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Contribution of tropical cyclones to meridional heat transport by the oceans

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Abstract. Tropical cyclones mix warm surface waters with cooler water within the thermocline, leaving pronounced, cold wakes that over a period of weeks are restored to normal conditions by mixing and surface fluxes. This restoration is associated with net, vertically integrated heating of ocean columns, which in statistical equilibrium must be balanced by oceanic heat transport out of the regions affected by the storms. Observed tropical cyclone tracks together with coupled ocean-atmosphere hurricane models are used to estimate the net ocean heating induced by global tropical cyclone activity during one calendar year (1996). This estimate, amounting to $(1.4 \pm 0.7) \times 10^{15}$ W, represents a substantial fraction of the observed peak poleward heat flux by the oceans, suggesting that tropical cyclones may play an important role in driving the thermohaline circulation and thereby in regulating climate. In particular, the strong sensitivity of tropical cyclone intensity to tropical ocean temperatures in turn implies that the net poleward heat flux by the ocean is sensitive to tropical temperature, reducing tropical climate sensitivity and increasing climate sensitivity at higher latitudes.

1. Introduction

The world's oceans carry roughly half of the net equator-to-pole heat flux necessary to balance the satellite-observed meridional distribution of net radiative flux at the top of the atmosphere [*Macdonald and Wunsch,* 1996] and thus play a critical role in setting the global temperature distribution. Of this half, a substantial fraction is thought to be carried by the deep, meridional overturning circulation (MOC) with the rest carried by the wind-driven lateral gyres in the principal ocean basins [*Hall and Bryden,* 1982; *Wang et al.,* 1995]. It is therefore of critical importance to understand the physics of the MOC, and especially the factors that control its poleward heat transport.

There are two basic physical processes that may, in principle, drive the observed MOC, as illustrated in Figure 1. The first relies on differential Ekman pumping driven by the curl of the surface wind stress and can, in principle, result in purely adiabatic (isopycnal) flow below the ocean's surface mixed layer. In this view, all the enthalpy sources and sinks are confined to the surface mixed layer and therefore play little or no role in the energetics of the MOC, as shown by the heat engine argument proposed by *Sandström* [1908] and, subsequently, modified by *Jeffreys* [1925]. This mechanism for driving the MOC has been discussed in some detail by *Toggweiler and Samuels* [1998].

The second mechanism relies on the downward turbulent mixing of relatively warm water into colder water at depth, thus introducing an enthalpy source at relatively high pressure, resulting in a circulation in which buoyancy plays an active role. In this second view, the rate of overturning is, in the long run, controlled by the rate of downward diffusion of enthalpy, as reviewed by *Munk and Wunsch* [1998], who estimated that 2.1×10^{12} *W* must be expended in turbulent mixing to explain the estimated 30 Sv of global deep water formation.

Three particular aspects of this second mechanism are noted here. First, to be effective, the turbulent mixing must occur on the warm side of the system; mixing in regions of deep water formation has little or no effect [*Marotzke and Scott,* 1999]. Second, to explain the steady state ascent of water through isopycnal surfaces, there must be a net, vertically integrated density sink, and this requires that the mixing ultimately communicate with the surface. Mixing that is purely internal to the ocean and does not affect the surface density cannot cause a vertically integrated density source and leads instead to the formation of internal mixed layers. Third, the required mixing can be highly localized in space (J. R. Scott and J. Marotzke, The location of diapycnal mixing and the meridional overturning circulation, *J. Phys. Oceanogr.*, submitted manuscript, 2000.) and, perhaps, in time as well.

The two aforementioned mechanisms are by no means mutually exclusive and may operate in concert. Here we apply the converse of the reasoning behind the second mechanism; that is, if one identifies a source of vertical turbulent mixing and can estimate the associated column-integrated heating, then in equilibrium, this heating must be balanced by the divergence of a lateral heat flux in the ocean.

Wunsch [1998] estimated the rates of ocean mixing which result from a 10-day mean, zonal surface wind. The work done on ocean currents was estimated using the relation

$$W = \langle \tau \bullet \mathbf{v}_g \rangle, \tag{1}$$

where $\boldsymbol{\tau}$ is the surface wind stress and $\mathbf{v}_g$ is the geostrophic ocean current at the sea surface, which was estimated using TOPEX/POSEIDON altimetric data, and the brackets denote a time mean. Ten-day average values of the surface stress

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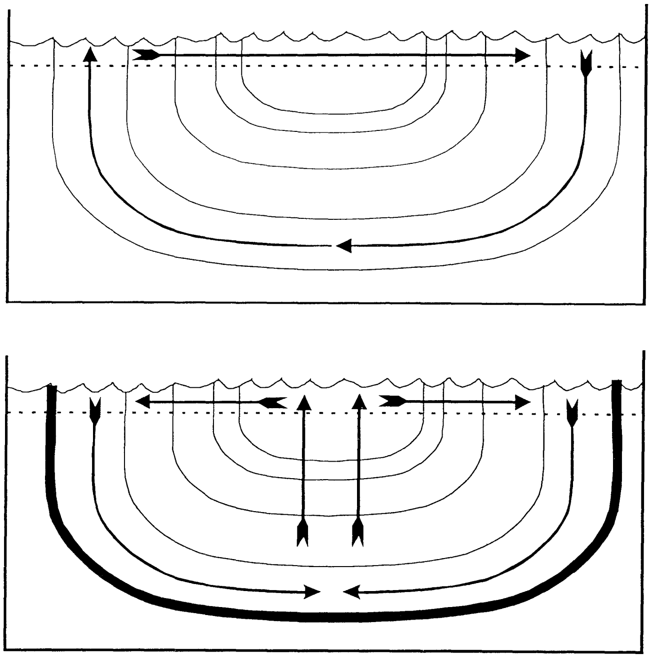

Figure 1. Two views of the meridional overturning circulation, showing vertical cross sections from pole to pole. In each panel, the dashed lines denote the base of the mixed layer and the thin sold lines are isopycnal surfaces. In the first view (top), Ekman pumping forces fluid down in one hemisphere and up in the other, and flow below the mixed layer is adiabatic. Since all density sources and sinks are in the mixed layer an thus approximately at the same pressure, buoyancy does no net work on the circulation. In the second view (bottom), cold water sinks in either or both hemispheres at high latitudes and returns by flowing across isopycnal surfaces on the warm side of the system. Downward turbulent mixing of warm water must balance upward advection by the mean flow in the return branch of the circulation. Net, column-integrated heating must occur in the tropics in this case, while most of the cooling is applied in the mixed layer. The induced poleward heat flux cannot extend poleward of the isopycnal surface (thick curve) corresponding to the greatest depth to which mixing extends in the tropics . These two views are not mutually exclusive.

from NCEP analyses were used. Since (1) is a quadratic relation, there are contributions to it from both the (10-days) mean geostrophic flow coupled with the mean stress and from the correlations between the fluctuating components of each quantity. *Wunsch* [1998] estimated that the contribution from the fluctuations was relatively small and calculated a global average contribution from the 10-day mean surface flow and stress of about 1×10^{12} *W*, most of which comes from the Southern Ocean. (If the dissipation is also dominated by the Southern Ocean, then this power deposition by the mean wind is insufficient to explain the observed oceanic heat flux out of the tropics.)

