sample for an analysis of interannual variability, and left these storms in; the resulting time series are nonetheless in good agreement with the theoretical predictions.

3. Results

3.1. Atlantic

[10] Figure 1a shows time series of wind speed, *V*, and potential intensity *P*, averaged over the first point of maximum intensity for each storm reaching at least hurricane intensity, for the Atlantic basin. The time series shown have not been detrended or normalized. Because storms

[1]Auxiliary materials are available in the HTML. doi:10.1029/2006GL028581.

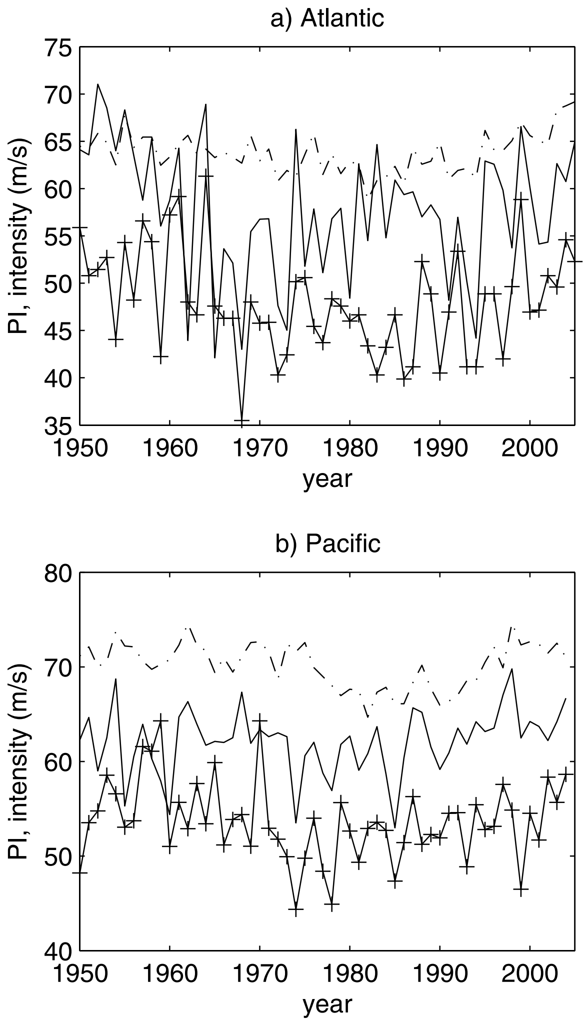

Figure 1. Time series of wind speed (crosses) and potential intensity (solid and dot-dash curves). Wind speed is computed by averaging the maximum intensities reached by all storms in a given year. Potential intensity is computed by averaging the values occurring at the first point on each track at which each storm reached its maximum intensity, over all storms for each year which reached at least hurricane intensity (solid curve). The dot-dash curve shows PI averaged over the main development region (10°–20°N, 80°–20°W for the Atlantic; 5°–25°N, 100°–160°E for the western North Pacific) and peak season (August–October for the Atlantic, July–October for the western North Pacific). Data for the (a) Atlantic and (b) western North Pacific.

exceeding their PIs are allowed in the sample, it is possible for the curves to cross, and this occurs once in the Atlantic record and twice in the Pacific. The basin average PI over the MDR for ASO is also shown. This plot shows that much of the variance in all three time series is interannual (as opposed to multidecadal, though some of the latter is certainly also present). It also shows that the MDR-averaged PI varies less, in absolute as well as relative terms, than does the PI averaged on the points of maximum intensity. This indicates that much of the PI variation experienced by the storms from year to year is due to track changes which cause storms to sample regions of greater or lesser PI, as opposed to PI changes at fixed locations. A similar conclusion was reached by *Kossin and Vimont* [2007].

