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## Global representation of tropical cyclone-induced ocean thermal changes using Argo data – Part 2: Estimating air–sea heat fluxes and ocean heat content changes

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## Abstract

We use Argo temperature data to examine changes in ocean heat content (OHC) and air–sea heat fluxes induced by tropical cyclones (TC)s on a global scale. A footprint technique that analyzes the vertical structure of cross-track thermal responses along all storm tracks during the period 2004–2012 is utilized (see part I). We find that TCs are responsible for 1.87 PW (11.05 W m<sup>-2</sup> when averaging over the global ocean basin) of heat transfer annually from the global ocean to the atmosphere during storm passage

- (0-3 days) on a global scale. Of this total,  $1.05 \pm 0.20 \text{ PW}$   $(4.80 \pm 0.85 \text{ Wm}^{-2})$  is caused by Tropical storms/Tropical depressions (TS/TD) and  $0.82 \pm 0.21 \text{ PW}$   $(6.25 \pm 1.5 \text{ Wm}^{-2})$
- <sup>10</sup> is caused by hurricanes. Our findings indicate that ocean heat loss by TCs may be a substantial missing piece of the global ocean heat budget. Net changes in OHC after storm passage is estimated by analyzing the temperature anomalies during wake recovery following storm events (4–20 days after storm passage) relative to pre-storm conditions. Results indicate the global ocean experiences a  $0.75 \pm 0.25$  PW (5.98 ±
- <sup>15</sup> 2.1 W m<sup>-2</sup>) net heat gain annually for hurricanes. In contrast, under TS/TD conditions, ocean experiences  $0.41 \pm 0.21$  PW  $(1.90 \pm 0.96$  W m<sup>-2</sup>) net ocean heat loss, suggesting the overall oceanic thermal response is particularly sensitive to the intensity of the event. The net ocean heat uptake caused by all storms is 0.34 PW.

## 1 Introduction

- <sup>20</sup> Observed variability in ocean heat content (OHC) has been shown to be a key indicator of climate change (Levitus et al., 2009, 2012). Recent observational studies have provided estimates of OHC trends (Domingues et al., 2008; Levitus et al., 2012; Cheng and Zhu, 2014). However, uncertainties exist within these records due to the sparseness of ocean observations, which is unevenly distributed in both space and time.
- <sup>25</sup> Because of these limitations, the current observation systems may miss sampling transient/severe weather events like tropical cyclones (TCs), which have been shown to re-



distribute heat and mass vertically and horizontally within regions experiencing storms, and perhaps on larger scales due to interactions with ocean circulations such as the meridional overturning circulation (Emanuel, 2001) or the subtropical overturning (Fedorov et al., 2010; Sriver et al., 2010). TCs also induce quick but strong ocean thermal changes by enhancing air–sea heat fluxes (Ginis, 1995; Mcphaden et al., 2009), ocean heat advection and ocean mixing (Emanuel, 1991). The role of TCs in the global ocean heat budget is still poorly understood.

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Previous efforts to quantify the impact of TCs on upper OHC have relied primarily on near-surface observations (Sriver and Huber, 2007; Sriver et al., 2008; Jansen et al.,

- 2010) and Altimetry sea level data (Mei et al., 2013), inferring a redistribution of heat via vertical mixing. Because there is a relatively firm theoretical understanding of the ocean's response to TC wind forcing (Price, 1981), global SST fields from satellites and reanalysis serve as the basis for first-order estimates of TC impacts, because they offer global coverage and relatively high spatial and temporal resolution. But these methods
- <sup>15</sup> come along with fundamental assumptions about the interior ocean response. For example, Sriver and Huber (2007) assume that the upper 50 m depth ocean is cooled homogeneously within TC cold wakes reflecting the temperature change observed at the surface, and all cooling is achieved through vertical mixing. Given these assumptions, they estimate 0.2–0.4 PW of heat is transported into the ocean interior globally.
- <sup>20</sup> Using improved methodologies, several additional observational studies have generally supported this result (Jansen et al., 2010; Sriver et al., 2008).