Here we note two limitations in estimating the power available for mixing from (1). First, it is by no means clear that all or even most of the energy applied to the mean surface currents is ultimately used to raise the potential energy of the ocean by mixing buoyancy; in this sense, (1) places an upper bound on the mixing power. (*Wunsch* [1998], and others assume that 15-20% of the energy deposited by wind stress is available for mixing.) However, on the other hand, (1) ignores the flow of wind energy into nongeostrophic flows, including surface waves and inertial oscillations, some of which may ultimately be used up in mixing. From this point of view, (1) may underestimate the mixing.

Although tropical cyclones are comparatively infrequent, with about 80 events each year around the globe [*Anthes,* 1982], they are very efficient mixers of the upper ocean because they produce rapid, unbalanced accelerations of the ocean's mixed layer, which in turn lead to violent mixing [*Price,* 1981]. Since this mixing occurs episodically and on timescales much less than characteristic thermal restoration times, it leads to a noticeable and often spectacular cooling of the upper ocean [e.g., *Leipper,* 1967]. Over a period on the order of several months [*Nelson,* 1996], the cold wake left by tropical cyclones recovers to near-normal temperatures. This process is illustrated schematically in Figure 2. If the volume-integrated negative temperature anomaly is known, then it is straightforward to calculate the amount of heating necessary to restore the upper ocean to climatology. In the long term, this heating, summed over all the events, must be balanced by lateral oceanic heat transport out of the regions affected by the storms. By measuring the volume-integrated cold anomaly left by each storm, one can obtain an estimate of the global effect of tropical cyclones on lateral heat transport without having to deal with the specific physics of the mixing process.

In section 2, observations of the upper ocean response to tropical cyclones are reviewed, with special attention to the magnitude, width, and depth of penetration of the cold wakes left by the storms. In section 3 we describe the methods and data used to estimate the tropical cyclone-induced heat flux into the upper ocean during the calendar year 1996. A discussion of the results of this estimate and their possible significance is presented in section 4.

2. Upper Ocean Response to Tropical Cyclones

Indications that tropical cyclones produce anomalous cold wakes date back at least to the work of *Fisher* [1958] and *Jordan* [1964]. The first systematic attempt to make comprehensive upper ocean measurements just after the passage of a hurricane was described by *Leipper* [1967]. Hydrographic cross sections of the Gulf of Mexico were made just after Hurricane Hilda passed over the gulf in 1964. By comparing these sections to several made before Hilda entered the gulf, it was evident that the hurricane caused the mixed layer temperature to decrease by as much as 5°C spanning a distance of around 400 km across the path of the storm. Subsequent observational studies by *Federov et al.* [1979] and *Pudov et al.* [1979] showed the distinctive rightward bias in the negative sea surface temperature anomaly pattern that accompanies Northern Hemisphere storms. *Withee and Johnson* [1976] described the pattern of response of the ocean to the passage of Hurricane Eloise of 1975 as measured by two moored buoys in the Gulf of Mexico. Cooling down to the level of the base of the undisturbed mixed layer was accompanied by warming at greater depths, suggesting that mixing (rather than surface heat loss or vertical advection) was the dominant effect on temperature. Current measurements down to 530 m exhibited strong oscillations of near-inertial period, showing that tropical cyclones cause substantial disturbance to the water column, even at large depths.

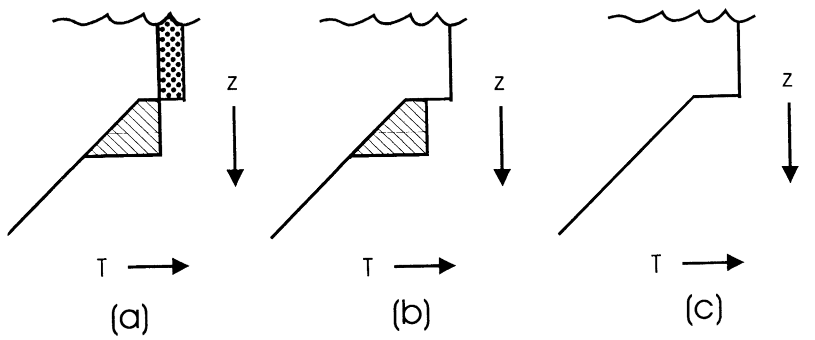

Figure 2. Three stages of mixing and recovery. In the first stage (a), strong mixing deepens the mixed layer, creating a cold anomaly at the top (dotted) and a compensating warm anomaly below (striped). There is no net column heating during this stage. In the second stage (b), the cold anomaly is removed by net surface enthalpy fluxes; it is during this stage that column-integrated heating takes place. In the third stage (c), the warm anomaly is removed advectively by buoyancy adjustment to the surrounding ocean. Stages (b) and (c) may take place concurrently.

Price [1981] modeled the response of the upper ocean to hurricanes and demonstrated that about 85% of the observed surface cooling results from entrainment, with the remainder owing to surface heat fluxes. He also showed that the rightward bias of the temperature anomalies results from the near resonance of the storm's wind field with inertial oscillations to the right of the track, where the wind veers around the compass. This, in turn, strongly suggests that much of the entrainment arises from turbulence associated with velocity shears at the base of the mixed layer, rather than through the effect of surface-generated turbulence. *Price* [1983] also showed that much of the energy deposited in near-inertial waves propagates rapidly downward through the thermocline, where it may be available for mixing. Observations reported by *Leipper* [1967] and by *Pudov* [1979] both show measurable temperature changes at depths of several hundred meters, though how much of this change results from mixing is not known.

An example of the cold wake produced by a hurricane, Edouard of 1996, is shown in Plate 1. The volume of surface waters cooled by entrainment induced by tropical cyclones can be impressive. For example, a wake whose average negative temperature anomaly of 3°C extends downward to 50 m with a cross-track dimension of 400 km and a track length of 2000 km would require nearly 5×10^{20} J of surface heating to restore it to prestorm temperature. When integrated over global tropical cyclone activity, this represents a very substantial amount of vertically integrated heating of ocean columns and leads us to seek a more accurate estimate of the average heating induced by tropical cyclones. Such an estimate is attempted in the following section.

