Observational Evidence for Oceanic Controls on Hurricane Intensity

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(Manuscript received 1 April 2010, in final form 20 September 2010)

ABSTRACT

The influence of oceanic changes on tropical cyclone activity is investigated using observational estimates of sea surface temperature (SST), air–sea fluxes, and ocean subsurface thermal structure during the period 1998–2007. SST conditions are examined before, during, and after the passage of tropical cyclones, through Lagrangian composites along cyclone tracks across all ocean basins, with particular focus on the North Atlantic. The influence of translation speed is explored by separating tropical cyclones according to the translation speed divided by the Coriolis parameter. On average for tropical cyclones up to category 2, SST cooling becomes larger as cyclone intensity increases, peaking at 1.8 K in the North Atlantic. Beyond category 2 hurricanes, however, the cooling no longer follows an increasing monotonic relationship with intensity. In the North Atlantic, the cooling for stronger hurricanes decreases, while in other ocean basins the cyclone-induced cooling does not significantly differ from category 2 to category 5 tropical cyclones, with the exception of the South Pacific. Since the SST response is nonmonotonic, with stronger cyclones producing more cooling up to category 2, but producing less or approximately equal cooling for categories 3–5, the observations indicate that oceanic feedbacks can inhibit intensification of cyclones. This result implies that large-scale oceanic conditions are a control on tropical cyclone intensity, since they control oceanic sensitivity to atmospheric forcing. Ocean subsurface thermal data provide additional support for this dependence, showing weaker upper-ocean stratification for stronger tropical cyclones. Intensification is suppressed by strong ocean stratification since it favors large SST cooling, but the ability of tropical cyclones to intensify is less inhibited when stratification is weak and cyclone-induced SST cooling is small. Thus, after accounting for tropical cyclone translation speeds and latitudes, it is argued that reduced cooling under extreme tropical cyclones is the manifestation of the impact of oceanic conditions on the ability of tropical cyclones to intensify.

1. Introduction

In this paper we examine the oceanic response to North Atlantic hurricanes as a potentially important environmental control on hurricane activity. Recent work has led to a developing consensus that a number of environmental controls are important for tropical cyclone activity. Cyclone activity depends on the thermodynamic state of the local atmosphere (Emanuel 1988; Knutson et al. 2010), vertical wind shear (Frank 1977), and easterly wave activity (Thorncroft and Hodges 2001), among other factors.

Observational evidence from various case studies indicates that hurricanes cause strong vertical ocean mixing that brings cold water to the ocean surface (Price 1981; Zedler et al. 2002; D'Asaro 2003; McPhaden et al. 2009). Studies of global satellite observations have confirmed the presence of widespread cyclone-induced SST cooling and have suggested potential impacts of these oceanic changes on climate through signatures in oceanic and atmospheric memory (Hart et al. 2007) and meridional ocean heat transport (Sriver and Huber 2007; Jansen et al. 2010). Observations (Cione and Uhlhorn 2003; Kaplan and DeMaria 2003), theoretical considerations (Pasquero and Emanuel 2008), and modeling studies (Schade and Emanuel 1999; Knutson et al. 2001; Jacob and Shay 2003; Shen and Ginis 2003; Emanuel et al. 2004) suggest that cyclone-induced oceanic changes—and thus the oceanic conditions that control these changes—can

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DOI: 10.1175/2010JCLI3763.1

influence the intensification of cyclones. But when taking an average over a large number of tropical cyclones for an extended period of time, to what extent does a composite of tropical cyclones agree with results for individual case studies? Moreover, what is the observational evidence, if it exists, for a systematic role of cyclone-induced oceanic changes on cyclone intensity?

Considering the oceanic response to cyclones, a null hypothesis may be that strong tropical cyclones should lead to greater SST cooling than weaker cyclones for a given cyclone size, translation speed, and degree of ocean stratification. This expectation is formed from considering only the atmospheric forcing from tropical cyclones, since greater atmospheric forcing is exerted by stronger cyclones. With other factors being equal, stronger winds should lead to larger SST cooling, since they will cause a greater net air–sea turbulent enthalpy flux out of the ocean, and drive wind-induced vertical mixing and upwelling. An alternative hypothesis, motivated by studies that show the cyclone-induced wake as a strong control on cyclone intensity (Schade and Emanuel 1999; Bender and Ginis 2000; Knutson et al. 2001; Cione and Uhlhorn 2003; Pasquero and Emanuel 2008), would be that the intensity–wake relationship should be more subtle since a strong wake may prevent a cyclone from maintaining its intensity.

To first order we assume that changes in the distribution of tropical cyclone size, translation speed, and ocean stratification—by which we mean the thermal structure of the upper ocean—will not significantly change across different tropical cyclone categories. With this assumption, we expect to observe a proportionality between tropical cyclone intensity and SST cooling. Any deviations from a monotonic SST anomaly–intensity relationship would then indicate differences in tropical cyclone size, translation speed, or ocean thermal structure (e.g., stratification) between tropical cyclone categories. These deviations could indicate a nonlinear response of SST to tropical cyclone intensity.

To isolate variables that control the cyclone-induced SST response, our preference would be to divide tropical cyclones according to the nondimensional term V/fL, for translation speed V, Coriolis parameter f, and characteristic length scale L, if all parameters were known. From inertial oscillation theory we expect the maximum SST response to occur when $V/fL \sim 1$. Thus, areas of the ocean exposed to tropical cyclones for longer time intervals should have greater air–sea enthalpy losses and total momentum fluxes, and stronger surface cooling (Lin et al. 2009). Surface cooling will also depend on tropical cyclone size and latitude. As a result of a lack of observational data for global tropical cyclones sizes, we divide cyclones by the criteria V/f. Assuming differences in cyclone size (specifically the radius of maximum winds) are relatively small (Kimball and Mulekar 2004), we anticipate that any nonmonotonic features in the SST–intensity relationship will be due to differences in upper-ocean stratification.

We expect that any differences in upper-ocean stratification across tropical cyclone categories will cause different degrees of negative ocean feedback (Bender et al. 1993; Schade and Emanuel 1999; Bender and Ginis 2000), and that increased SST cooling may significantly reduce hurricane intensity. For a given atmospheric forcing, regions with weaker stratification will produce larger SST cooling, which could constrain the maximum intensity of hurricanes. The sensitivity of the upper ocean to atmospheric forcing would then be important; regions with shallower mixed layer and thermocline, which also tend to have lower oceanic heat content, have a stronger sensitivity to atmospheric forcing.

To explore these processes, a Lagrangian composite approach is used to analyze SST anomaly changes associated with tropical cyclones in the North Atlantic and other ocean basins for a 10-yr period (1998–2007). We outline the observational data and composite methods used in section 2. Results are presented in section 3, where we find that the SST–intensity relationship is nonmonotonic for small V/f criteria, indicating the importance of upper-ocean stratification for the strongest tropical cyclones. Discussion of this result is given in section 4, where we consider the role of large-scale ocean conditions and oceanic feedback on hurricane intensity.

2. Data and methods

a. Observations

The Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) sensor allows for daily observations of SST even under clouds (Wentz et al. 2000) and has led to new insights into air–sea interaction processes (Xie et al. 1998; Chelton et al. 2001; Lloyd and Vecchi 2010). We use observations from the TMI satellite to examine the SST response along tropical cyclone tracks during the period 1998–2007. To estimate the air–sea enthalpy flux associated with tropical cyclones, we use Yu–Weller fluxes (Yu and Weller 2007) during the period 1998–2004 [a comparison with the National Centers for Environmental Prediction (NCEP) reanalysis-2 data (Kanamitsu et al. 2002) during 1998–2007 is shown in appendix A]. For the Yu–Weller fluxes, reanalysis and satellite retrieval data are combined using a variational objective analysis technique, and sensible and latent heat fluxes are computed following bulk aerodynamic formulas following Liu et al. (1979). Anomalies of SST and air–sea enthalpy flux are computed from a monthly climatology based on 1998–2007. We also use $1° \times 1°$ monthly gridded analysis of Coupled Ocean Data Assimilation (CDA; Zhang et al. 2007) during 1998–2007 to explore the large-scale oceanic state during cyclones. The CDA consists of an ensemble filter applied to the Geophysical Fluid Dynamics Laboratory (GFDL) Coupled Climate Model version 2(CM2), with real ocean observations (e.g., XBTs, CTDs, Argo floats, etc.) assimilated into the coupled system.

As a measure of near-surface stratification, we examine the depth at which ocean temperature is 2 K below the temperature at a depth of 10 m, which we denote by T(10 m) − 2 K. We do not use SST because subsurface observations in the CDA data start at 10-m depth, and we choose a 2-K depth below the 10-m temperature value in order to focus on stratification changes large enough to be well represented in the CDA. The T(10 m) − 2-K depth is one measure of near-surface stratification, and provides a rough approximation as to the sensitivity of SST to atmospheric perturbations. Gridded daily analyses of CDA data were not available, so our results do not capture rapid time transient behavior of stratification on rapid time scales. The T(10 m) − 2-K depth for monthly climatologies averaged over June–November for the Northern Hemisphere cyclone season and December–May for the Southern Hemisphere cyclone season are shown in Figs. 1b and 1c, respectively.