A key uncertainty missing in the above studies is the effect of enthalpy exchange at the air-sea interface (Emanuel, 1991, 1999). Typically these fluxes are estimated by measuring the humidity near the sea surface using mooring instruments, and calculat-

ing the latent heat flux according to a parameterized bulk equation. A so-called drag coefficient is needed to approximate the efficiency of heat transfer, which is one of the sources of uncertainties in the high-wind weather conditions (Powell et al., 2003). Another source of uncertainty arises due to the spatial and temporal variations in humidity, temperature and wind speed under highly chaotic wind conditions, since the winds un-



dergo rapid changes in direction and magnitude in TC conditions (Wright et al., 2001). Additionally, the effect of sea spray must also been taken into account, since it represents an efficient mechanism transporting enthalpy in the air–sea boundary layer under strong winds. These uncertainties make the direct measurements of the air–sea fluxes

difficult, and thus it is perhaps necessary to accumulate large numbers of measurements to achieve statistically significant results. Despite these limitations, studies have shown the air–sea heat fluxes within TCs are on the order of 2 to 3 times greater than the background heat fluxes in quiescent conditions (D'Asaro, 2003; Lin et al., 2009). These studies improve the understanding of the air–sea heat exchange by TCs, but
 global scale estimates are difficult.

Following the wind-drag based methodology, Trenberth and Fasullo (2007) use a high resolution model to examine the TC-induced water and energy budgets on the global scale. Their estimate of heat flux is based on an empirical relationship between heat flux and wind speed (Trenberth et al., 2007). They find that 0.17 PW heat is released from ocean to atmosphere within 400 km of the storm center and 0.58 PW within 1600 km.

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These previous studies suggest that TCs induce a substantial amount of heat exchange between the atmosphere, oceanic mixed layer, and upper thermocline, thus these events may play an important role in regulating global heat budgets and climate. However, comprehensive global estimates of air-sea exchanges, ocean interior thermal changes, and net effects on OHC have not yet been estimated using global, vertically resolved ocean observations.

The Argo system provides a new opportunity to achieve a global representation of TC-induced effects on the ocean temperatures and heat fluxes. These data are the primary source of in situ ocean observations since 2000. They are vertically resolved and capable of capturing ocean variability from intra-seasonal to decadal timescales (von Schuckmann and Le Traon, 2011; Willis et al., 2008). A preliminary study used Argo floats to estimate the net OHC changes by TCs in the northwestern pacific region (Park et al., 2011) and found a seasonal effect of the TC impact on upper ocean



temperature (i.e. TCs cause more cooling during warming seasons). On the contrary, subsurface OHC changes are not significant for weak storms compared to background variability. The authors attribute the lack of a subsurface response to a diminished role of entrainment in weaker TCs compared combined with enhanced air-sea heat fluxes

<sup>5</sup> and vertical advection. For strong storms, subsurface warming is generally consistent with mixed layer cooling (same magnitude and opposite sign).

In part I of this study, we present a new technique using ARGO data that utilizes a method aggregating global TC-induced ocean thermal responses into a Langragian footprint coordinate system. In brief, the footprint is created as a function of horizon-

- tal distance across the track (perpendicular to the storm's direction of motion), water depth, and elapsed time after the TC passed. We separate TCs into two categories: tropical storm/tropical depression (TS/TD) and hurricanes. We analyze the thermal response for two distinct time periods: the forcing stage (0–3 days relative to storm passage) and the recovery stage (4–20 days relative to storm passage). This tech-
- nique enables the characterization of the major sources of variability in the TC-induced ocean response, including cross-track variation, differences in storm intensity, and response time scale. We find this technique robustly captures the known characteristics of the vertical structure of the cross-track oceanic temperature response, and it provides a useful tool for examining basin-to-global representations of TC effects as a function of storm intensity.

Here we use the footprint method developed in part I to quantify OHC changes caused by TCs on a global scale. We focus on air-sea heat fluxes during storm passage and net ocean heat changes in the wakes of storms, relative to pre-storm conditions. The aim of the paper is to estimate observation-based OHC changes by TCs on

<sup>25</sup> basin and annual scales using vertically resolved Argo data and to explore implications for air-sea exchange and heat budgets. The paper is organized as follows. In Sect. 2, we present the method estimating the air-sea heat flux and ocean heat changes. We analyze the results of TC-induced air-sea heat fluxes in Sect. 3. In Sect. 4, we in-



2912

vestigate TC-induced net OHC changes after storm passage. The uncertainties of our results are analyzed in Sect. 5. We discuss the conclusions and implications in Sect. 6.