3. Model-Based Estimate of Column-Integrated Heating Induced by Tropical Cyclones

The approach here is to use a very simple, coupled hurricane model to "hindcast" individual tropical cyclones over a period of 1 year. This model is described by *Emanuel* [1999] and has been shown to yield surprisingly accurate hindcasts of tropical cyclone intensity, which depend critically on the negative feedback of ocean cooling. Since the intensity hindcasts work well, it is reasonable to suppose that the model has skill in simulating the storm-induced cooling of the upper ocean near the storm center, where the storm is sensitive to the sea surface temperature.

The model consists of an axisymmetric atmospheric tropical cyclone model which assumes gradient and hydrostatic balance and uses a representation of cumulus convection based on the postulate of subcloudlayer quasi-equilibrium [*Raymond,* 1995]. The model is phrased in "potential radius" coordinates [*Schubert and Hack,* 1983], where the potential radius R is defined by

$$fR^2 = 2rV + fr^2, \tag{2}$$

where f is the Coriolis parameter, r is the physical radius from the storm center, and V is the azimuthal component of the velocity. The right side of (2) is just twice the absolute angular momentum per unit mass. Convection is assumed to maintain moist adiabatic lapse rates on surfaces of constant R, in keeping with the theory of slantwise moist convection [*Emanuel,* 1983]. The use of potential radius coordinates also gives an exceptionally high radial resolution in the critical eyewall region of the storm, where angular momentum surfaces are closely packed. Details of the atmospheric model are given by *Emanuel* [1995].

The ocean model was developed by *Schade* [1997] and consists merely of a string of independent vertical columns along the track of the hurricane, which must be specified. (The model does not predict the track of the storm.) The initial thermal profile is assumed to have the shape shown in Figure 2, with the initial mixed layer depth and submixed layer thermal stratification taken from the monthly mean climatology of *Levitus* [1982]. Salinity effects are not accounted for. The bulk horizontal velocity of the ocean's mixed layer is predicted according to

$$\rho_O \frac{\partial (hu)}{\partial t} = \tau_S \tag{3}$$

where ρ_0 is the density of water, h and u are the local mixed layer depth and horizontal velocity, and τ_s is the surface stress supplied from the hurricane model. Horizontal advection and Coriolis accelerations are neglected, and it is assumed that the water below the mixed layer is at rest.

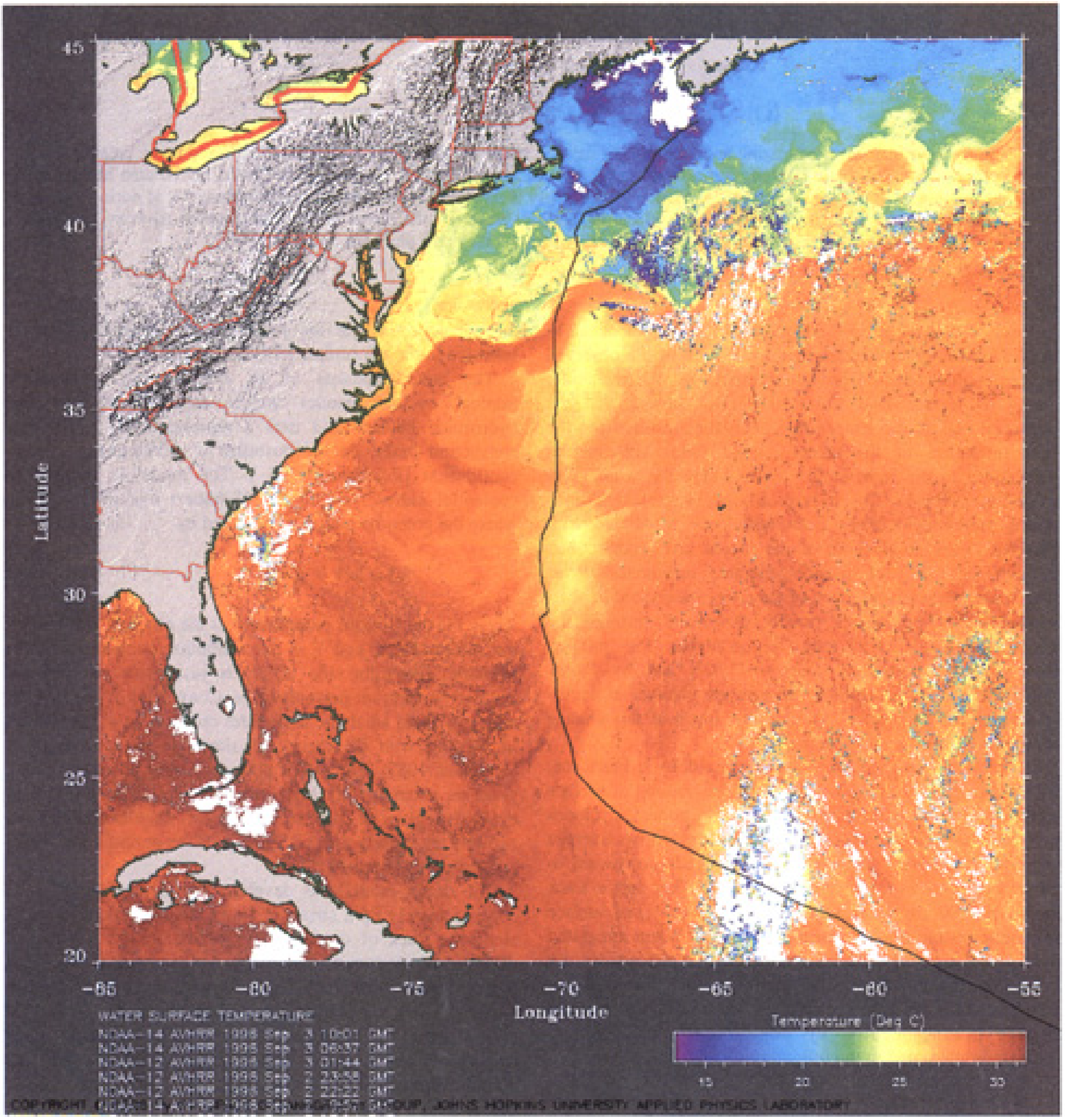

Plate 1. Satellite-derived water surface temperature shortly after the passage of Hurricane Edouard in 1996. Storm center track indicated by thin black line. Courtesy of the Johns Hopkins University Applied Physics Laboratory.

