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Seasonal versus permanent thermocline warming by tropical cyclones

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[1] Recent studies suggest that the enhanced upper ocean mixing caused by tropical cyclones significantly contributes to the ocean heat transport. However, existing studies that try to quantify this contribution make the assumption that all heat pumped below the mixed layer by tropical cyclones is finally released in higher latitudes. Tropical cyclones occur primarily during summer and early fall, when the ocean mixed layer is generally shallow. As the mixed layer deepens in the following winter, any warm anomaly deposited within the seasonal thermocline will be reabsorbed by the mixed layer and lost to the atmosphere. Analysis of satellite sea surface temperature and sea surface height data, together with climatological subsurface ocean data, suggests that only about one quarter of the heat that is mixed downward by tropical cyclones eventually makes it into the permanent thermocline. **Citation:** 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.

1. Introduction

[2] Ocean heat transport (OHT) is crucial to maintaining the observed global climate. *Emanuel* [2001] suggested that ocean mixing driven by the strong surface winds in tropical cyclones (TCs) is an unappreciated contributor to the ocean heat transport. The argument is that TCs mix warm surface waters into the ocean interior which subsequently propagate poleward and contribute to the total OHT. However, it remains an open question as to whether the heat pumped into the ocean interior in the wake of TCs actually propagates towards higher latitudes.

[3] *Emanuel* [2001] estimated the ocean heat uptake in the wake of TCs by simulating all storms during the year 1996 using a simple axisymmetric hurricane model coupled to a bulk-Richardson number formulation for the ocean mixed layer (ML). The strong winds associated with TCs drive upper ocean mixing reaching far beneath the undisturbed summer ML, thereby cooling the ocean surface while leaving a warm anomaly below the depth of the pre-storm ML. Assuming that the near surface cold anomaly is entirely restored by surface fluxes, he estimated the average column integrated ocean heat uptake in the wake of TCs during the year 1996 to be 1.4 ± 0.7 PW. *Sriver et al.* [2008] repeated the calculation using observed SST anomalies from the Tropical Rainfall Measuring Mission's (TRMM) Microwave Imager (TMI) and employing climatological subsurface stratification data to estimate the depth to which the cold anomaly extends. They found an averaged TC-induced ocean heat uptake of about 0.48 PW for the years 1998–2005.

[4] The total OHT out of the tropics is about 3 PW [e.g., *Wunsch*, 2007]. If heat anomalies generated in the wake of TCs propagate poleward, the estimates by *Emanuel* [2001] and *Sriver et al.* [2008] imply that TCs contribute a large fraction of the total heat transport. *Pasquero and Emanuel* [2008] find that not all the heat pumped into the ocean by TCs contributes to OHT. They performed idealized simulations of the response of an ocean model to sudden mixing of the upper ocean over the regions where TCs occur. For a few months after the mixing event the ocean heated up, but only about half of the heat was transported out of the mixing region while the other half was lost back to the atmosphere within the same region. However, their model had no seasonal cycle, which is crucial to determining the fate of warm anomalies generated by upper ocean mixing in the real ocean.

[5] The goal here is to use a combination of observations and theory to study the fate of the heat anomalies in the months following the passage of a TC and thus to provide more accurate estimates of the net contribution of TC mixing to OHT. The ocean ML deepens in winter due to cooling of the atmospheric temperatures and stronger winds. Once the ML deepens, it entrains most of the heat anomalies generated by TCs the previous summer and returns their heat content back to the atmosphere. Detailed analysis of available observations show that most of the heat mixed by TCs into the ocean interior is directly lost back to the atmosphere in the following winter.

2. Data

[6] We estimate the net ocean heat uptake in the wake of TCs using a combination of satellite-derived sea surface temperature (SST), sea surface height (SSH), and monthly climatological data for ML depth and sub-mixed-layer stratification. SSH anomalies are derived from the SSALTO/DUACS multi-mission gridded product, which combines data from satellite altimeter missions Jason-1, Topex/Poseidon, Envisat, GFO and ERS-1 & 2 [*Dibarboure et al.*, 2009]. SSH is provided globally every 7 days on a $1/3° \times 1/3°$ mercator grid, but was interpolated on a $1° \times 1°$ lat-lon grid to match the resolution of the ML depth and stratification climatologies.