## 2 Methods

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In part I, we presented a technique that quantifies the average TC-induced ocean ther-<sup>5</sup> mal changes using a footprint strategy. We create a cross-track representation of the average oceanic thermal effects induced by TCs for all storms occurring globally between 2004 and 2012. The 3-dimensional cross-track footprint is a function of distance from the storm track center (dist), water depth (depth) and time after storm passage ( $\delta t$ ). Two different stages of ocean responses are examined including the forcing stage (0–3 days relative to storm passage) and restoring stage (4–20 days relative to storm passage). The functional form of the footprint is represented as:  $F_{\text{TSTD}}$  (dist, depth,  $\delta t$ )

for weak tropical cyclones (TS/TD conditions) and  $F_{Hur}$  (dist, depth,  $\delta t$ ) for strong tropical cyclones (or Hurricane conditions). We use this footprint method to characterize the ocean thermal response to TCs using global Argo data.

# **2.1** Estimation of ocean heat content changes during 0–3 days after storm passage

Since the air-sea exchanges in the TC affected regions within 0–3 days of the storm passage are dominated by TCs effects, which is several times larger than the background air-sea heat exchanges, it is reasonable to assume that the net ocean heat change within the TC-affected regions during this period is totally induced by TCs. Our choice of 3 days is to, at least partially, average out any float drift caused by inertial oscillations (see part I for more discussion).

This strategy is based on the notion that the ocean must be the energy source of a storm, so the total heat loss in the ocean during the storm is transported to the air as air–sea heat flux during the storm passage (0–3 days). The averaged OHC change is



calculated as follows (in W):

$$QA_{TSTD} = L_{track-TSTD} \left[ \int_{dist=-8}^{dist=8} \frac{1}{T_{3 days}} \int_{\delta t=0}^{\delta t=3 depth=1200} \rho c_{p} F_{TSTD}(dist, depth, \delta t) d_{depth} d_{\delta t} d_{dist} \right] / T_{year}$$

$$QA_{track} = QA_{track}$$

$${}_{5} L_{\text{track-Hur}} \left[ \int_{\text{dist}=-8}^{\text{dist}=8} \frac{\delta t=3 \text{ depth}=1200}{\int_{\text{dist}=-8}^{5} \int_{\delta t=0}^{5} \int_{\text{depth}=0}^{5} \rho c_{\rho} F_{\text{Hur}}(\text{dist, depth}, \delta t) d_{\text{depth}} d_{\delta t} d_{\text{dist}} \right] / T_{\text{year}}$$

where

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 $F_{\text{TSTD}}$ ,  $F_{\text{Hur}}$  – footprints of the track-averaged ocean responses obtained in part I of this study

 $c_p$  – heat capacity of sea water ~ 4186 J (kg °C)<sup>-1</sup>

 $\rho$  – density of sea water, which are calculated by using Argo salinity, pressure and temperature measurements before the storm

 $\delta t\text{-}$  elapsed time after the storm passage, which is averaged from 0 to 3 days after storm passage

dist - cross-track distance from the location of the Argo pair to the track, which is set

to 8° across the track

depth – vertical position of the measurement which is integrated from 0 to 1200 m  $T_{year}$  – duration of one calendar year in seconds, to calculate an annual mean

 $L_{\text{track-TSTD}}$ ,  $L_{\text{track-Hur}}$  – averaged track length within one year, about  $1.4 \times 10^8$  m,  $8.3 \times 10^7$  m for TS/TD and Hurricanes respectively, which are obtained by averaging the track length from 2004 to 2012.

We choose the time period between 0–3 days, because it likely captures the majority of the air–sea heat exchange during storm passage. However, other mechanisms may



also influence air-sea heat flux in this time period, such as storm induced cooling via mixing and wave generation. TC-induced surface cooling can cause a reversal of surface fluxes in the days following storm passage, which marks the transition between the forcing stage and the recovery stage. Some studies suggest fluxes may reverse sign

- <sup>5</sup> around 2 days after TC passage (Dare and McBride, 2011; Lloyd and Vecchi, 2011), The exact timing of this reversal depends on many factors, such as storm intensity, translation speed and regional conditions, and the best choice is unclear in a global context. Thus, we use 3 days as conservative estimate. We performed sensitivity tests of the TC response to different choices of the time length (0–2, 0–2.5, 0–3.5 and 0– 1 days) but the meeting (ast shows) and the meeting end intervention are sensed.
- <sup>10</sup> 4 days), but the results (not shown), and the results and interpretations are generally consistent for all time scales.