We average the data from TMI–SST, Yu–Weller fluxes, and CDA using the positions of North Atlantic tropical cyclones from the International Best Track Archive for Climate Stewardship (IBTRACS) tropical cyclone dataset (Knapp et al. 2009) to generate Lagrangian composite response for 10 yr of tropical cyclones (see section 2b). The IBTRACS dataset records the 10-min maximum wind speed along with tropical cyclone track locations at 6-h intervals, which we convert to 1-min values by dividing by 0.88 (Knapp et al. 2009). This conversion allows wind speeds from IBTRACS data to be compared on the Saffir–Simpson scale, which uses a 1-min maximum wind speed definition. A plot of global tropical cyclones from the IBTRACS dataset for 1998–2007, divided by tropical cyclone category on the Saffir–Simpson scale, is shown in Fig. 1a. The analysis presented in this study was initially conducted for the North Atlantic. As an independent test, we subsequently repeat the analysis for all other ocean basins, and present these results in tandem with those for the North Atlantic. As a result of the lack of category 5 tropical cyclones outside of the North Atlantic during the period of study, we include results combining tropical cyclones in categories 4 and 5.

b. Lagrangian composite method

A Lagrangian composite analysis was performed over 2489 tropical cyclone track locations in the North Atlantic, and 10 325 tropical cyclone track locations in all other ocean basins (see Tables 1 and 2 for detailed cyclone counts). In this approach, evaluation of the SST anomaly (and other variables) was made before, during, and after a tropical cyclone passes every recorded position for all tropical cyclone tracks. The TMI–SST data were sampled at daily intervals for $1^\circ \times 1^\circ$ areas centered on the tropical cyclone track location. A composite response was calculated by averaging over all track positions for all tropical cyclones. The composite response was computed separately by category on the Saffir–Simpson scale, according to the magnitude of the 10-min maximum wind speed, and was also divided according the criteria V/f, for the tropical cyclone's translation speed divided by the Coriolis parameter $f = 2\Omega \sin\lambda$, for angular frequency Ω and latitude λ. A threshold criteria of $V/f = 1$, where 1 unit = 100 km, was used to prescribe approximately equal numbers of fast-moving (or low latitude) tropical cyclones for $V/f < 1$, and slow-moving (or high latitude) tropical cyclones for $V/f > 1$. For category 0 tropical cyclones, tropical depressions with maximum wind speeds of 17 m s^{-1} or less were excluded. To examine the mean composite response two days after the passage of the tropical cyclone, normal statistics were used to estimate the uncertainty k in the sample mean $\overline{x}$, for true mean $\mu = \overline{x} \pm k$. The uncertainty is given by $k = st_{\alpha/2}/\sqrt{n}$ where s is the sample standard deviation, n is the number of tropical cyclone track points, and $t_{\alpha/2}$ is the t statistic with significance level α, which was set to the 90% and 95% confidence intervals.

The composite response was further divided in some analyses to examine subregions within ocean basins. In the North Atlantic, we examined the Greater Antilles and western Caribbean region (15°–25°N, 90°–60°W) to test the dependence of our analyses on geographic location. This region was chosen because of its range of tropical cyclone intensities (see Fig. 1a). Tropical cyclones were also divided into two classifications: intensification or decay. The intensity tendency $\partial I/\partial t$ over a 36-h period (containing 6 data points) starting at day 0 was used to define intensifying or decaying tropical cyclones, such that $\partial I/\partial t > 0$ for intensifying tropical cyclones, and $\partial I/\partial t < 0$ for decaying tropical cyclones. Tropical cyclones reaching landfall within the 36-h period were excluded from either classification. In addition to examining the SST cooling for a given intensification or decay phase, tropical cyclones were divided into different bands of SST cooling, at 1-K intervals, to examine the fraction of tropical cyclones that intensify or decay for a given band of cooling.

3. Results

For all tropical cyclone categories, across all ocean basins, tropical cyclones on average experience larger cooling

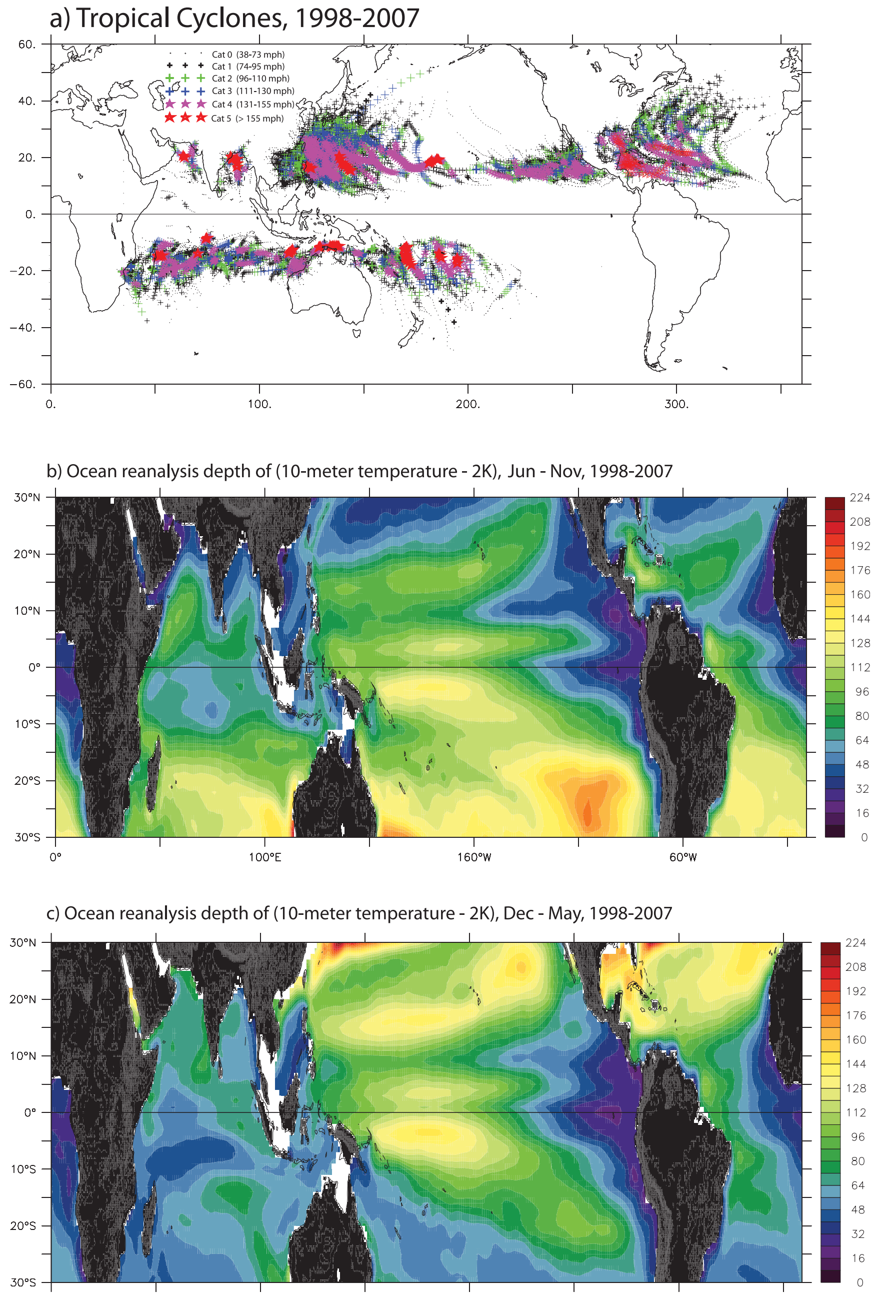

FIG. 1. (a) Global tropical cyclones during the period 1998–2007, from IBTRAC data. Colors indicate the tropical cyclone category on the Saffir–Simpson scale for each individual track position. (b) Ocean reanalysis (CDA) data, showing monthly climatology during June–November for the $T(10\text{m}) - 2$-K depth at 30°S–30°N. (c) Ocean reanalysis (CDA) data, showing monthly climatology during December–May for the $T(10\text{m}) - 2$-K depth at 30°S–30°N.

for smaller values of V/f (Fig. 2). The dashed line at $V/f = 1$ in Fig. 2 indicates the division between slow-moving (or high latitude) and fast-moving (or low latitude) tropical cyclones, and approximately divides storms by equal numbers in the North Atlantic. Because we are interested in the influence of oceanic feedback on tropical cyclones, we focus on the criteria $V/f < 1$, for which the average SST cooling is greater.

On average, North Atlantic tropical cyclones are preceded by a positive SST anomaly on the order of 0.2 K, which we set as a reference value from 12 to 2 days before a cyclone reaches a position (Fig. 3a). As the cyclone

TABLE 1. Tropical cyclone counts for each ocean basin divided by intensity on the Saffir–Simpson scale. The IBTRACS dataset during 1998–2007 is used and each data point represents an observation at 6-h intervals. Data are shown for all translation speeds and latitudes (all V/f values).

Ocean basin	North Atlantic	All basins except North Atlantic	West Pacific	East Pacific	South Pacific	South Indian	North Indian
Total	2489	10 325	4110	2325	1082	2399	409
Category 0	1433	5928	2033	1440	685	1454	316
Category 1	493	1847	825	392	171	422	37
Category 2	211	1029	526	206	86	200	11
Category 3	149	934	499	150	65	201	19
Category 4	144	525	211	123	55	116	20
Category 5	59	62	16	14	20	6	6

reaches a position, there is a rapid surface cooling, with the SST anomaly reaching a minimum two days after the cyclone passes (day 0). Though the SST anomaly returns to climatology at approximately day +20, SST does not return to the precyclone positive SST anomaly values even after 50 days. Accordingly, the passage of a tropical cyclone results in an average postcyclone SST decrease of approximately 0.2 K from precyclone SST. Similar behavior is observed in other ocean basins (Fig. 3c).

Figure 4 confirms that the amplitude of the cooling is dependent on the intensity of the tropical cyclone and its value of V/f. For tropical cyclones with $V/f < 1$ (slow moving or high latitude) the oceanic response shows a stronger dependence on hurricane intensity, likely due to longer exposure times between hurricanes and specific points on the ocean surface. As may be expected from our initial null hypothesis, when tropical cyclone intensity increases from category 0 to category 2 the maximum SST cooling response for a $1° \times 1°$ Lagrangian composite intensifies monotonically (Fig. 4a). However, from category 2 to category 5 the SST response has a statistically significant decrease in magnitude. This result is repeated in a smaller region focused on the Greater Antilles and western Caribbean (Fig. 5c), excluding geographical dependence as a possible explanation for the observed SST–intensity response. When combining all other ocean basins there is no significant difference in the mean SST cooling between category 2 and category 5 (see Fig. 4c and appendix B). In addition, for tropical cyclones with $V/f < 1$ the SST cooling is larger, and the nonmonotonic dependence of SST response on hurricane intensity is more evident when compared to tropical cyclones for which $V/f > 1$ (Fig. 4). Finally, when comparing intensifying or decaying tropical cyclones, it is found that decaying tropical cyclones are associated with larger SST cooling, and have an SST–intensity relationship that is more markedly nonmonotonic (Fig. 6). Division of tropical cyclones into different bands of cyclone-induced SST cooling shows that tropical cyclones that have larger SST cooling are more likely to decay (Fig. 7). For bands of smaller cyclone-induced SST cooling, tropical cyclones are more likely to intensify (Figs. 7a,c), while category 4 and 5 tropical cyclones have a more equal chance of intensification or decay (Figs. 7b,d). The interpretation of these results is discussed in section 4.