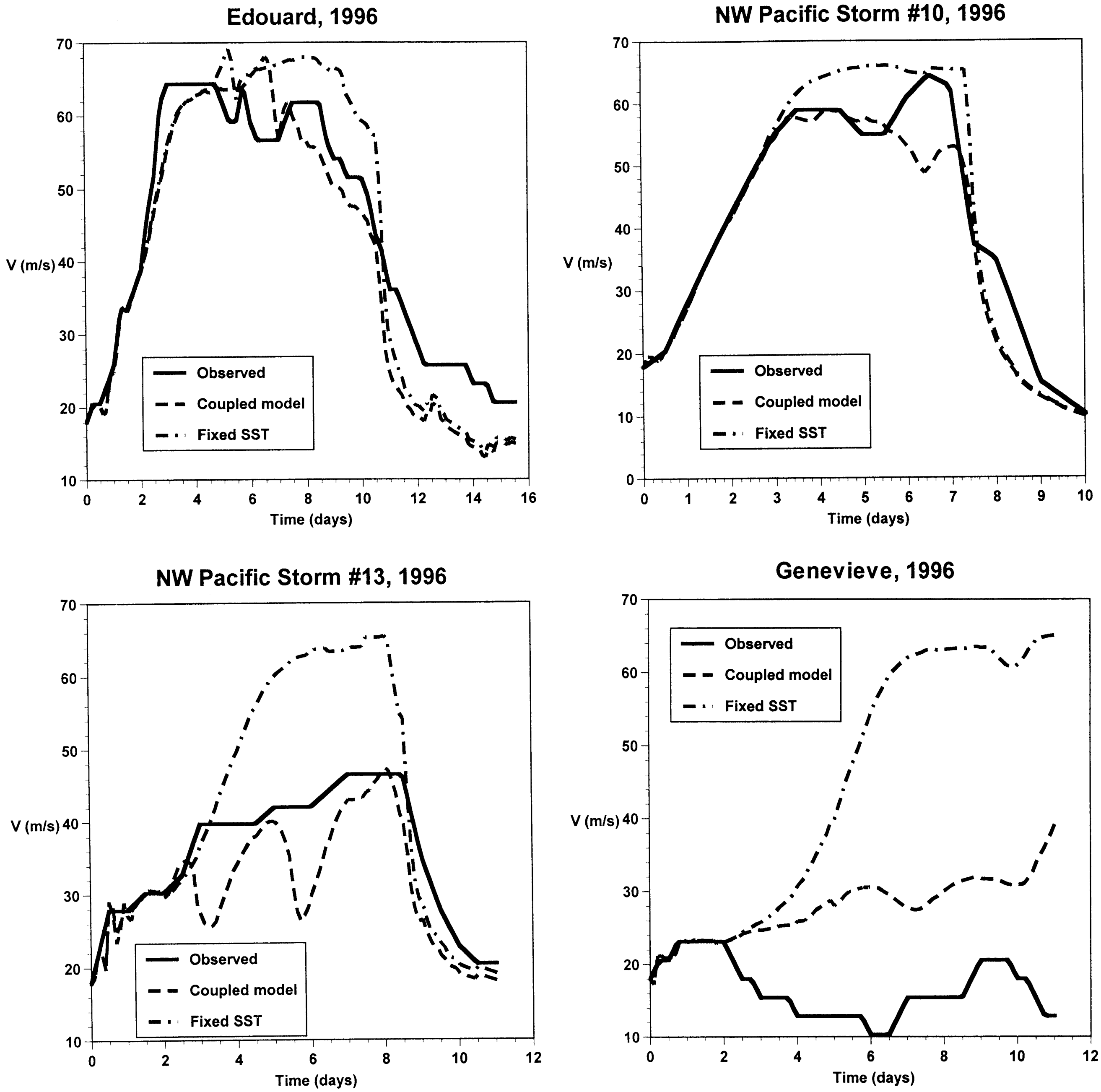

Figure 3. Observed (solid) and model-simulated (longdashed) evolutions of the maximum wind speed in four tropical cyclones during 1996. In each panel, the shortdashed line shows a model simulation with fixed sea surface temperature.

Entrainment into the mixed layer proceeds at the rate required to keep a bulk Richardson number near its critical value [*Price,* 1979]. This Richardson number is defined as

$$Ri \equiv \frac{\Delta B h}{u^2} \tag{4}$$

where ΔB is the buoyancy jump across the base of the mixed layer. In this model we enforce $Ri=1$ at each time step.

As the model storm approaches an ocean water column, (3) is integrated forward in time using the surface wind stress predicted by the model. The mixed layer temperature anomaly (and thus the associated ΔB) is diagnosed from the mixed layer depth anomaly, invoking conservation of vertically integrated enthalpy. (The reduction of column-integrated enthalpy by surface enthalpy fluxes is ignored.) The integration is discontinued once the storm center passes the ocean column in question, since (3) cannot be expected to apply once the wind has reversed direction. The sea surface temperature fed back into the atmospheric model is a weighted average of the predicted temperature at the given radius ahead of the storm and the predicted temperature at the storm center, given the absence of predictions of ocean temperature along the storm track to the rear of the storm center.

At first glance it would seem that this representation of upper ocean behavior is far too crude to produce realistic simulations. In particular, the neglect of Coriolis turning of the ocean currents removes the resonant mechanism shown by *Price* [1981] to be so important for creating the cold wake of the storm. Nonetheless, confidence in the use of (3) and 4) arises from two different observations. First, *Schade* [1997] compared the evolution of hurricane intensity in the coupled model described above with that simulated by the same atmospheric model but coupled to the fully three-dimensional ocean model of *Cooper and Thompson* [1989], in the case of a storm translating at constant velocity. The intensities of storms simulated with the simple ocean model were within a few percent of those simulated using the fully three-dimensional ocean model (which, however, uses the same closure for mixed layer entrainment as the simple model). *Schade* [1997] ascribes this lack of difference to the fact that (1) little Coriolis turning is observed in the full model between the onset of strong winds and the passage of the storm center and (2) the storm thermodynamics are indifferent to sea surface temperatures in the cold wake region, far outside the storm's eyewall. (For this reason, the correspondence between the simple and the complete ocean models is likely to fail in the case of storms that turn back over their own cold wakes.)

The second source of confidence arises from use of the simple coupled model to hindcast real events. Figure 3 shows particular examples of hindcasts and compares them to hindcasts made with the same model but assuming fixed ocean temperature. This demonstrates that the coupling is both important and has been modeled with some success, from the point of view of simulating the feedback on storm intensity. Other examples are shown by *Emanuel* [1999].