## 2.2 Estimation of air-sea fluxes during TC passage

We examine the air-sea heat transfer rate in the TC-affected regions, by averaging air-sea heat flux as follows (in  $Wm^{-2}$ ):

<sup>15</sup> 
$$H_{\text{TSTD}} = \text{QA}_{\text{TSTD}} / (L_{\text{track-TSTD}} \times R)$$
  
 $H_{\text{Hur}} = \text{QA}_{\text{Hur}} / (L_{\text{track-Hur}} \times R)$ 

where *R* is the cross-track size of the TC-affected region which is set to be  $16^{\circ}$  ( $\pm 8^{\circ}$  across the track). The other variables are consistent with QA<sub>TSTD</sub> and QA<sub>Hur</sub>.

## 2.3 Geographical distribution of air-sea heat fluxes

<sup>20</sup> To calculate the geographical distribution of the TC-induced air–sea heat fluxes, we bin the global ocean using 1° by 1° grid boxes. From 2004 to 2012, the air–sea heat flux from each TC is calculated in each grid box (denoted as *i* and *j* for latitude and longitude respectively):



$$H_{i,j} = \left(\sum_{\mathsf{ID}_1(i,j)}^{\mathsf{ID}_2(i,j)} H_{\mathsf{TSTD}} + \sum_{\mathsf{ID}_1(i,j)}^{\mathsf{ID}_2(i,j)} H_{\mathsf{Hur}}\right) \Big/ T_{9\,\mathsf{years}}$$

where  $T_{9 \text{ years}}$  is the duration of 9 years in years. If a grid box is affected by two or more individual storms at the same time, only the heat flux due to the stronger storm is included. This avoids potential double-counting of storm effects. Annual air–sea heat flux is calculated by calculating the 9 year average (2004–2012). In this calculation, we assume that the heat exchange is uniform over the TC-affected region.

#### 2.4 Estimation of net ocean heat content changes

We estimate the net OHC changes by examining the average temperature response between 4 to 20 after storm passage, referenced to pre-storm conditions. We choose 20 days as the maximum duration, because sea surface temperatures are typically restored by this time, and it is difficult to separate TC effects from seasonal signals on timescales greater than 3 weeks. Tests on the different choices of the time length are also conducted (4–18, 4–19 and 4–21 days). The results of these tests (not shown) suggest the magnitude of the TC signal is relatively insensitive to the choice of timescales.

We calculate OHC changes using (in W):

$$QN_{TSTD} = L_{track-TSTD} \left[ \int_{dist=-8}^{dist=8} \frac{1}{T_{17 \text{ days}}} \int_{\delta t=4}^{\delta t=20 \text{ depth}=1200} \rho c_{\rho} F_{TSTD}(dist, depth, \delta t) d_{depth} d_{\delta t} d_{dist} \right] / T_{year}$$



 $QN_{Hur} = L_{track-Hur} \left[ \int_{dist=-8}^{dist=8} \frac{1}{T_{17 \, days}} \int_{\delta t=4}^{\delta t=20 \, depth=1200} \rho c_{\rho} F_{Hur}(dist, depth, \delta t) d_{depth} d_{\delta t} d_{dist} \right] / T_{year}$ 

where  $\delta t$  is the elapsed time after storm averaged from 4 to 20 days,  $T_{17 \text{ days}}$  is the duration of 17 days in seconds. The other variables are the same to those in calculating QA<sub>TSTD</sub> and QA<sub>Hur</sub>.

#### 2.5 Methodological limitations

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Limitations of this methodology include:

- To create the geographical distribution of air-sea heat fluxes, we assume that ocean response to a storm is the same everywhere (based on the composite analysis of the footprints). This assumption neglects the regional differences in the ocean response due to differences in the background state and seasonal effects.
- 2. The net OHC changes induced by TCs are averaged within 4–20 days after storm passage, which represents the restoration stage. However, the ocean changes may not be fully restored during this time interval. As noted previously, we choose this time period because TC signals are difficult to separate from the background seasonal cycle on longer time scales.
- 3. The internal waves generated by TCs induce fluctuations in temperature, which could potentially bias our results. However, we hypothesize that these wave effects average to zero because we are using a large number of Argo pairs (~ 4410). We discuss potential biases further in the following section.



#### 3 Results and discussion

#### 3.1 Estimate of air-sea heat flux

Here we calculate a global estimate of air-sea heat exchange during TCs by integrating the ocean heat differences within storm-affected regions during a 3 day interval sur-

<sup>5</sup> rounding storm passage. We assume that during this period, the net column-integrated ocean heat loss is caused by heat transfer from the ocean to the atmosphere. We use the footprint methodology described in part I, which has two spatial dimensions: vertical depth and cross-track distance relative to the storm's direction of motion. The footprint averages over the along-track direction. Thus the footprint represents a 2-dimensional insulated box that is ±8° across the storm track relative to the storm center and 1200 m deep, with an opening at the air–sea interface. The heat exchange between the box and its surroundings occurs only at the surface.