[7] SST is derived from the TRMM-TMI. TMI is particularly useful for this study because it operates at low

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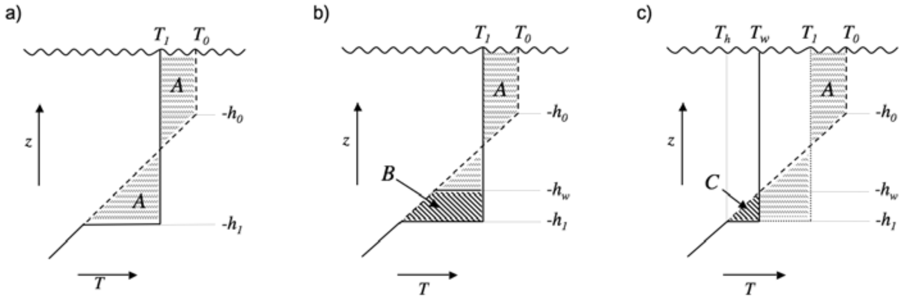

Figure 1. (a) Sketch of the upper ocean temperature profile subject to a tropical cyclone mixing event. The dashed line shows the temperature before, and the solid line shows the temperature after, the mixing event. T_0 and h_0 denote the SST and ML depth before the passage of the storm, T_1 is the SST after the storm, and h_1 is the depth reached by the storm-induced mixing. The shaded areas A are similar for a pure vertical mixing event in which the total heat content of the column is not changed. (b) The area B indicates the amount of heat that is mixed below the climatological winter ML depth h_w. (c) The area C indicates the fraction of the heat anomaly that remains after the winter. The winter SST minimum is denoted by T_w and T_h denotes the temperature at the mixing depth h_1. Stability requires that the ML deepens all the way to the mixing depth of the storm if $T_w < T_1$.

frequency and can see through the clouds that persist for multiple days in the aftermath of TCs. The SST data come as 3-day averages on a 1/4° × 1/4° grid, and were interpolated on the same 1° × 1° grid as the SSH and climatological data.

[8] ML depth and sub-mixed-layer stratification are taken from the OCCA reanalyzed climatology [*Forget*, 2010]. OCCA is obtained by assimilating observations into the MIT ocean model to ensure that the reanalyzed fields are dynamically consistent and the global heat, freshwater, and momentum budgets are closed. However, the main conclusions in the paper do not depend on the climatology used, as discussed in the auxiliary material.[1]

[9] Finally, TC tracks are taken from the Best Track data set, based on track information from the National Hurricane Center and the Joint Typhoon Warning Center [see *Emanuel*, 2001].

3. Estimating the Global Ocean Heat Uptake

[10] The large wind stresses associated with TCs cause very efficient mixing of the upper ocean, which often leads to spectacular cold SST wakes. Most of this cooling is associated with entrainment of colder water induced by vertical mixing, rather than by air/sea heat loss [e.g., *D'Asaro*, 2003] and results in surface cooling and subsurface warming (see Figure 1). The cold SST is subsequently restored to pre-storm values by anomalous air-sea heat fluxes (e.g. reduced latent heat loss), entailing a net positive heat flux into the ocean.

[11] If the near-surface cold anomaly is entirely restored by surface heat fluxes, the net ocean heating resulting from the passage of the TC is associated with the warm anomaly located below the base of the ML marked as the area "*A*" in Figure 1. The subsurface heating is equal to the surface heating necessary to restore the cold wake to pre-storm condition, because the cold surface anomaly is equal to the warm anomaly indicated by equal areas "*A*" in Figure 1a. This heating is given by,

$$OHU = -\frac{1}{\tau}\sum_i \int\int\int_{\Delta T_i<0} \rho c_p \Delta T_i(x,y,z)dzdxdy = \frac{1}{\tau}\sum_i \int\int \rho c_p A_i(x,y)dxdy, \quad (1)$$

where c_p and ρ are the heat capacity and density of seawater. ΔT is the temperature anomaly generated by the TC and we integrate in the vertical over the depth where $\Delta T < 0$, which gives the "area" marked as A in Figure 1. The horizontal integration is carried out over the wake of each storm and the sum is over all TCs (here defined as all tropical storms with maximum sustained wind speeds exceeding 17.5 m/s) over a time period τ. *Sriver et al.* [2008] used (1) and estimated the average TC-induced OHU during the years 1998–2005 to be about 0.48 PW.