On the other hand, the object of the present exercise is to predict the whole of the ocean temperature anomaly produced by the storm, including its wake. To do this, we first use the simple model to predict the sea surface temperature anomaly and mixed layer depth anomaly at the storm center, and then assume that the product of these two anomalies at the storm center is proportional to the same product integrated over the whole volume of the actual storm wake. This is tantamount to assuming that all storm wakes are geometrically similar. The constant of proportionality is calculated by comparing the storm center anomaly product to the column-integrated product in the model that uses the fully three-dimensional ocean component, in a particular integration of the latter.

The amount of net surface heating necessary to restore the cold wake to prestorm temperature is given by

$$Q = \iiint \rho_o C_l \Delta T dh dW dL \tag{5}$$

where C_l is the heat capacity of seawater, ΔT is the magnitude of the negative part of the temperature anomaly, h is the depth to which the negative part of the temperature anomaly extends, and W and L are the cross-track and long-track extents, respectively, of the storm-induced temperature anomaly.

To evaluate (5) using the simple coupled model, we approximate it as

$$Q \cong \gamma \int \rho_o C_l \Delta T_c h_c dL \tag{6}$$

where ΔT_c and h_c are the magnitude of the cold part of the temperature anomaly and the depth to which it extends, both evaluated at the storm center. The constant of proportionality, γ, has the dimensions of length and is given (by comparing (5) to (6)) by

$$\gamma = \frac{\iint \Delta T dh dW}{\Delta T_c h_c} \tag{7}$$

Once again, the assumption made here is that both the temperature anomaly and the depth anomaly at any point in the wake scale with those quantities evaluated at the storm center, while the width of the wake is considered not to vary among storms. The latter is not on the face of it a good assumption, since the geometric size of tropical cyclones is highly variable, but on the other hand, there is no known correlation between storm size and storm intensity as measured, for example, by maximum wind speed [*Merrill*, 1984], so using an average value of storm size in (7) should not bias the estimate of Q over a sufficiently large sample of events.

The single value of γ used here is calculated from a single run of the hurricane model coupled to the three-dimensional ocean model of *Cooper and Thompson* [1989], as described by *Schade and Emanuel* [1999]. This run assumes that the hurricane is moving at a uniform speed of 6 m s^{-1} over an ocean mixed layer of uniform, unperturbed depth of 30 m. Figure 4 shows the mixed layer depth and temperature anomalies used to estimate γ from (7). We perform the integral in (7) over a cross section taken at a representative position in the wake of the storm (Figure 4). We arrive at a value of γ of

$$\gamma = 970 \text{ km}. \tag{8}$$

The actual width of the wake is closer to 400 km; the larger value of γ reflects the greater magnitude of the temperature and depth anomalies in the wake than at the center of the storm. Because we used only one run of the full physics model to estimate the value of γ, some uncertainty in our heat flux estimates will arise from lack of a more precise estimate of its value.

Given this value of γ, the heating necessary to restore the wake to its prestorm temperature is estimated by running the simple coupled model of *Emanuel* [1999] and using (7), with the integral taken along the whole track of the storm. This is done for all tropical cyclones recorded during calendar year 1996, using "best track" data sets obtained from Colorado State University for all Atlantic and eastern North Pacific storms, and from the Joint Typhoon Warning Center for storms in the western North Pacific, Indian Ocean, and all tropical cyclones in the Southern Hemisphere. There were 83 recorded tropical cyclones in 1996, slightly more than the long-term average of around 78.

Model runs were completed for 67 of the 83 events of 1996; the remaining 16 were not run either because the maximum recorded wind during the lifetime of the event did not exceed 17 m s^{-1} or because no wind reports were available for the event in question. It is not likely that the missing events contributed in any significant way to the total, because they were probably very weak. As in the work of *Emanuel* [1999], the model storms were run along their observed tracks but through monthly climatological upper ocean and atmosphere thermodynamic conditions. The thermodynamic state of the atmosphere and the unperturbed sea surface