[11] Figure 2a shows a scatter plot of the wind speed and potential intensity data averaged over the points of maximum intensity as in Figure 1. For this plot, both time series have been detrended and normalized by their respective climatological means. The two quantities have a significant correlation of $r = 0.52$. The slope of the least squares regression line calculated for normalized wind speed vs. PI is 0.49. The theoretical slope in this case is 0.65, which is well within the 95% confidence limits of the slope obtained from the data. These results were computed including all storms reaching hurricane intensity in each year; restricting the sample to storms occurring in the peak August to October (ASO) season does not significantly change the results.

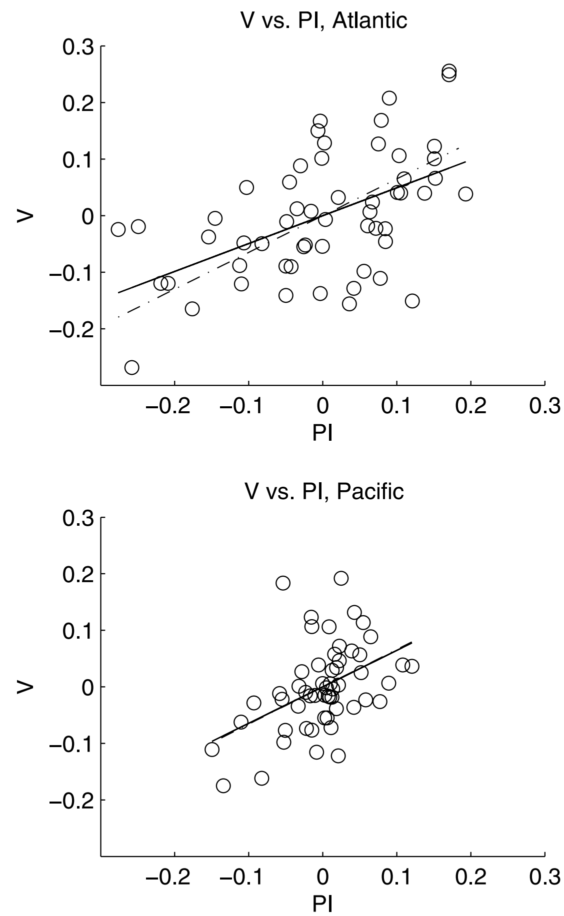

Figure 2. Scatter plots of wind speed and potential intensity. Wind speed is computed by averaging the maximum intensities reached by all storms in a given year, for those storms reaching at least hurricane intensity. Potential intensity is computed by averaging the values occurring at the first point on each track at which each storm reached its maximum intensity, over all storms for each year reaching at least hurricane intensity. Data for the (a) Atlantic and (b) western North Pacific. Both time series have been normalized and detrended, and have had their means removed. Least-squares regression lines computed from the data (solid curve) as well as the theoretical linear relationship (dot-dash curve) are also shown.

Table 1. Linear Regression Statistics[a]

Points Included	Correlation	Slope	Theory
Atl - tracks	**0.52**	0.49	0.65
Atl-TS - tracks	**0.36**	0.70	0.79
Atl - MDR	**0.46**	1.63	—
Pac - tracks	**0.45**	0.64	0.66
Pac-TS - tracks	**0.39**	1.05	0.79
Pac - MDR	0.11	—	—

[a]Data are shown for PI computed on the tracks, as in Figure 2, for the Atlantic (Atl - tracks) and Pacific (Pac - tracks); and for PI averaged over the main development regions (MDRs) and peak seasons, compared to intensity also sampled only over peak seasons, for the Atlantic (Atl - MDR) and Pacific (Pac - MDR). Atl-TS - tracks and Pac-TS - tracks are sampled as in Atl - tracks and Pac - tracks, but tropical storms are included, and storms whose actual intensities exceeded their potential intensities at the time of maximum intensity are excluded. The first column shows correlation coefficients, in bold if significant at 95% (see text for details). The second column shows the slope of the least-squares linear regression line, while the third column shows the theoretical slope, as computed from (1). In all cases in which the correlation is significant and a theoretical slope is shown, the regression slope is within the 95% confidence limits computed assuming each year is independent; accounting for autocorrelation will widen the confidence limits.