[12] We repeat their calculation, assuming the simplified subsurface temperature profile sketched in Figure 1: ML of constant temperature, given by the observed pre-storm SST, extends to the ML depth inferred from the monthly OCCA climatology and lies above an upper thermocline with constant stratification. The stratification is estimated by a linear fit of the climatological subsurface temperature profile from the base of the summer ML to a depth of 125 m. Fitting a second order polynomial reveals that ignoring the curvature in the temperature profile introduces negligible systematic biases.

[13] The horizontal integral in (1) is computed along cross-sections at each 6-hourly TC location provided by the Best Track data set. The SST before (T_0) and after (T_1) the TC are estimated from TMI data using three-day averages before and after the crossing of the storm center. For TCs from 1998–2006, the OHU is estimated to be 0.58 PW, a result in reasonable agreement with *Sriver et al.* [2008] our estimate is slightly larger, likely because of a different assumption for the wake width (see auxiliary material).

4. Role of the Seasonal Cycle

[14] The above calculation assumed that the near surface cold anomaly generated by the passage of the TC is entirely

[1] Auxiliary materials are available in the HTML. doi:10.1029/2009GL041808.

restored by surface heat fluxes, while the warm anomaly below is permanently transferred into thermocline water. This is likely to be a considerable overestimate of the ocean heat uptake, because only water that escapes the ML in late winter is subducted into the permanent thermocline and contributes to the ocean heat uptake [*Stommel*, 1979]. Any warm anomaly found in the seasonal thermocline will be reabsorbed by the ML and lost to the atmosphere in the following winter. One can estimate what fraction of the warm anomaly generated in the wake of TCs is deposited at depths greater than the deepest winter ML depth h_w; this is given by the area B sketched in Figure 1b. Following the same approach described in section 3, but now integrating over B instead of A, using h_w from the OCCA climatology, we find that only about 0.30 PW or about half of the heat that is mixed downward by TCs is deposited in the permanent thermocline. This suggests that at least half of the heat pumped by TCs into the ocean in summer is returned to the atmosphere in winter.

[15] The above estimate, however, is mostly qualitative, because the deepening of the winter ML will itself be modified by the TC mixing event. A quantitative estimate requires a detailed heat budget of the ML between the time when the TC hits until the following winter. As a first step, we consider a one dimensional ML and ignore advection by the Ekman and geostrophic circulations, which are shown to be of secondary importance below. We also assume that the stability of the upper ocean is dominated by temperature and neglect the effect of salinity, a good approximation in the tropics. Under these approximations, the evolution of heat anomalies through seasons is sketched in Figure 1. Winter cooling will first return surface temperatures back towards T_1, i.e. the SST generated in the wake of the TC. If the surface temperatures drop below T_1, convection penetrates through the warm anomaly down to h_1, because that part of the water column is still unstratified. The progression of winter will therefore cool the whole warm anomaly down to the minimum winter temperature T_w. The fraction of the heat anomaly that is expected to remain in the permanent thermocline is given by the area C sketched in Figure 1c and is always less than the area B used for the qualitative estimate above. If $T_w < T_h$ no anomaly will survive the winter.

[16] The fraction of heat that remains in the permanent thermocline after the winter can be estimated as

$$OHU = \frac{1}{\tau}\sum_i \int\int \rho c_p C_i(x,y)dxdy\,,$$
$$C_i = \begin{cases} A & \text{for} \quad T_w > T_1 \\ \frac{1}{2}(T_w - T_h)^2\left(\frac{dT}{dz}\right)^{-1} & \text{for} \quad T_h \le T_w \le T_1\,, \\ 0 & \text{for} \quad T_w < T_h \end{cases} \tag{2}$$

where A is defined above, T_h is defined in Figure 1 and is calculated from T_0, T_1, h_0 and dT/dz. Using the same data and approach described in section 3 but now integrating over C_i, we find that $\simeq$0.14 PW, or about one quarter of the heat mixed downward by TCs, is expected to remain in the permanent thermocline after the winter.