We also find that the nonmonotonic SST anomaly–intensity relationship is enhanced for cyclones that cause the largest SST cooling: when dividing the cyclone-induced SST response into quantiles, we find that the upper quantiles (largest cooling) show an SST–intensity response that is more markedly nonmonotonic (Figs. 3b,d). For lower quantiles (smaller cooling) the SST–intensity relationship is still somewhat nonmonotonic; however, the range of SST cooling for the lower quantiles over different hurricane categories is smaller. The reduced mean SST cooling from stronger tropical cyclones across all quantiles indicates oceanic conditions that are relatively insensitive to atmospheric perturbations.

TABLE 2. As in Table 1, but for slow-moving (or high latitude) tropical cyclones with $V/f < 1$.

Ocean basin	North Atlantic	All basins except North Atlantic	West Pacific	East Pacific	South Pacific	South Indian	North Indian
Total	1408	4955	1900	1054	552	1226	223
Category 0	822	2884	895	698	393	730	168
Category 1	282	893	385	189	81	220	18
Category 2	117	516	284	86	39	101	6
Category 3	78	416	235	46	12	111	12
Category 4	72	223	96	29	18	64	16
Category 5	37	23	5	6	9	0	3

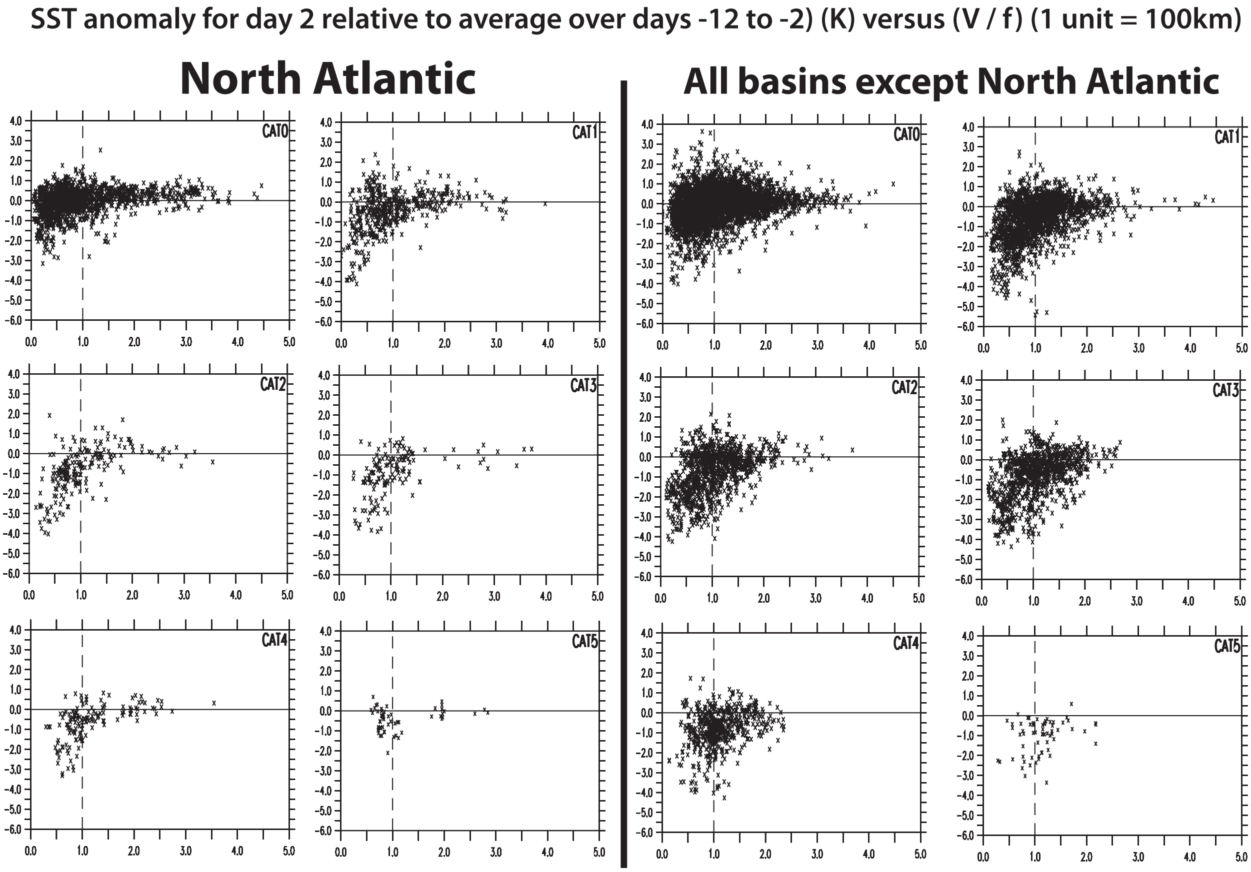

FIG. 2. The 1° × 1° SST anomaly response for day 2 minus the average over days −12 to −2 is plotted on the y axis, against translation speed (m s^{-1}) divided by the Coriolis parameter (s^{-1}), V/f, on the x axis, with 1 unit = 100 km. The six plots left of the central dividing line are for the North Atlantic, with each plot showing results for different cyclone categories on the Saffir–Simpson scale, starting with category 0 on the top left and category 5 on the bottom right. The six plots to the right of the central line are for all ocean basins except the North Atlantic for different cyclone categories. The dashed line at $V/f = 1$ indicates the division used in this study between slow-moving (or high latitude) and fast-moving (or low latitude) tropical cyclones.

Consistent with our initial expectation, as tropical cyclone intensity increases the net air–sea enthalpy flux becomes larger (Figs. 8a,c). While momentum and air–sea enthalpy fluxes increase monotonically for tropical cyclones, the SST response is nonmonotonic, becoming smaller for the strongest cyclones (Figs. 4a,c). This result confirms that the observed behavior of SST, momentum, and air–sea enthalpy fluxes cannot be explained by atmospheric forcing alone. As predicted in previous studies (e.g., Schade and Emanuel 1999; Bender and Ginis 2000; Shen and Ginis 2003), there is a role for ocean feedback in the development of hurricanes to their maximum intensity, which involves ocean dynamical changes, and these feedbacks are responsible for the nonmonotonic SST–intensity relationship. By isolating tropical cyclones according to the V/f criteria, the nonmonotonicity in the SST anomaly–intensity relationship must be due to differences in upper-ocean stratification for different tropical cyclone categories.

Our interpretation of the SST–intensity relationship is that the most intense tropical cyclones (categories 4–5) can only develop when there is no more than a modest SST cooling. The state of the upper ocean can modulate the thermal sensitivity of SST to atmospheric perturbations (Knutson et al. 2001), with stronger near-surface stratification increasing the sensitivity of SST. Therefore, we hypothesize that this relationship is indicative of a role for the ocean in modulating intensity, because of (i) feedbacks of cooling on cyclone intensity, (ii) the role of oceanic mixing and upwelling in the cyclone-induced cooling, and (iii) the role of upper-ocean thermal stratification in modulating the sensitivity of SST to atmospheric perturbations.

This conjecture is supported by CDA data, for which the $T(10\ \text{m}) - 2$-K depth is greater for the most intense tropical cyclones. This result is consistent with large-scale oceanic conditions conducive to a lower ocean sensitivity to atmospheric forcing, for which ocean stratification is weaker and the mean SST response is reduced (Fig. 7b).