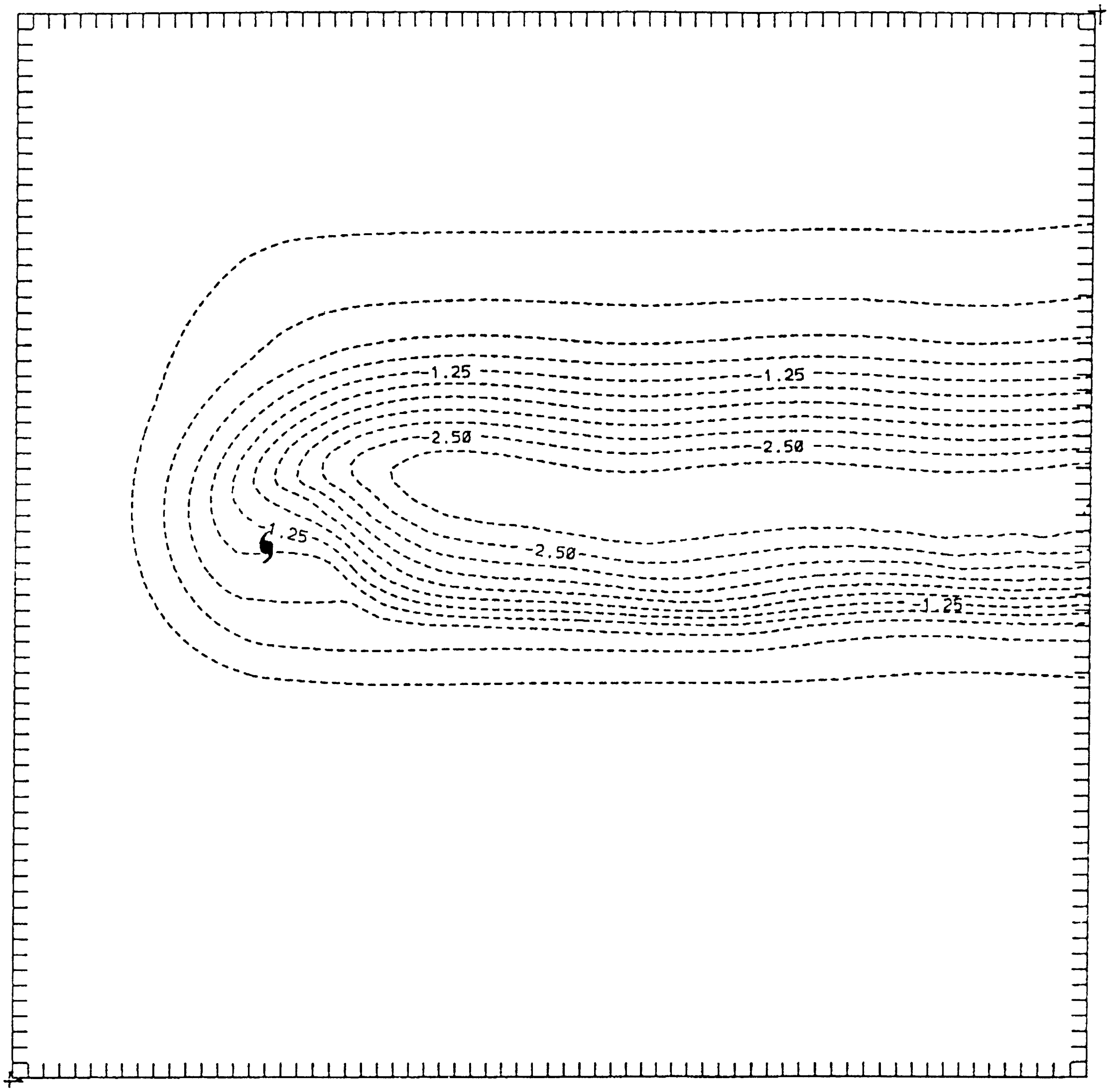

Figure 4. Mixed layer depth (solid, in meters) and sea surface temperature (dashed, in Kelvin) anomalies produced by a hurricane translating from right to left at a uniform speed of 7 m s^{-1}, over an ocean with an initial mixed layer depth of 40 m. The current location of the model storm is denoted by the hurricane symbol. Tick marks on boundary are spaced 20 km apart. Model is as described by *Schade and Emanuel* [1999].

temperature are represented in the model (as in the work of *Emanuel,* 1999) by a single quantity: the potential intensity, which is the maximum wind speed that can be achieved in a tropical cyclone according to an energy bound [e.g., *Emanuel,* 1987]. The monthly climatological values of potential intensity [e.g., *Bister and Emanuel,* 1998] used in the hurricane model were derived from reanalysis data supplied by the National Centers for Environmental Prediction (NCEP), while the monthly climatological values of the upper ocean mixed-layer depths were derived from data supplied by *Levitus* [1982]. The monthly value was assigned to the 15th day of each month and linearly interpolated to the date in question.

Examination of each of the individual model runs shows that while some of the simulations either underpredicted or overpredicted the storm intensity, there was little systematic bias evident. The results of some of the simulations are shown in Figure 3.

Summing the heat input needed to restore the cold wakes to prestorm temperature over all 67 events, using (6) and (8), and dividing by 366 days yields an average heating rate of 1.4×10^{15} *W.* Given the uncertainties involved together with year-to-year fluctuations in global tropical cyclone activity, this estimate is likely to be in error by as much as 50%, thus we estimate the average annual column-integrated heating induced by tropical cyclones to be $(1.4\pm0.7)\times10^{15}$ *W.*

A possible weakness of this analysis is our assumption that the cold anomalies produced by tropical cyclones are entirely eliminated by surface heat flux over a period of a few weeks. Although observations show rather conclusively that these cold anomalies extend down to below the depth of the unperturbed mixed layer and that the surface temperature associated with these anomalies recovers in a matter of weeks, it is not clear that the restoration of the surface temperature extends through the entire depth occupied by the cold anomaly. Here we present some evidence that this restoration is indeed rapid and extends through an appreciable depth.

Figure 5 shows a sequence of sea surface height anomalies

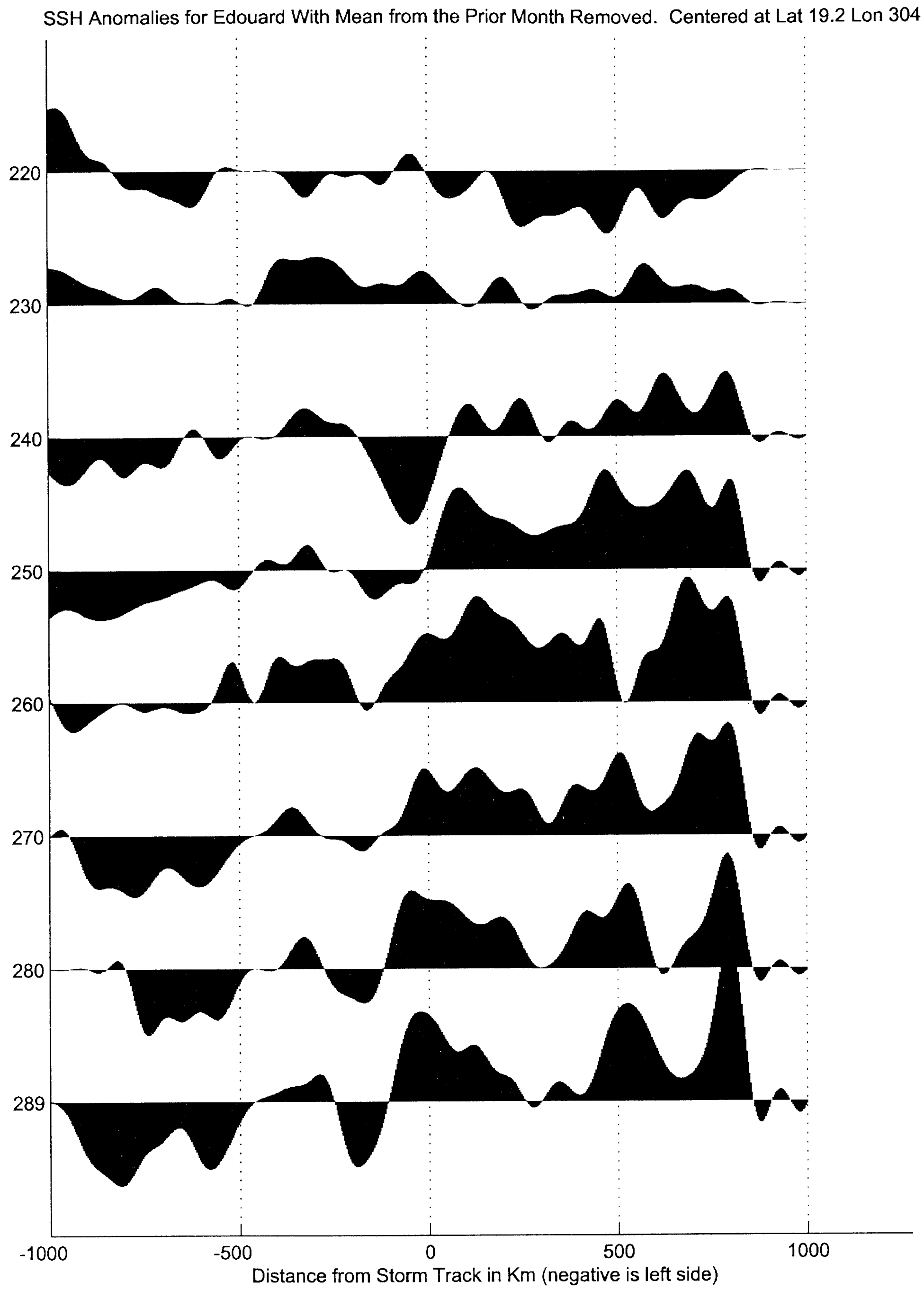

Figure 5. Cross-track sections of the sea surface height anomaly from TOPEX/POSEIDEN at 10-day intervals in August-September, 1996. The height anomaly corresponding to the vertical separation between the transects is 20 cm. The transect is centered at 19.2°N, 56°W, and the time of the transect is indicated 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 of the Massachusetts Institute of Technology.)