[17] A full analysis of the evolution of the heat anomaly generated by TC winds must consider the effects of vertical mixing, Ekman pumping or suction, ML instabilities and advection by the large scale flow. The effect of vertical mixing and Ekman pumping can be studied with a simple one-dimensional ML model as proposed by *Price et al.* [1986] (hereinafter PWP). Simulations using the PWP model suggest that vertical diabatic mixing can increase the fraction of the heat anomalies surviving the winter, resulting into $\simeq$0.26 PW of heat uptake (see auxiliary material). This is because mixing partly restratifies the region that was homogenized by the TC so that the ML does not necessarily deepen as much as assumed in our estimate (2), and some heat can diffuse downwards into the permanent thermocline before the winter deepening of the ML sets in. Ekman pumping and suction are shown to have negligible effects within the precision of this study.

[18] In a full 3-dimensional picture, the lateral gradients associated with the localized temperature anomaly generated by the TC are dynamically unstable. As described by *Boccaletti et al.* [2007], isotherms will slump from the vertical to the horizontal as a result of baroclinic instabilities. The slumping results in a spreading and rising of the warm anomaly towards its neutrally buoyant level. Simple scaling arguments, presented in the auxiliary material, suggest that typical warm anomalies will rise by about 50 meters between the TC season and late winter as a result of slumping. Since most storms mix only about as deep or slightly deeper than the winter ML depth, a rise of $\simeq$50 m will bring the anomaly back into the winter ML, thereby strongly reducing the amount of heat that remains in the permanent thermocline. This effect is shown to likely dominate over vertical mixing.

[19] We conclude that the total amount of heat permanently deposited below the ML is $\simeq$0.15 PW. However due to uncertainties in the data (see auxiliary material) and the details of the ML dynamics, the uncertainty is on the order of 100%. Ocean heat uptake caused by TC-induced upper ocean mixing is thus estimated to be in the range of 0–0.3 PW, much smaller than previously calculated. To put this result in perspective, the annual average ocean heat uptake integrated over the tropics is on the order of 3 ± 1 PW [e.g., *Wunsch*, 2007] and the expected contribution of TC mixing is well within the error bars of this estimate.

5. Propagation of the Anomaly

[20] So far we assumed that the anomalies generated by TC-induced mixing remain in the same region until the onset of winter. This assumption is now tested by tracking SST and SSH anomalies generated in the aftermath of TCs on time scales of weeks to months. In addition to triggering irreversible mixing of the upper ocean, TCs drive a divergent surface Ekman flow associated with a net upwelling vertical velocity reaching deep into the main thermocline [e.g., *Greatbatch*, 1984]. The resulting (reversible) lifting of density surfaces results in a column integrated cooling and SSH depression in the center of the storm-wake, balanced by an opposite signal outside of the wake. Note that this reversible displacement of the thermocline does not influence the net ocean heat uptake.

[21] The SSH depression caused by a single TC is generally smaller than the background ocean variability. Hence we construct an ensemble average of SST and SSH anomalies in the wake of all TCs for the years 1998 to 2006

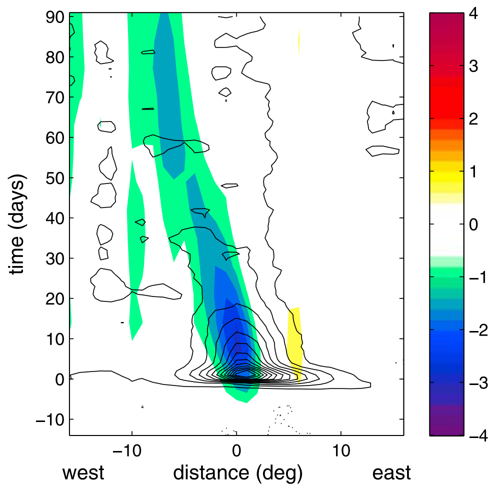

Figure 2. Hovmöller diagram showing the SSH (shading - in cm) and SST (contours - CI: 0.1 K) response to the passage of a tropical cyclone. Shown is a composite zonal section using all poleward propagating tropical cyclones with maximum sustained winds exceeding 17.5 m/s for the time period 1998 to 2006. Anomalies are calculated relative to the respective values before the passage of the storm and the seasonal cycle is removed. The shading/contouring shows only values which are statistically significant on a 95% level, as estimated from the background variability. For the SST, solid lines here indicate negative values, i.e., cooling, while dotted lines indicate positive values.

and between 40°S and 40°N. Figure 2 shows an ensemble mean zonal section through all storms moving within 45° of the poleward direction.