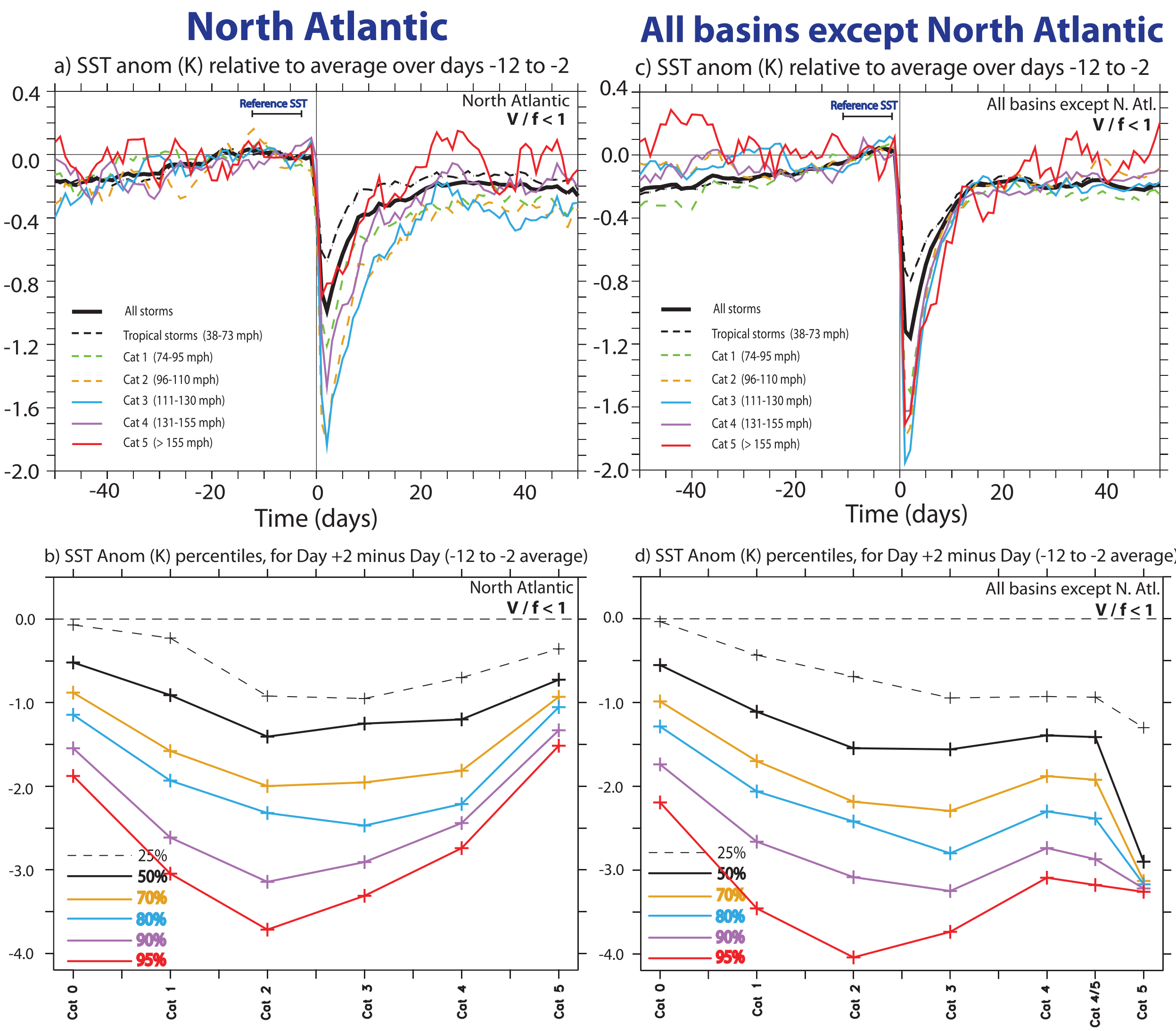

FIG. 3. (a) Lagrangian composite SST anomaly response relative to the average over days −12 to −2, for the North Atlantic. The composite SST anomaly is averaged over a 1° × 1° region centered over tropical cyclone track positions. Negative values on the time axis indicate days before a tropical cyclone passes a track position. Day 0 is when the tropical cyclone reaches the track position, and positive values indicate the track position response after the tropical cyclone has passed. The *y* axis indicates the SST anomaly (K). Lines are divided into mean response and different categories on the Saffir–Simpson scale, which are indicated by the color key. (b) Lagrangian composite SST anomaly for different percentiles, with higher percentile values for larger SST anomalies. The *y* axis shows the magnitude of the SST anomaly response (K), while the *x* axis indicates the tropical cyclone category. (c),(d) As in (a),(b), but they contain results for all ocean basins except the North Atlantic. In addition, (d) also shows results for a combination of categories 4 and 5. All panels are presented for tropical cyclones with $V/f < 1$.

4. Discussion

Tropical cyclones with $V/f < 1$ tend on average to produce greater SST cooling (Fig. 2). For tropical cyclones with $V/f > 1$, which have a reduced mean SST response, oceanic feedback is weaker, and atmospheric forcing tends to dominate the SST response (Figs. 4b,d). We focus on $V/f < 1$ in this paper because we are interested in the effects of oceanic feedback on hurricane intensity.

As tropical cyclone intensity increases, the air–sea enthalpy flux out of the ocean increases monotonically (Figs. 8a,c). If the cyclone-induced SST response is forced by the atmosphere—the ocean feedback onto the cyclone itself is not important—we expect that cyclone-induced SST responses should increase monotonically with cyclone intensity, since momentum (and induced vertical mixing and upwelling) and air–sea enthalpy fluxes increase with cyclone intensity.

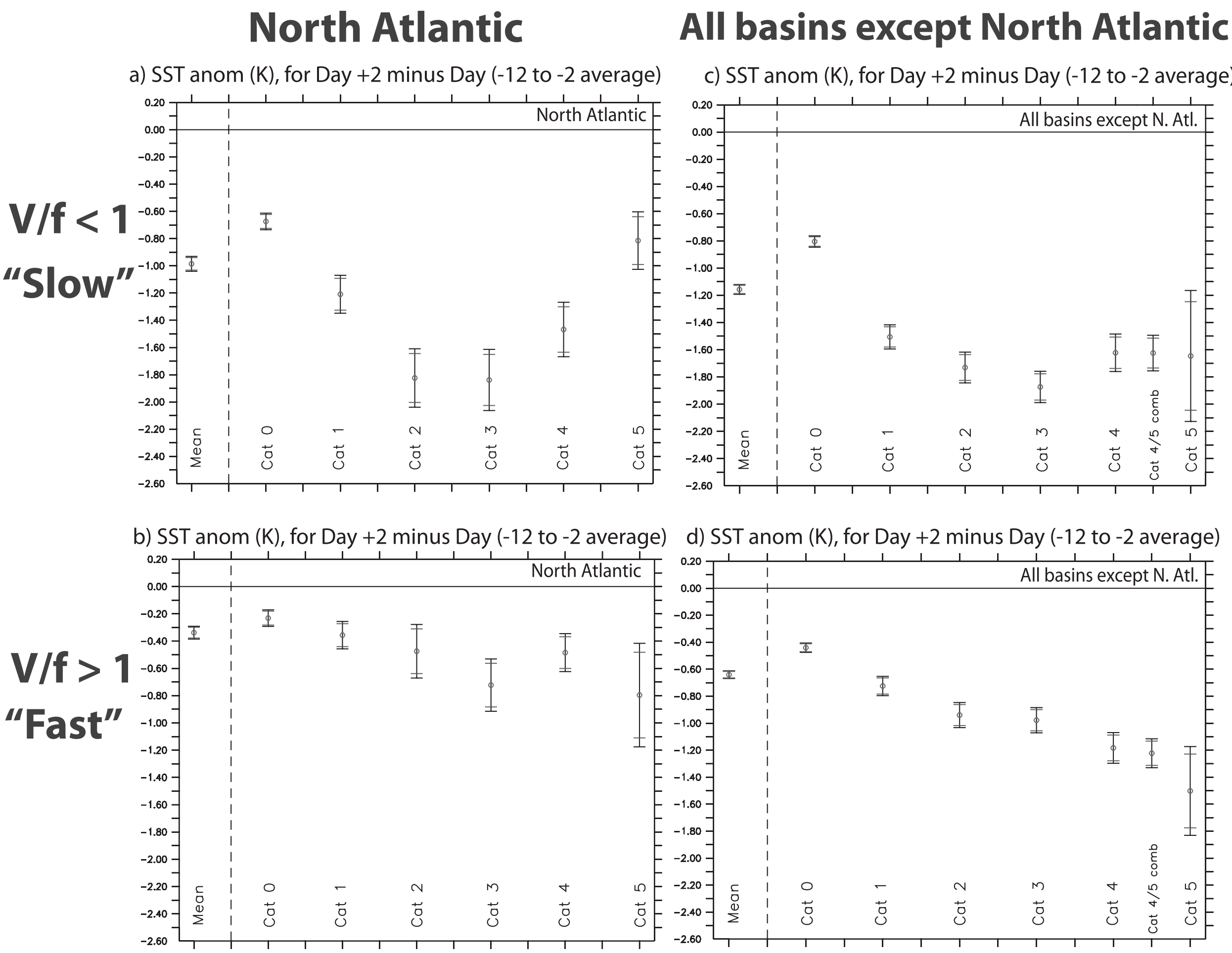

FIG. 4. The Lagrangian composite for the SST anomaly response on day +2 minus the day −12 to −2 average. The *y*-axis values indicate the mean SST response, with error bars showing the 90% and 95% significance levels for errors in the mean. (a) Slow-moving (or low latitude) tropical cyclones with $V/f < 1$ and (b) fast-moving (or high latitude) cyclones with $V/f > 1$ for the North Atlantic. The data point on the far left indicates the mean composite response for all tropical cyclones, while other points mark different categories on the Saffir–Simpson scale. (c) Slow (or low latitude) and (d) fast (or high latitude) tropical cyclones in all other ocean basins.

However, we find that the SST response is nonmonotonic, with stronger cyclones producing more cooling up to category 2 but producing less or approximately equal cooling for categories 3–5 tropical cyclones (Figs. 4a,c). The nonmonotonic SST response is particularly pronounced in the North Atlantic. For ocean basins outside the North Atlantic, the SST cooling from category 2 to category 5 tropical cyclones does not change significantly, with the exception of the South Pacific (see Fig. 4c and appendix B). This observation may in part be due to a lack of category 5 tropical cyclones outside the North Atlantic over the period of study. Other possible explanations for the ‘‘saturation’’ of SST cooling for categories 2–5 tropical cyclones outside the Atlantic merit further investigation. For example, it could be that increased ocean mixing does not bring colder water to the surface after category 2. Alternatively, potential intensity or other atmospheric constituents (e.g., vertical wind shear) in ocean basins outside the North Atlantic could be a more dominant factor for intensification, with a lesser role for oceanic controls. The SST–intensity pattern is most apparent for $V/f < 1$ since slower-moving cyclones (and cyclones at higher latitudes) induce greater SST cooling due to enhanced vertical mixing and upwelling and greater air–sea enthalpy flux losses. Additionally, slower-moving tropical cyclones have longer exposure to cyclone-induced cooling.

The nonmonotonic SST–intensity result of Fig. 4a may at first appear counterintuitive. However, this result is the type of relationship one would expect if cyclone-induced wakes impacted the cyclone itself, as has been previously suggested by modeling studies (Bender and Ginis 2000; Knutson et al. 2001; Shen and Ginis 2003). Thus, atmospheric forcing of the ocean alone is not adequate