from prestorm conditions (with tidal influences removed), in a transect across the path of Hurricane Edouard of 1996, whose effects on sea surface temperature are shown in Plate 1and was one of the storms in our sample. These transects are separated by 10-day intervals. Edouard passes through the center of the transect on Julian day 239. After one day, the strongest signal in the sea surface altitude is the depression very near the track of the storm. This very likely reflects the Ekman pumping induced by the strongly cyclonic wind stress near the storm center. This brings cooler water upward and thus decreases the column heat content. Neglecting any purely barotropic response to the moving storm and salinity perturbations, the sea surface height anomaly is related hydrostatically to the depth-integrated temperature anomaly by

$$h' = \int_0^\infty \beta T' dz, \tag{9}$$

where T' is the temperature anomaly and β is the coefficient of thermal expansion of seawater. Note that since vertical mixing cannot directly alter the enthalpy of the column, no sea surface height anomaly can result directly from vertical mixing. However, Ekman pumping, by creating a net divergence of the column-integrated enthalpy flux, reduces the integrated temperature and thus the sea surface height through (9).

However, 11 days after Edouard's passage, on Julian day 250 in Figure 5, a prominent positive anomaly has formed to the right of Edouard's track, where one expects the vertical mixing to be most vigorous. This is precisely the signal one expects from (9) if there has been a net downward surface flux into the ocean, increasing the column enthalpy. We may estimate the total anomaly of ocean heat content, integrated across the storm path, using (9):

$$Q_L = \int_{-\infty}^{\infty} \frac{1}{\beta} \rho_0 C_l h' dW, \tag{10}$$

where W is the cross-track distance. From Figure 5 at Julian day 250, this amounts to about 8×10^{14} J m^{-1}. At the same time, the model prediction of Q_L, using (6) and (8) without integrating along the track, is about 4×10^{14} J m^{-1}. Thus if anything, the model is underestimating the induced heating by this storm, perhaps because Edouard was rather larger than the average tropical cyclone.

4. Discussion

The peak meridional heat flux carried by the ocean has been estimated at around $2\times10^{15} W$ [*Macdonald and Wunsch*, 1996]. Here we estimate that $(1.4\pm0.7)\times10^{15} W$ of net column heating is required globally to restore the cold wakes of tropical cyclones to prestorm conditions. This net column heating must be balanced, on average by lateral heat flux away from the tropics. Thus it is conceivable that much of the observed lateral heat flux carried by the ocean is induced by tropical cyclones.

One possible objection to this inference is that much of the net column heating induced by tropical cyclones could be balanced by lateral fluxes that redistribute enthalpy locally within the Tropics without much export to higher latitudes. Some evidence that this is not the case is provided by *Scott and Marotzke* [2000], who performed numerical experiments that show the response of the world oceans to vertical mixing that is highly localized in the tropics. These clearly show that the response is global and entails a vigorous meridional overturning component. However, this model did not include wind driving of the ocean, and so there remains the possibility that in a more realistic model, the lateral heat transport induced by tropical cyclone activity would be confined within the tropics. However, certain aspects of the observed ocean heat transport are consistent with what one would expect if tropical cyclones were an important contribution. For example, the divergence of the meridional heat flux is largest in the Pacific basin, where tropical cyclones are most plentiful, and the heat flux in the South Atlantic (where there are no tropical cyclones) is northward and nearly free of divergence [*Trenberth et al.*, 2000]. The northward flux is consistent with the existence of large vertical mixing in the Pacific and Indian Ocean basins with deep water formation in the North Atlantic.

It is obvious from the lower part of Figure 1 that the poleward heat flux induced by vertical mixing in the tropics cannot extend poleward of the outcropping latitude of that

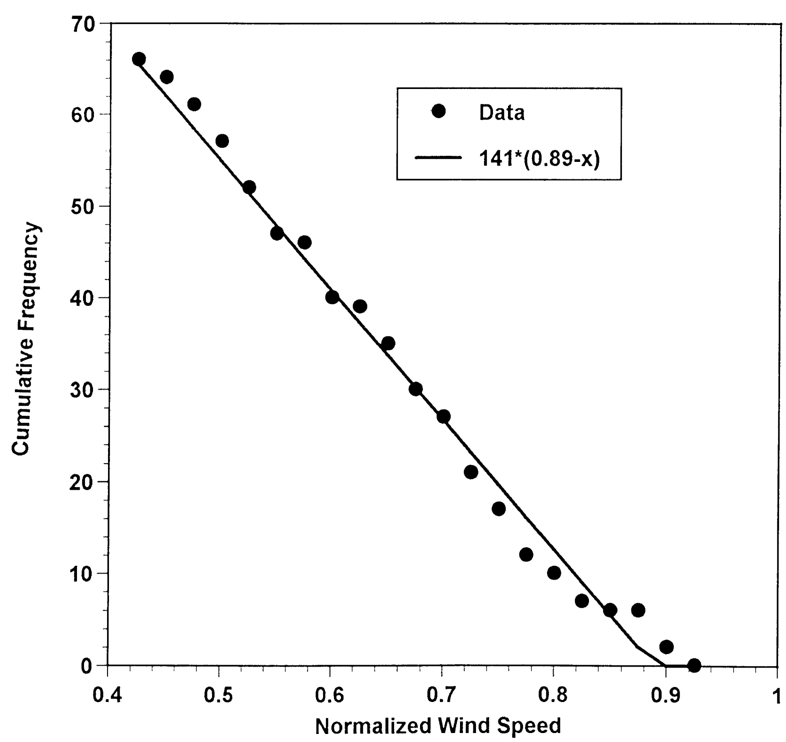

Figure 6. Cumulative distribution function (CDF) of lifetime maximum wind speeds of the North Atlantic hurricanes after 1957 whose lifetime maximum intensity was not limited by declining potential intensity. Wind speed is normalized by monthly climatological potential wind speed at the reported position of the tropical cyclone. The ordinate shows the total number of events whose normalized lifetime maximum wind speed exceeds the value on the abscissa. The straight line is from linear regression. [From *Emanuel*, 2000].

isopycnal surface that represents the greatest depth to which the mixing extends in the tropics. The observed sea surface temperature anomalies caused by tropical cyclones require mixing into the thermocline. The calculations described in the previous section show a mixing that extends no deeper than about the 17°C isotherm; nominally, this implies that the induced poleward heart flux should not extend poleward of the 17°C sea surface isotherm (neglecting the effects of salinity and pressure on the density of seawater). However, current measurements near hurricanes show pronounced inertial oscillations even near the base of the thermocline [*Price,* 1981], so it is possible that, in the long term, tropical cyclone-related mixing effects may extend through the entire thermocline.