[12] We also compared the average maximum intensity of storms reaching at least hurricane intensity during the ASO season to PI averaged spatially over the MDR and over the ASO season. The slope of the regression line calculated for the detrended and normalized time series is 1.63, and the correlation coefficient is 0.51. The storms' intensities are consistent with PI theory when actual PI is considered, as shown in Figure 2a, but many storms reach their maximum intensities outside the MDR, and PI varies spatially within the MDR as well. Mean PI variations over the MDR are thus not directly relevant to the interannual variations in maximum intensities of the actual storms.

[13] When tropical storms are included in the sample, the correlation of actual and potential intensity (both sampled on the tracks at the first point of maximum intensity) is reduced to 0.38, and the regression slope is 0.42, compared to a theoretical prediction of 0.77. When tropical storms are included, but storms with actual intensity exceeding PI at the time of maximum intensity are excluded (similar but not identical to the "modified maximum intensity" of *Emanuel* [2000]), the correlation between the two quantities is reduced to 0.36, which is marginally significant. The least-squares slope is increased to 0.70. The theoretical slope of 0.79 is well within the 95% confidence limits.

[14] The results discussed above include variability on all time scales resolved by the data, excluding linear trends but including both interannual and multidecadal time scales. We also performed calculations for the Atlantic using only the period 1970–2005, removing the linear trend over that period. This is a crude proxy for the removal of multidecadal variations, as these are approximately manifest as a rise during the 1970–2005 period following a decline earlier [*Goldenberg et al.*, 2001; *Emanuel*, 2005a; *Webster et al.*, 2005]. The Atlantic during this period is also of interest because the best track data set for that time and place is more reliable than any other long-term TC data set in existence. Compared to the results from the full record, this calculation gives a lower, but still marginally significant correlation of 0.35, when hurricanes only are included. The least-squares regression slope in this case is found to be 0.32, while the theoretical value is 0.64. When tropical storms are included in this calculation, but those for which actual intensity exceeded PI at the time of maximum intensity are excluded, the correlation between PI and intensity is not significant.

3.2. Western North Pacific

[15] An analysis analogous to that described above was also performed using the western North Pacific data. The results are shown in Figures 1b and 2b. Comparing the detrended and normalized time series of annual average potential intensity and annual average wind speed, including points of maximum intensity for storms reaching hurricane strength or greater (Figure 2b) a correlation coefficient of r = 0.45 and a slope of 0.64 (very close to the theoretical slope of 0.66, see Table 1) were obtained. When the sample is restricted to storms occurring in the July to October (JASO) peak season, the correlation increases to 0.57. The slope also increases to 0.92, but the theoretical slope of 0.66 is still inside the 95% confidence interval. This change in slope is nonetheless larger than that found in the Atlantic, where, as described above, limiting the sample to the peak season did not affect the result significantly. While we cannot explain the Pacific behavior in detail, its difference from that in the Atlantic may be related to the tendency of Atlantic storms to occur primarily during or near their peak seasons, while western North Pacific storms can occur all year, with significant variations in their genesis locations and tracks over the annual cycle [*Lander*, 1996]. When tropical storms are included, but storms exceeding their PI at peak intensity are excluded, the regression slope is 1.05, and the theoretical slope of 0.79 is well within the confidence limits.

[16] The comparison of the basin average (5°–25°N, 100°–160°E) PI and intensity did not yield a statistically significant correlation. This is not entirely surprising, as it is known that local SST is not correlated with TC intensity in this region on an interannual basis [*Chan and Liu*, 2004]. PI variations with ENSO also tend to have opposite sign in eastern and western portions of the western North Pacific basin, tending to cancel over a region as large as we use here. While it might be possible to choose the averaging region so as to obtain a significant correlation, any fixed region will be problematic given the large interannual track variations.

4. Conclusions

[17] We used available data for the North Atlantic and western North Pacific to test the predictions of potential intensity (PI) theory. The prediction is that a given percentage change in the PI experienced by tropical cyclones (TCs) will on average lead to a percentage change in actual intensity that is comparable to, but somewhat smaller than that in PI if a lower bound (such as tropical storm or hurricane) is imposed on the intensity data. For the Atlantic and Pacific basins, the theoretical normalized slope is around 0.65 for hurricanes only, and 0.8 when tropical storms are included.