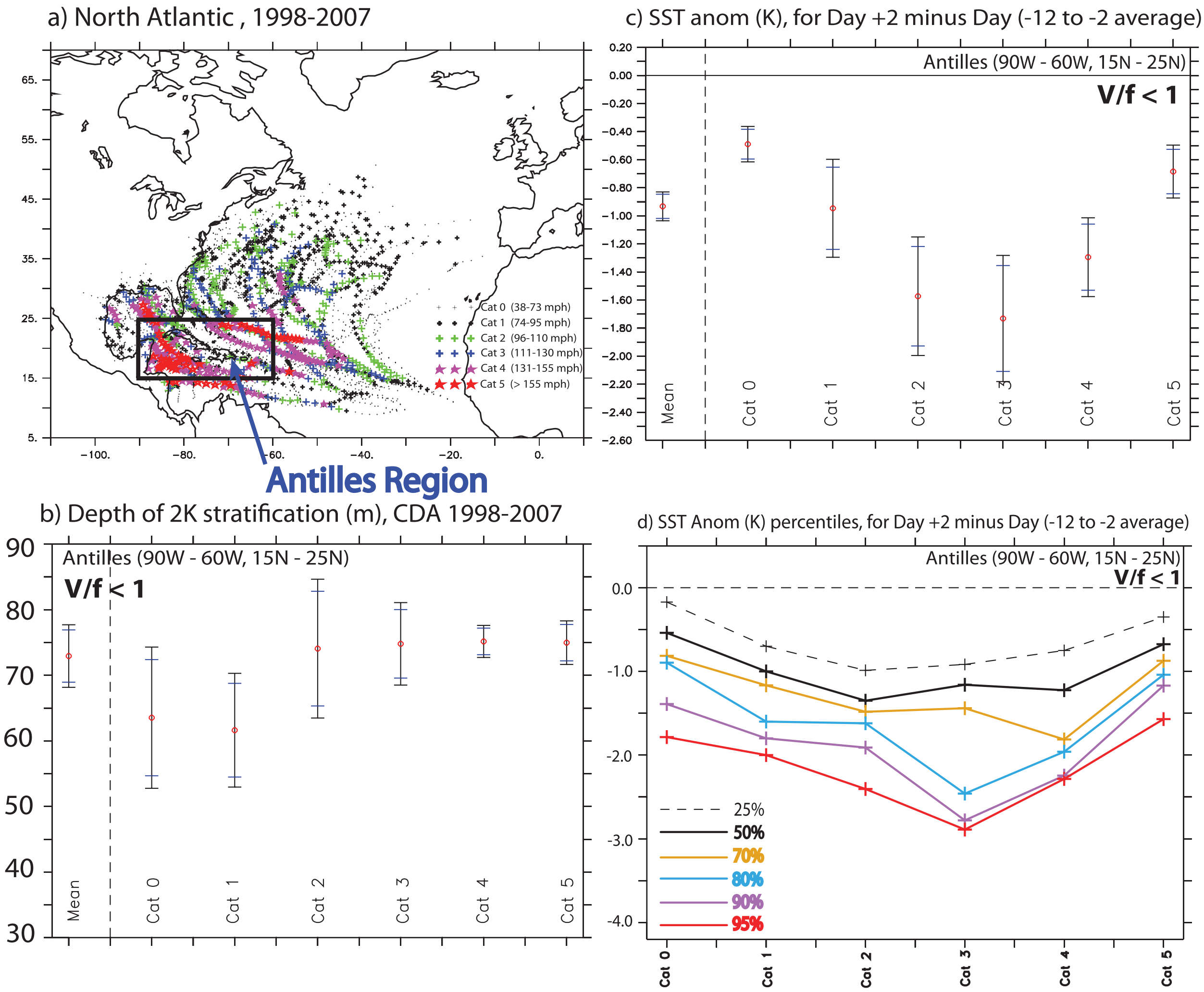

FIG. 5. (a) As in Fig. 1a, but highlighting Greater Antilles region (15°–25°N, 90°–60°W). (b) Ocean reanalysis (CDA) depth for T(10m) − 2-K stratification, for the Greater Antilles region. (c) As in Fig. 3a for Lagrangian composite SST anomaly (K), but for the Greater Antilles region. (d) As in Fig. 3b, but showing SST cooling percentiles for the Greater Antilles region.

to account for the observed relationship between SST cooling and cyclone intensity. In turn, this reasoning leads to the conclusion that upper-ocean stratification and the depth of the warm layer are important, and that if pronounced SST cooling is present tropical cyclones will not as readily intensify to categories 4 or 5. Further supporting this claim, we find that for tropical cyclones that induce the largest (upper quantile) SST cooling, the nonmonotonicity in the SST–intensity pattern becomes more evident (Fig. 3b). In addition, the SST–intensity response for intensifying and decaying tropical cyclones indicates that decaying tropical cyclones have larger SST cooling (see Fig. 6). The larger (smaller) SST cooling and more markedly nonmonotonic (monotonic) SST–intensity response indicate that decaying (intensifying) tropical cyclones tend to occupy the upper (lower) quantiles of SST cooling in Fig. 3b.

Of greater relevance to prediction, however, may be the reverse question: for a given magnitude of SST cooling, is the tropical cyclone more likely to intensify or decay? We find that when dividing tropical cyclones into 1-K bands of SST cooling, tropical cyclones with greater cooling are more likely to decay (Fig. 7), supporting a significant role for negative oceanic feedback on intensity. Thus, the enhanced nonmonotonicity in the SST–intensity pattern observed for $V/f < 1$, decaying tropical cyclones, and the upper quantiles of cyclone-induced cooling is consistent with an enhanced role for ocean feedback. This result holds true after accounting for geographic location, specifically for the Greater Antilles and western Caribbean

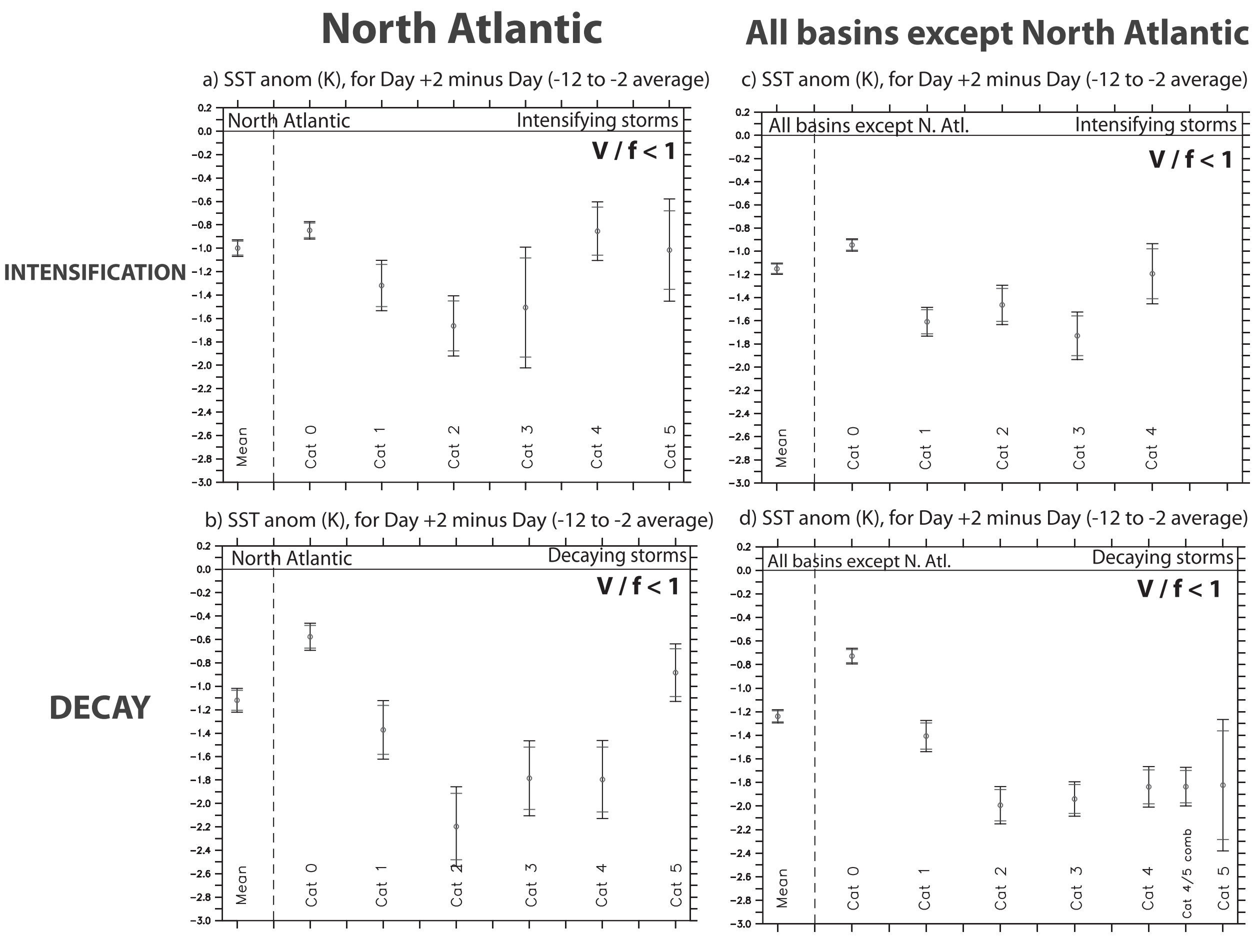

FIG. 6. (a) SST anomaly response for day +2 relative to the average over days −12 to −2, for the North Atlantic, for intensifying tropical cyclones ($\Delta I > 0$) for which $V/f < 1$. The bars mark 90% and 95% confidence intervals on the composite mean values. (b) As in (a), but for decaying tropical cyclones ($\Delta I < 0$). (c),(d) As in (a),(b), but for all basins except the North Atlantic. Results for categories 4 and 5 combined are also presented.

region (Fig. 5d), and for ocean basins outside the North Atlantic.

Given that ocean feedback is important for cyclone intensity, we are led to the conclusion that the strongest cyclones will most frequently develop in areas where ocean cooling is inhibited by large-scale oceanic conditions. We explore ocean reanalysis (CDA) data during 1998–2007 and find that, indeed, category 4 and 5 cyclones form where there is a deeper near surface warm layer [T(10 m) − 2 K], and the mixed layer depth is deep enough to inhibit strong cooling (Figs. 8b,d). As a crude estimate, from category 1 to 5 tropical cyclones the T(10 m) − 2-K depth increases by approximately 8.9 m per category in the North Atlantic, and approximately 3.2 m per category in other basins (Figs. 8b,d). The larger gradient of T(10 m) − 2-K depth versus intensity in the North Atlantic is consistent with the greater nonmonotonicity of the SST–intensity response (Fig. 4a), as SST cooling becomes more significantly inhibited for the strongest tropical cyclones. An increase in the T(10 m) − 2-K depth is also observed in the Greater Antilles and western Caribbean region (Fig. 5b).