To test whether assumptions about the footprint hold, particularly related to insulation from horizontal advection at the sides of the box and vertical heat exchange at the base, we calculate the box-averaged air–sea heat flux at different horizontal and vertical spatial scales, ranging from  $dx = \pm 1$  to  $\pm 15^{\circ}$  (with 0.5° increment) and dz = 100to 1900 m (with 200 m increment). As shown in Fig. 1, the averaged air–sea heat flux stabilizes for  $dx > \sim 6^{\circ}$ . The decreasing trend for  $dx > \sim 7^{\circ}$  is a "dilution" effect, which is caused by enlarging the box size while the OHC change within TC-affected region remains unchanged. Note this effect is generally linear for large dx, which is expected since the depth is held constant. In the vertical direction, the air–sea flux estimates are unchanged for dz greater than 700 m (corresponding to dx >~ 5°). This result suggests the method requires at least  $\pm 7^{\circ}$  and 700 m in order for the insulated box assump-

tion to be considered valid. Thus, we use a terminal depth dz = 1200 m based on the <sup>25</sup> availability of data in the upper ocean (a large portion of Argos stop near 1200 m), and  $dx = 8^{\circ}$ .

The annual contribution of TCs to the air–sea heat fluxes for  $dx = 8^{\circ}$  and dz = 1200 m is about ~ 4.80 and ~ 6.25 W m<sup>-2</sup> for TS/TD and Hurricanes, respectively. The positive



heat flux represents the net ocean heat loss. We calculate the total global air–sea heat exchange in Fig. 2. Given the methodology described previously, the integrated heat transport should converge to the total TC contribution as we increase the domain size. Consistent with Fig. 1, this convergence occurs for  $dx > \sim 6^{\circ}$ . For TS/TD, heat exchange

- <sup>5</sup> continues to increase to dx > 9. However, this increase is probably not caused by TCs given the large spatial scale. Therefore, we estimate the global annual air–sea heat exchange during TCs to be 1.05 and 0.82 PW for TS/TD and hurricanes, respectively. The total heat transfer is 1.87 PW annually, which represents the total heat loss from ocean to atmosphere during the forced stage (0–3 days).
- The global air-sea fluxes derived in Fig. 1 correspond to 584 W m<sup>-2</sup> for TS/TD and 10 761 W m<sup>-2</sup> for hurricanes, when averaging fluxes within storm-affected regions ( $\pm 8^{\circ}$ across the track) These values are consistent with previously published case study estimates, such as the mooring observations during the category-4 hurricane Nargis (Lin et al., 2009; Mcphaden et al., 2009), which estimated storm-induced air-sea fluxes of ~ 400–900  $Wm^{-2}$ . Furthermore, our global estimates are consistent to first order 15 with the estimated heat required to bring the troposphere into thermodynamic equilibrium (Emanuel, 1991) with the ocean  $\sim 10^8 \text{ Jm}^{-2}$ , which is equivalent to  $\sim 3 \text{ Wm}^{-2}$ annually over the global ocean basin. This estimate generally agrees with our results (4.80 and 6.25 Wm<sup>-2</sup> for weak and strong storm categories respectively). A recent observational study (Bell et al., 2012) shows that the mean TC enthalpy fluxes from 20 CBLAST field program increases from  $764 \text{ Wm}^{-2}$  at wind speeds of  $52 \text{ ms}^{-1}$  (category 3) to 2189  $\text{Wm}^{-2}$  at wind speeds of 72 ms<sup>-1</sup> (category 5) near the storm center. The result of the category 3 conditions is similar to our estimate averaging over all hurricanes (category 1 to 5). In addition, Braun, (2006) estimates a 1.34 PW heat loss from the ocean caused by hurricanes, which is  $\sim 0.52 \text{ PW}$  larger than our estimates. 25 Trenberth and Fasullo, (2007) estimates the TC-induced enthalpy exchange caused by hurricanes is about 0.58 PW in total for 1990-2005, ranging from 628, 703, 783 and 895 to 1019 Wm<sup>-2</sup> for categories 1 to 5 respectively, where the category-3 estimate is similar to our estimates averaged over all hurricane conditions. In total, their result is



about 0.24 PW smaller than our result for similar conditions. As noted previously, the three day averaging period used here may capture some of the recovery stage associated with a reversal of air-sea fluxes and the reheating of anomalously cold surface waters after storm passage, thus these results may be considered slightly conservative 5 compared to previous estimates (e.g. Braun, 2006).