[18] Table 1 summarizes key statistics computed from the observations. We evaluated both PI and actual intensity at the first point of maximum intensity on each TC track, averaged the resulting PI and actual intensity values over each

year to create time series with annual resolution. In this case, the correlations between variations in PI and actual maximum intensity in both basins are significant and the slopes of the least-squares linear regressions are consistent with the predictions of PI theory, within uncertainties, in both basins.

[19] When PI is instead averaged over the nominal main development region (MDR) and over the peak season in each basin, the comparison of the resulting time series to that of average maximum actual intensity gives results which vary by basin, but are not in either case predicted by PI theory. In this case the storms' intensities are being compared to PI values which may be quite different from those which the storms actually experienced near their times of maximum intensities. Variations in mean MDR PI are smaller than those experienced by the storms, indicating that the latter are in large part induced by variations in the TC tracks, rather than by variations in PI at fixed locations. This implies that prediction of TC intensity variations induced by climate changes requires not only prediction of the PI field, but also of variations in TC tracks. This requirement may reduce the utility of PI theory for prediction of TC intensity, either on a seasonal-to-interannual or longer term basis. On the other hand, if track changes were more predictable than changes in the PI field, and PI changes experienced by storms were dominated by track changes (as is apparently the case for observed interannual variability), that dominance could actually improve the prospects for prediction. This scenario is not far-fetched, as track changes in response to some well-known modes of climate variability do have a systematic and presumably predictable component; consider the response of tracks to El Niño and La Niña events in the western North Pacific [e.g., *Chan*, 2005; *Camargo et al.*, 2007], or to the "Atlantic Meridional Mode" in the Atlantic [e.g., *Kossin and Vimont*, 2007].

Appendix: Theoretical Slopes

[20] The mean of a distribution that is uniform between two bounds, say α and β, is the average of the bounds, $(\alpha + \beta)/2$. Thus let the distribution of maximum TC intensities be uniform between a lower bound that is a fixed fraction (say γ) of PI, and the PI itself, so $\gamma P < V < P$, where V is maximum TC intensity and P is PI. Then taking the average over a large sample of storms

$$\overline{V} = \left(\frac{\gamma + 1}{2}\right)\overline{P},$$

which implies that if PI is varied by a given percentage, V will vary by the same percentage, as stated by *Emanuel* [2000]. Now if instead the lower bound is a fixed number, δ, rather than a fixed fraction of P, so $\delta < V < P$, we have

$$\overline{V} = \frac{\delta + \overline{P}}{2}.$$

Normalizing and removing the mean, we find

$$\frac{\overline{V} - \langle\overline{V}\rangle}{\langle\overline{V}\rangle} = \left(\frac{\langle\overline{P}\rangle}{\delta + \langle\overline{P}\rangle}\right)\left(\frac{\overline{P} - \langle\overline{P}\rangle}{\langle\overline{P}\rangle}\right), \tag{1}$$

where the angle brackets represent the average over the time series. Thus the normalized slope is $\langle\overline{P}\rangle/(\delta + \langle\overline{P}\rangle)$, which varies between the limits of 1 (when $\delta = 0$) and 0.5 (as $\delta \rightarrow \langle\overline{P}\rangle$).

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S. J. Camargo, International Research Institute for Climate and Society, The Earth Institute at Columbia University, Lamont Campus, 225 Monell Building, 61 Route 9W, Palisades, NY 10964, USA.

A. H. Sobel, Department of Applied Physics and Applied Mathematics and Department of Earth and Environmental Sciences, Columbia University, 500 W. 120th Street, Room 202, New York, NY 10027, USA. (ahs129@columbia.edu)

A. A. Wing, Department of Earth and Atmospheric Sciences, Cornell University, Ithaca, NY 14853, USA. (aaw28@cornell.edu)