These results show that the T(10 m) − 2-K metric is a useful indicator for hurricane intensity. While other metrics such as the depth of the 26°C isotherm are often used, the advantage of the T(10 m) − 2-K metric is that it accounts for the possibility of a 2-K SST cooling causing a reduction in hurricane intensity, even though the absolute value of SST may remain above 26°C. This perspective is reinforced by Schade and Emanuel (1999), who find that the cyclone-induced SST cooling, and hence the upper-ocean thermal structure, is more important than the final absolute value of SST. Further study should identify the causes for differences in the gradient of T(10 m) − 2-K depth versus tropical cyclone intensity, and investigate the relationship between SST,

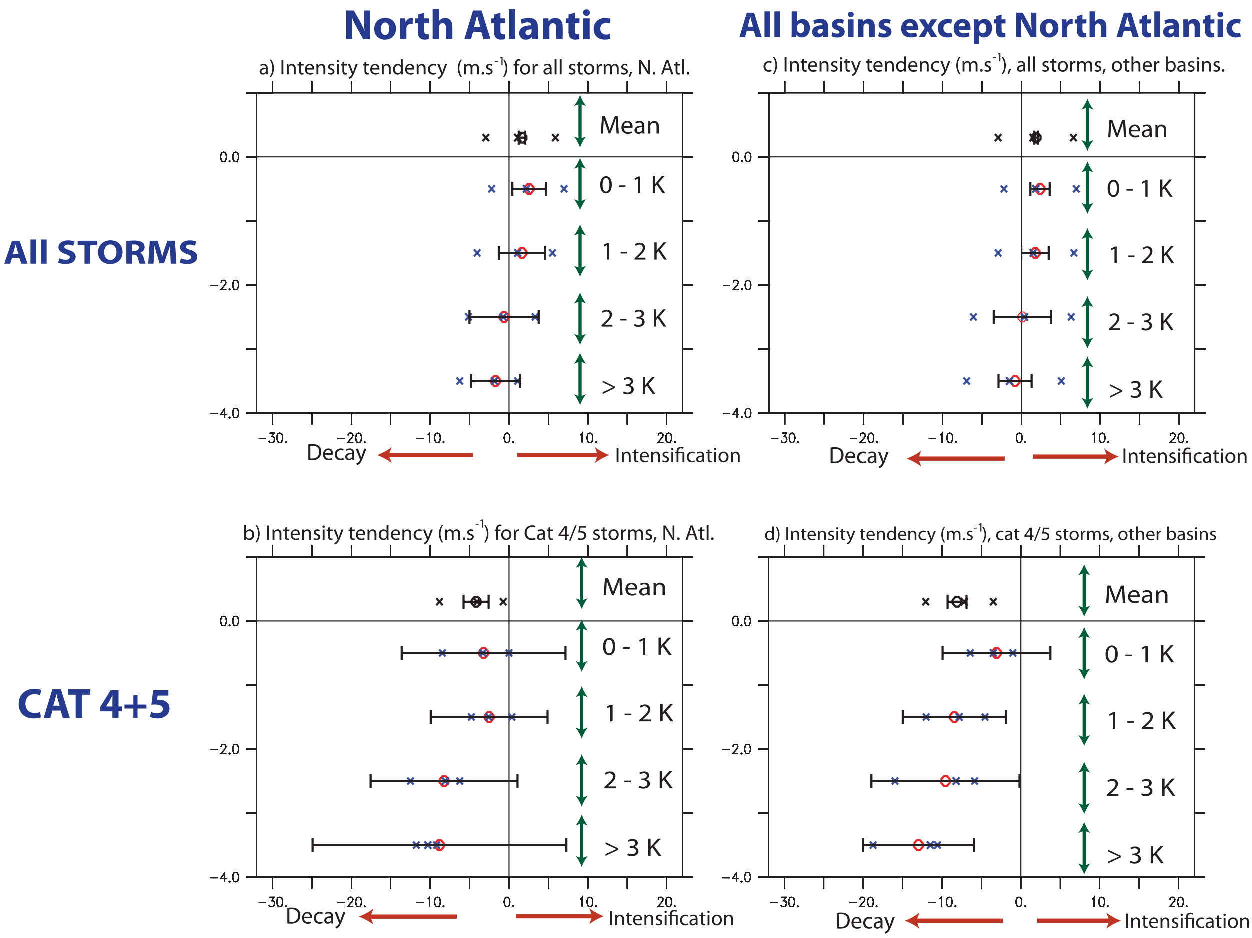

FIG. 7. (a) $\partial I/\partial t$ (m s^{-1}) for all tropical cyclones in the North Atlantic, for different bands of SST cooling. For each SST cooling band (0–1, 1–2, 2–3, and greater than 3 K) the mean $\partial I/\partial t$ is indicated by a red circle. Bars show the 95% confidence in the mean, and the 25th, 50th, and 75th percentiles are marked by blue crosses. The mean, uncertainty, and percentiles for all tropical cyclones across all SST bands are shown in black at the top. (b) As in (a), but for category 4 and 5 tropical cyclones combined. (c),(d) As in (a),(b), but for all ocean basins apart from the North Atlantic.

tropical cyclone intensity, and upper-ocean stratification when comparing individual ocean basins. We also note that our CDA data only accounts for aspects of the upper-ocean stratification that change on time scales of months and are resolved by the 1° × 1° resolution of the model (Zhang et al. 2007); transient, short time-scale features such as warm core rings, which impact intensification, are not as well captured.

We confirm through observations that the characteristic relationship between SST, momentum and enthalpy fluxes, and ocean stratification to be an indication that large-scale ocean conditions can impact hurricane intensity, and that SST changes are a feedback on intensification. Given favorable environmental conditions for hurricane development, intensification of cyclones to categories 4 and 5 will occur when upper-ocean heat content is large enough, so that cyclone-induced SST cooling is limited. In contrast, for cases in which tropical cyclones pass areas of lower-ocean heat content with shallower thermocline and mixed layer, rapid SST cooling will lead to strong negative ocean feedback, inhibiting the development of hurricanes.

The findings of this paper are in agreement with modeling studies of individual tropical cyclones (Bender et al. 1993; Bender and Ginis 2000; Jacob and Shay 2003) and observations of North Atlantic hurricanes (Cione and Uhlhorn 2003; Kaplan and DeMaria 2003) that indicate the importance of oceanic structure in controlling the response of hurricanes. Further support for oceanic control on hurricane intensity is manifest in observations of hurricane intensification over the passage of a warm loop current or warm core ring for which there is an increase in the ocean to atmosphere air–sea enthalpy flux (Bender and Ginis 2000; Hong et al. 2000; Jacob and Shay 2003;

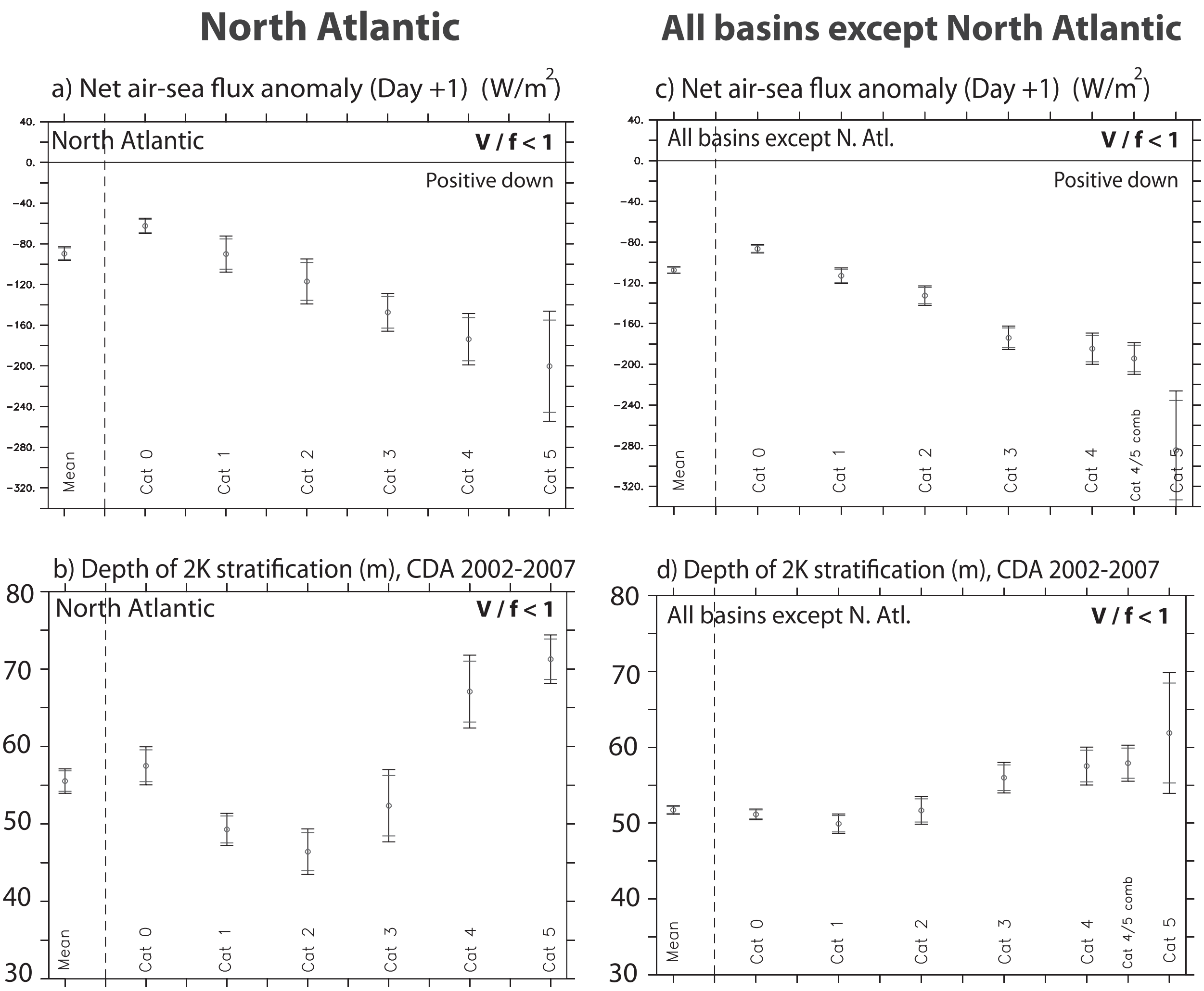

FIG. 8. (a) Lagrangian composite air–sea enthalpy flux anomaly (W m^{-2}) on day +1 for 2.5° × 2° average over tropical cyclone track positions in the North Atlantic during the period 1998–2004 using Yu–Weller fluxes. (b) Lagrangian composite of the T(10m) − 2-K stratification depth in the North Atlantic for monthly averaged 1° × 1° CDA data during 2002–07. The T(10m) − 2-K depth is divided by category on the Saffir–Simpson scale and is shown for $V/f < 1$. (c),(d) As in (a),(b), but for all basins excluding the North Atlantic and also show results for categories 4 and 5 combined. The bars mark 90% and 95% confidence intervals on the composite mean values.