[22] We find pronounced SSH and SST responses on the order of 3 cm and 1 K, respectively. Note that anomalously strong TCs, the focus of many case studies [e.g., *Price*, 1981], can produce considerably larger SST anomalies, but the ensemble average contains many weak storms with no significant cold wakes. The SST cooling appears instantly after the passage of the storm. The SSH response seems to develop about one week after the storm, likely an artifact of the longer sampling period of the altimeter. The SST response is slightly biased to the east of the storm center, with the maximum cooling displaced by 50–100 km from the center. This is caused by the anticyclonic rotation of the wind stress on this side of the storm, which excites inertial oscillations and promotes stronger mixing [e.g., *Price*, 1983, and references therein].

[23] The SSH and SST depressions propagate westward at a speed of about 0.1°/day ≈0.1 m/s. As discussed in the auxiliary material, the propagation characteristics as well as the height of the SSH anomaly are in good agreement with the assumption that the observed depression is dominantly a baroclinic signal - Ekman upwelling driven by the TC winds lifts the thermocline resulting in balanced jets and eddies. The fact that the SSH and SST signals track together suggests that the disturbance is the result of a nonlinear eddy which, unlike a linear Rossby wave, transports material properties such as temperature. Analysis of a meridional section through the remaining, zonally moving, storms shows no clear indication of significant meridional movement (not shown). In summary the anomalies generated by storm events propagate primarily westward over a distance of at most 1000 km in a season. The anomalies therefore stay within the same climatological region until the following winter, supporting our previous analysis that neglected any propagation.

[24] It should also be noted that Figure 2 shows that the cold SST anomaly decays with an average e-folding time scale of about 1 week. This suggests that the restoring times of about 5 and 20 days found for hurricanes Fabian (2003) and Frances (2004) by *Price et al.* [2008], are representatives of the average TC cold wake, with Fabian being the somewhat more typical case. As discussed by *Price et al.* [2008], one expects only a shallow surface layer on the order of 10 m to be restored on this time scale. This in turn serves as a likely explanation for the fact that no significant thermal sea level rise is observed in Figure 2. If the restoring of the cold anomaly extended to a depth on the order of 50 m (a typical depth scale for the near surface cold anomaly), one would expect a thermal sea level rise on the order of 1.5 cm for a typical SST cooling of 1 K, but such a signal is not visible in Figure 2. The lack of such a signal confirms that the upper ocean cold anomaly is on average not fully restored by surface fluxes before the onset of winter, when the ML deepens and reabsorbs the subsurface warm anomaly.

6. Conclusions

[25] It was shown that the ocean heat uptake due to tropical cyclone upper ocean mixing has been systematically overestimated in the past. Previous studies assumed that the near surface cold anomaly is entirely restored by surface fluxes, while the warm anomaly beneath is left untouched. However, as the ML deepens in winter, not only the near surface cold anomaly but also the underlying warm anomaly gets reabsorbed by the ML, and its heat is released back to the atmosphere. It is estimated that between 0–0.3 PW eventually make it into the permanent thermocline and thus contribute to ocean heat transport.

[26] Several recent studies [*Emanuel*, 2002; *Korty et al.*, 2008; *Jansen and Ferrari*, 2009; *Fedorov et al.*, 2010] speculated that TC mixing might have played a more important role in the global heat budgets of past climates, such as the Eocene or Pliocene. This study shows that the influence of TC mixing on ocean heat uptake is largely limited by the mixing depth of the storms relative to the winter ML depth. The possible role of TCs in the ocean heat budget during past climates thus depends not only on the strength and frequency of TCs but, perhaps more importantly, on the depth of the winter ML during these climate epochs.

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