If tropical cyclones do indeed contribute substantially to the MOC, the implications for regulation of the Earth's climate are interesting. Poleward heat flux by the ocean would be strongly affected by net global tropical cyclone activity. Recent work on the effect of climate change on tropical cyclone activity allows us to make some tentative speculations about how tropical cyclones, and thus perhaps the thermohaline circulation, might respond to climate forcing.

If one characterizes the intensity of each tropical cyclone by the peak wind speed achieved in the course of its life and normalizes this peak wind by the potential intensity derived from monthly mean climatology (see section 3), and then sums over all the events in particular ocean basins over a range of years, a cumulative distribution of normalized tropical cyclone intensity can be created. This was done for Atlantic and northwest Pacific tropical cyclones over several decades, as reported by *Emanuel* [2000]. An example of such a cumulative distribution function, for Atlantic storms of hurricane strength, is shown in Figure 6. In both the Pacific and the Atlantic the cumulative distribution function of normalized tropical cyclone intensity is distinctly linear for storms of hurricane strength, with a zero intercept at a normalized wind speed of close to unity. To the extent that this linear function is truly universal, one can characterize the long-term climatology of tropical cyclone activity by only two parameters: the overall frequency of events and the potential intensity, which is easily calculated from climatological sea surface and atmospheric temperature [*Bister and Emanuel,* 1998].

Unfortunately, there is almost no understanding of the climate factors that control the global frequency of tropical cyclones [e.g., *Henderson-Sellers et al.,* 1997]. On the other hand, the calculation of potential intensity is straightforward, if the mean climate of the tropics is known. From basic thermodynamic considerations and from the results of a global climate simulation, *Emanuel* [1987] estimated that the potential intensity of tropical cyclones increases about 7 m s^{-1} for every 2°C increase in tropical sea surface temperature. This sensitivity is also consistent with the work of *Knutson et al.* [1998], who used a three-dimensional tropical cyclone model to calculate the intensities of storms in a greenhouse-warmed climate, as simulated by a global climate model.

If we assume for the moment that the global frequency of events remains constant and that the cumulative distribution function of tropical cyclone intensity remains linear, then a 2°C increase in tropical sea surface temperatures would result in an increase of potential intensity of about 10% and the actual intensities of all storms of hurricane strength would also increase by 10%. However, the mixing effect in the ocean increases more nearly as the cube of the surface wind speed, and so the upper ocean heating induced by global tropical cyclone activity would increase by about 30%. This implies a change in the equilibrated poleward heat flux by the MOC of about $4\times10^{14} W$. Dividing this value by the surface area of the tropical oceans gives an average cooling of the upper ocean of about 3 Wm^{-2}, which is comparable to the magnitude of the climate forcing needed to warm the tropical oceans by about 2°C. Thus we may postulate that the response of tropical cyclone activity to climate forcing may constitute a strong negative feedback on changes in tropical upper ocean temperature, through its effect on lateral heat export. Conversely, *increases* in the climate forcing of tropical ocean temperature would be expected to magnify the thermal response of higher latitudes, where the MOC deposits heat. Thus over timescales long enough for the MOC to adjust to equilibrium, the strength of the poleward heat flux by the MOC may be far more sensitive to tropical ocean temperatures (specifically, to hurricane potential intensity) than to thermodynamic conditions in regions of deep water formation. Of course, factors that alter the frequency of tropical cyclones would also change the poleward heat transport by the MOC.

5. Summary

Tropical cyclones mix enthalpy through the upper ocean, resulting in negative temperature anomalies that extend through an appreciable depth and affect large volumes of seawater. Over a period of several months, this cold water is restored to near climatological conditions by net enthalpy fluxes through the sea surface, while the warm anomaly at depth undergoes an advective adjustment to climatology. In the long-term mean the net surface enthalpy flux must be balanced by lateral enthalpy transport in the ocean. Thus tropical cyclones, by vertically mixing the upper ocean, induce an overturning circulation. It has been postulated by *Munk and Wunsch* [1998] that the existence of the MOC requires vertical mixing induced by tides and/or wind stress; here we argue conversely that vertical mixing and associated restoration of the surface temperature by net surface fluxes must result in an overturning circulation that transports heat away from the mixing zone.

By using a coupled air-sea tropical cyclone model to "hindcast" most of the observed tropical cyclones in one calendar year (1996), we have estimated the average net enthalpy flux through the sea surface necessary to restore the cold wakes of the storms to be $(1.4\pm0.7)\times10^{15} W$. This value is comparable to the observed peak meridional heat transport by the MOC as estimated by *Macdonald and Wunsch* [1996]. We infer that tropical cyclones may be an important and even a dominant mechanism for driving the thermohaline circulation. While the direct mixing induced by tropical cyclones cannot explain the ocean heat flux poleward of about the 17°C sea surface temperature isotherm, indirect effects owing to breaking of near-inertial oscillations deeper in the thermocline may drive lateral enthalpy fluxes farther poleward. We need to attain a better understanding of these secondary effects of tropical cyclones on the oceans and also to examine the effects of highly localized vertical mixing in a more realistic model that includes wind-driving of the ocean.

To the extent that tropical cyclones drive much of the poleward heat transport by the ocean, the Earth's climate will be sensitive to the processes that control global tropical cyclone activity. In particular, warming of tropical oceans will increase the potential intensity of tropical cyclones and with it the actual intensity of the storms, if the linear cumulative distribution function of storm intensity found in the present climate [*Emanuel,* 2000] continues to hold. Unless this is offset by a rather large decrease in storm frequency, the tropical cyclone-driven part of the thermohaline circulation will increase. We estimate that a 2°C increase in tropical ocean-mixed layer temperature would give rise to a 30% increase in the tropical cyclone-induced oceanic poleward heat flux, cooling the tropics and warming higher latitudes. Thus to the extent that tropical cyclones drive the MOC, they greatly reduce tropical climate sensitivity while, at the same time, increasing the sensitivity of climate at higher latitudes. This predicted aspect of climate sensitivity constitutes a hypothesis that can be tested using reconstructions of past climate change.

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