Lin et al. 2005). These findings are consistent with improvements in hurricane forecasting, resulting from the inclusion of ocean coupling and estimation of oceanic heat content (DeMaria et al. 2005; Bender et al. 2007; Mainelli et al. 2008). The results presented in this paper, showing large-scale observational evidence for the importance of oceanic conditions, are consistent with these past studies. By compositing over a large number of tropical cyclones, across all ocean basins, and stratifying the TMI–SST response by tropical cyclone intensity on the Saffir–Simpson scale, we hope to reinforce past work indicating the importance of oceanic controls on hurricane intensity.

It should be noted that the influence of different tropical cyclone sizes on SST cooling has not directly been addressed in this study. Results from Kimball and Mulekar (2004) suggest that in the North Atlantic the average radius of maximum winds does not change significantly between hurricane categories. There is, though, a larger change in the average radius of 17 m s^{-1} winds, which is expected for stronger hurricanes. Further work is needed to quantify the relationship between tropical cyclone size and the magnitude of SST cooling, especially in ocean basins outside the North Atlantic.

Because the TMI–SST data is poor at recording measurements when heavy precipitation is present, we assume that the maximum SST cooling on day +2 is a good indicator of the SST cooling under a tropical cyclone when it passes a location on day 0. This assumption merits further study from in situ SST measurements. However,

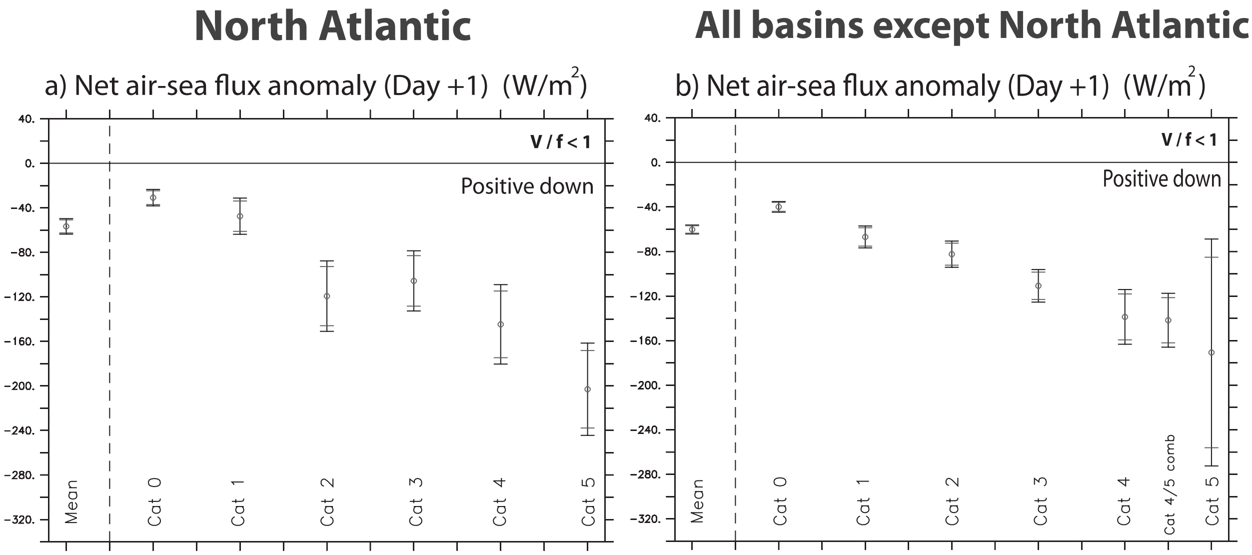

FIG. A1. (a) Lagrangian composite air–sea enthalpy flux anomaly (W m^{-2}) on day +1 for 2.5° × 2° average over tropical cyclone track positions in the North Atlantic during the period 1998–2007 using NCEP reanalysis-2 fluxes. (b) As in (a), but for all basins excluding the North Atlantic and also show results for categories 4 and 5 combined.

the result that SST cooling decreases in monotonic fashion from days +2 to +20 (Figs. 3a,c) gives support that the SST cooling on day +2 is a reasonable indication of the SST cooling on day 0, for a composite over many tropical cyclones. The nonmonotonic SST–intensity response is also reproduced for SST cooling averaged from days 0 to +5 (not shown). Furthermore, the day +2 SST cooling directly affects slow-moving tropical cyclones because, for a cyclone moving in a straight line, SST cooling at the front of the cyclone will affect the back of the cyclone 2 days later (e.g., a cyclone with a diameter of 400 km will traverse its diameter in 2 days if it travels at 2.3 m s^{-1}).

The residual SST cooling of 0.2 K after the tropical cyclone passage (Figs. 3a,c), which removes a warm SST that preceded the storm and restores climatology, prompts a question for further study: do tropical cyclones act to remove warm SST anomalies from the ocean and restore climatology, thus providing a thermostat on climate? The result is consistent with ideas supporting atmospheric and oceanic memory of tropical cyclones in Hart et al. (2007). However, using a Reynolds–optimal interpolation (OI) SST ¼° daily product to perform a similar Lagrangian composite during the period 1981–2005, Hart et al. (2007) do not see a residual SST cooling from prestorm to poststorm SST conditions. Whether this discrepancy is a result of differences between TMI–SST and Reynolds–SST measurements, or is a reflection of different temporal sampling, merits further investigation.

It appears that, for a composite over a large number of tropical cyclones, large-scale upper-ocean thermal stratification is an important factor for hurricane intensity, and that ocean dynamical changes can affect hurricane development. Consistent with past observational work (Zedler et al. 2002; D'Asaro 2003; Cione and Uhlhorn 2003; Kaplan and DeMaria 2003) we speculate that this oceanic dependence, in addition to the influence of changes in atmospheric state, may have consequences for the response of the most intense cyclones to changing climate conditions. For example, under anthropogenic greenhouse warming, radiative heating of the oceans will, in the tropical mean, increase near-ocean stratification (though regional patterns may vary; e.g., Vecchi and Soden 2007). Increased stratification would lead to larger SST sensitivity, and act to inhibit intensification, but the amplitude of this effect must be quantified. The results of Knutson et al. (2001), who examine CO_2-induced ocean stratification on hurricane intensity in an idealized framework for the Second Coupled Model Intercomparison Project (CMIP2), suggest that this ocean–thermodynamic effect on hurricane intensity is relatively minor compared to the impact of CO_2-induced warming on intensity. It would be valuable to explore the extent to which the results of Knutson et al. (2001) hold for climate projections with newer global climate models. Yet, at this stage, there is no reason to dismiss the dominance of atmospheric and SST changes on the CO_2 response of intensity presented by Knutson et al. (2001). Furthermore, the tropical Atlantic exhibits changes in its near-surface thermal structure on interannual time scales that may be relevant to intensification (e.g., Doi et al. 2010). Accordingly, we must understand the oceanic role on cyclone

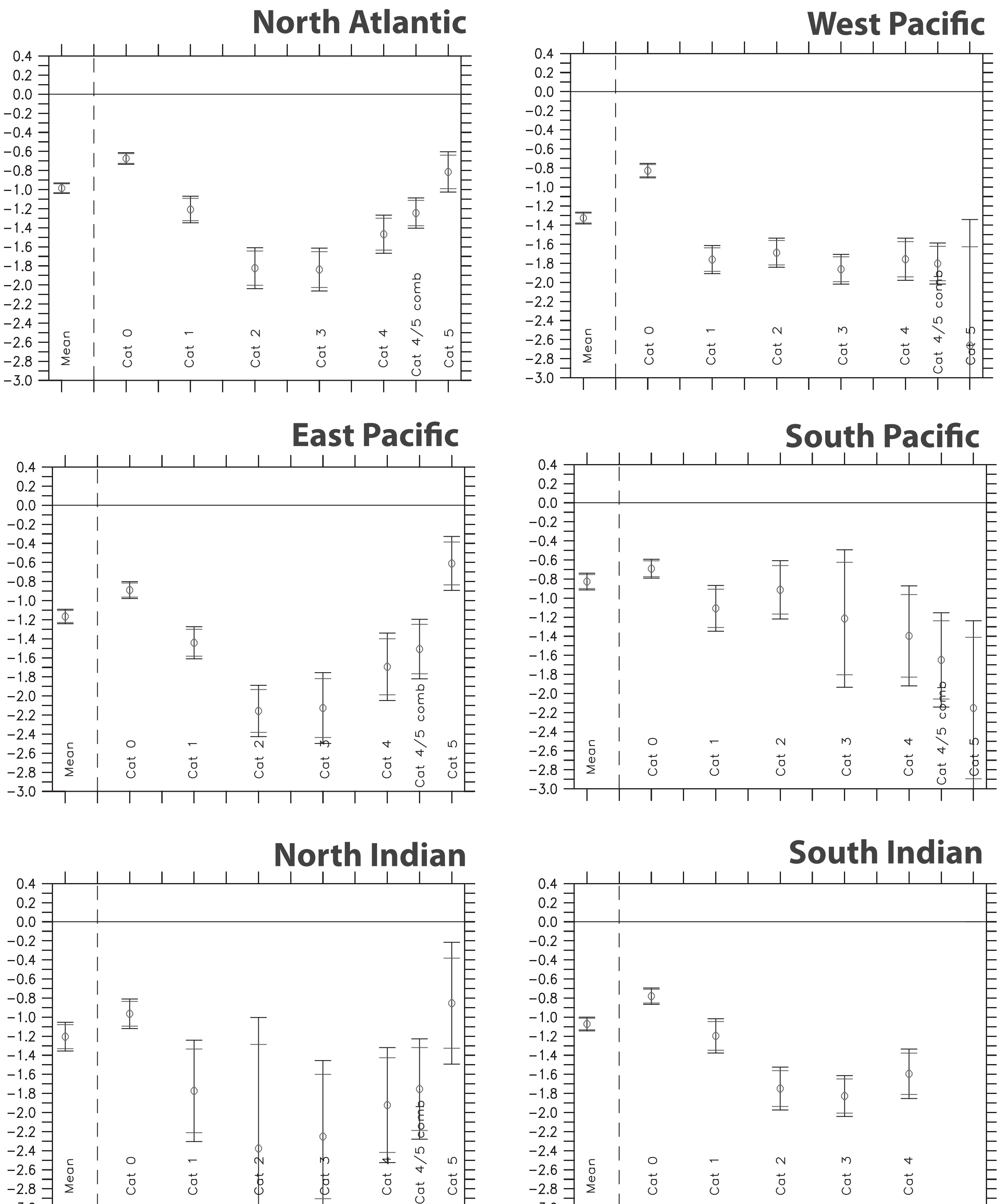

FIG. B1. Lagrangian composite anomaly for individual basins showing SST cooling for tropical cyclones that satisfy $V/f < 1$. Bars mark 90% and 95% confidence intervals on the composite mean values.

intensity, and its influence and magnitude relative to changes in atmospheric conditions. Predictions of hurricane intensity in regional and global climate models may benefit from ocean initialization, and predictions of hurricane intensity on interannual, decadal, and centennial time scales should account for changes in ocean stratification.

Acknowledgments. The authors thank Anand Gnanadesikan, Gabriel Lau, Leo Donner, Timothy Marchok, and Thomas Knutson for helpful comments and assistance during the course of this work. This report was prepared by Ian Lloyd under Award NA08OAR4320752 from the National Oceanic and Atmospheric Administration, U.S. Department of Commerce. The statements, findings, conclusions, and recommendations are those of the authors and do not necessarily reflect the views of the National Oceanic and Atmospheric Administration or the U.S. Department of Commerce.

APPENDIX A

NCEP Air–Sea Flux Analysis

To test the Lagrangian composite results from Yu–Weller fluxes, the composite analysis was repeated with NCEP reanalysis-2 fluxes (Kanamitsu et al. 2002) during the period 1998–2007 (Fig. A1). Both the Yu–Weller fluxes and NCEP reanalysis-2 fluxes indicate a monotonic increase in net air–sea enthalpy fluxes out of the ocean as a function of tropical cyclone category.

APPENDIX B

Lagrangian Composite Analysis for Individual Ocean Basins

Lagrangian composite SST anomalies for individual ocean basins are shown in Fig. B1. For all basins except the South Pacific, stronger tropical cyclones (categories 4 and 5) tend to have small or approximately equal SST cooling compared to category 2 tropical cyclones.

REFERENCES

Bender, M. A., and I. Ginis, 2000: Real-case simulations of hurricane–ocean interaction using a high-resolution coupled model: Effects on hurricane intensity. *Mon. Wea. Rev.,* **128,** 917–946.

——, ——, and Y. Kurihara, 1993: Numerical simulations of tropical cyclone–ocean interaction with a high-resolution coupled model. *J. Geophys. Res.,* **98,** 23 245–23 263.

——, ——, R. E. Tuleya, B. Thomas, and T. Marchok, 2007: The operational GFDL coupled hurricane-ocean prediction system and a summary of its performance. *Mon. Wea. Rev.,* **135,** 3965–3989.

Chelton, D. B., S. K. Esbensen, M. G. Schlax, N. Thum, and M. H. Freilich, 2001: Observations of coupling between surface wind stress and sea surface temperature in the eastern tropical Pacific. *J. Climate,* **14,** 1479–1498.

Cione, J. J., and E. W. Uhlhorn, 2003: Sea surface temperature variability in hurricanes: Implications with respect to intensity change. *Mon. Wea. Rev.,* **131,** 1783–1796.

D'Asaro, E., 2003: The ocean boundary layer below Hurricane Dennis. *J. Phys. Oceanogr.,* **33,** 561–579.

DeMaria, M., M. Mainelli, L. K. Shay, J. A. Knaff, and J. Kaplan, 2005: Further improvements to the Statistical Hurricane Intensity Prediction Scheme (SHIPS). *Wea. Forecasting,* **20,** 531–543.

Doi, T., T. Tozuka, and T. Yamagata, 2010: The Atlantic meridional mode and its coupled variability with the Guinea Dome. *J. Climate,* **23,** 455–475.

Emanuel, K. A., 1988: The maximum intensity of hurricanes. *J. Atmos. Sci.,* **45,** 1143–1155.

——, C. DesAutels, C. Holloway, and R. Korty, 2004: Environmental control of tropical cyclone intensity. *J. Atmos. Sci.,* **61,** 843–858.

Frank, W. M., 1977: The structure and energetics of the tropical cyclone. Part I: Storm structure. *Mon. Wea. Rev.,* **105,** 1119–1135.

Hart, R. E., R. N. Maue, and M. C. Watson, 2007: Estimating local memory of tropical cyclones through MPI anomaly evolution. *Mon. Wea. Rev.,* **135,** 3990–4005.

Hong, X., S. W. Chang, S. Raman, L. K. Shay, and R. Hodur, 2000: The interaction between Hurricane Opal (1995) and a warm core ring in the Gulf of Mexico. *Mon. Wea. Rev.,* **128,** 1347–1365.

Jacob, S. D., and L. K. Shay, 2003: The role of oceanic mesoscale features on the tropical cyclone–induced mixed layer response: A case study. *J. Phys. Oceanogr.,* **33,** 649–676.

Jansen, M. F., R. Ferrari, and T. A. Mooring, 2010: Seasonal versus permanent thermocline warming by tropical cyclones. *Geophys. Res. Lett.,* **37,** L03602, doi:10.1029/2009GL041808.

Kanamitsu, M., W. Ebisuzaki, J. Woollen, S.-K. Yang, J. J. Hnilo, M. Fiorino, and G. L. Potter, 2002: NCEP-DOE AMIP-II Reanalysis (R-2). *Bull. Amer. Meteor. Soc.,* **83,** 1631–1643.

Kaplan, J., and M. DeMaria, 2003: Large-scale characteristics of rapidly intensifying tropical cyclones in the North Atlantic basin. *Wea. Forecasting,* **18,** 1093–1108.

Kimball, S. K., and M. S. Mulekar, 2004: A 15-year climatology of North Atlantic tropical cyclones. Part I: Size parameters. *J. Climate,* **17,** 3555–3575.

Knapp, K. R., M. C. Kruk, D. H. Levinson, and E. J. Gibney, 2009: Archive compiles new resource for global tropical cyclone research. *Eos, Trans. Amer. Geophys. Union,* **90,** doi:10.1029/2009EO060002.

Knutson, T. R., R. E. Tuleya, W. Shen, and I. Ginis, 2001: Impact of CO_2-induced warming on hurricane intensities as simulated in a hurricane model with ocean coupling. *J. Climate,* **14,** 2458–2468.

——, C. Landsea, and K. A. Emanuel, 2010: Tropical cyclones and climate change: A review. *Global Perspectives on Tropical Cyclones: From Science to Mitigation,* J. C. L. Chan and J. D. Kepert, Eds., World Scientific Publishing Company, 243–286.

Lin, I.-I., C.-C. Wu, K. Emanuel, I.-H. Lee, C.-R. Wu, and I.-F. Pun, 2005: The interaction of Supertyphoon Maemi (2003) with a warm ocean eddy. *Mon. Wea. Rev.,* **133,** 2635–2649.

——, I.-F. Pun, and C.-C. Wu, 2009: Upper-ocean thermal structure and the western North Pacific category 5 typhoons. Part II:

Dependence on translation speed. *Mon. Wea. Rev.,* **137,** 3744–3757.

Liu, W. T., K. B. Katsaros, and J. A. Businger, 1979: Bulk parameterization of air-sea exchanges of heat and water vapor including the molecular constraints at the interface. *J. Atmos. Sci.,* **36,** 1722–1735.

Lloyd, I. D., and G. A. Vecchi, 2010: Submonthly Indian Ocean cooling events and their interaction with large-scale conditions. *J. Climate,* **23,** 700–716.

Mainelli, M., M. DeMaria, L. K. Shay, and G. Goni, 2008: Application of oceanic heat content estimation to operational forecasting of recent Atlantic category 5 hurricanes. *Wea. Forecasting,* **23,** 3–16.

McPhaden, M. J., and Coauthors, 2009: Ocean-atmosphere interactions during Cyclone Nargis. *Eos, Trans. Amer. Geophys. Union,* **90,** 53–54.

Pasquero, C., and K. Emanuel, 2008: Tropical cyclones and transient upper-ocean warming. *J. Climate,* **21,** 149–162.

Price, J. F., 1981: Upper ocean response to a hurricane. *J. Phys. Oceanogr.,* **11,** 153–175.

Schade, L., and K. Emanuel, 1999: The ocean's effect on the intensity of tropical cyclones: Results from a simple coupled atmosphere–ocean model. *J. Atmos. Sci.,* **56,** 642–651.

Shen, W., and I. Ginis, 2003: Effects of surface heat flux-induced sea surface temperature changes on tropical cyclone intensity. *Geophys. Res. Lett.,* **30,** 1933, doi:10.1029/2003GL017878.

Sriver, R. L., and M. Huber, 2007: Observational evidence for an ocean heat pump induced by tropical cyclones. *Nature,* **447,** 577–580, doi:10.1038/nature05785.

Thorncroft, C., and K. Hodges, 2001: African easterly wave variability and its relationship to Atlantic tropical cyclone activity. *J. Climate,* **14,** 1166–1179.

Vecchi, G. A., and B. J. Soden, 2007: Global warming and the weakening of the tropical circulation. *J. Climate,* **20,** 4316–4340.

Wentz, F. J., C. Gentemann, D. Smith, and D. Chelton, 2000: Satellite measurements of sea surface temperature through clouds. *Science,* **288,** 847–850.

Xie, S.-P., M. Ishiwatari, H. Hashizume, and K. Takeuchi, 1998: Coupled ocean-atmospheric waves on the equatorial front. *Geophys. Res. Lett.,* **25,** 3863–3866.

Yu, L., and R. A. Weller, 2007: Objectively analyzed air–sea heat fluxes for the global ice-free oceans (1981–2005). *Bull. Amer. Meteor. Soc.,* **88,** 527–539.

Zedler, S. E., T. D. Dickey, S. C. Doney, J. F. Price, X. Yu, and G. L. Mellor, 2002: Analyses and simulations of the upper ocean's response to Hurricane Felix at the Bermuda Testbed Mooring site: 13–23 August 1995. *J. Geophys. Res.,* **107,** 3232, doi:10.1029/2001JC000969.

Zhang, S., M. J. Harrison, A. Rosati, and A. Wittenberg, 2007: System design and evaluation of coupled ensemble data assimilation for global oceanic climate studies. *Mon. Wea. Rev.,* **135,** 3541–3564.