The geographical patterns of the TC-induced air-sea heat fluxes are presented in Fig. 3, using spatial and temporal averaging consistent with our global estimates. We find significant spatial variability in the flux estimates, with the largest fluxes occurring in regions with the most TC activity (e.g. northwestern Pacific). These results indicate the

- <sup>10</sup> zonally averaged TC contribution to the total annual air-sea enthalpy flux budget may be as large as ~ 9.1 Wm<sup>-2</sup>, with peak value of 20 Wm<sup>-2</sup> at latitudes experiencing the most TCs. These fluxes could account for as much as ~ 10 % of the total annual ocean latent heat flux (90–110 Wm<sup>-2</sup>) (Trenberth et al., 2009) derived using NCAR/NCEP reanalysis (Kalnay et al., 1996), shown as the black curve in Fig. 3b.
- As a simple check of the TC contributions of net surface fluxes, we compare the results from Argo data to the background surface fluxes using the NCEP/NCAR reanalysis. Specifically, we calculate the net climatological air–sea heat fluxes along TC tracks using NCEP/NCAR reanalysis for the same Argo sampling criteria ( $dx = 8^\circ$ , dt = 0-3 days). However, these climatological fluxes represent the daily averages over
- <sup>20</sup> a 20 year period (1990–2012), rather than from specific TC days. In other words, the plot shows the surface fluxes along storm tracks for non-TC conditions. The NCEP/NCAR reanalysis product generally predicts a net oceanic heat uptake in background climatological conditions during TC seasons (Fig. 3c). In the absence of TCs, typical conditions would favor a net ocean heat uptake, on the order of 1 W m<sup>-2</sup> zonally
- <sup>25</sup> averaged across the global ocean. This background warming signal is of opposite sign to the TC effect, which tends to cool the ocean through enhanced surface fluxes during the forcing stage.

Figure 3a shows a prominent peak in air–sea heat flux in the Western Pacific Ocean, reaching values as large as  $65 \text{ Wm}^{-2}$ . Because this region experiences the most TC



activity, this peak is probably due to more TCs occurrences relative to other regions. In Fig. 4a and b, we present the frequency at each 1° by 1° grid box which is affected by TCs per year, given the affected regions (cross-track distance) defined to be  $\pm 1$  and  $\pm 8^{\circ}$  relative to storm center. As expected, the figure shows a higher frequency of TC

- <sup>5</sup> occurrences in grid boxes as we increase the size of the affected region. For example, in the western and eastern Pacific Ocean, between 1 and 1.5 storms pass directly through a single 1° by 1° grid box annually, but this activity can contribute to as much as  $12 \sim 14$  storms affecting the same grid boxes when we increase the TC-affected region to  $\pm 8^{\circ}$ .
- To check whether the peak in heat flux is caused by increased frequency of TCs, we assume that one grid box can be affected by only one storm within 20 days. The annual air–sea heat fluxes for this method are presented in Fig. 5, showing the peak fluxes in the northwestern Pacific decrease to 30 Wm<sup>-2</sup> while the overall fluxes in the other basins are relatively unaffected (to within ~ 5–10 Wm<sup>-2</sup>). Since the nature of air–sea heat exchanges is complicated by the close proximity of storms in active TC regions, such as in northwestern and northeastern Pacific, our geographical map of air–sea heat flux should be regarded as a first-order approximation.

#### 3.2 Estimates of net OHC changes after storm

Net OHC changes are estimated by calculating the difference between the average post-storm temperature during wake recovery (between 4 and 20 days after storm passage) and the pre-storm conditions. We choose 20 days as the upper limit of the post-storm temperature, because it represents the time scale of SST recovery after storm passage. Temperature anomalies are plotted in Fig. 6 as a function of the spatial and vertical extents of the TC footprint. Within 3° of the storm center, both TS/TD and Hur ricanes induce column-averaged cooling at all depths. The cooling effect decreases for increasing footprint size, approaching zero for TS/TD and a net warming for Hurri-

for increasing footprint size, approaching zero for TS/TD and a net warming for Hurricanes. This result suggests that upwelling and heat loss to the atmosphere near the storm center are partly (for TS/TD) or fully (for Hurricane) compensated by post-storm



surface fluxes. Furthermore, when the footprint size is larger than  $8-9^{\circ}$ , the absolute value of the average temperature anomalies decreases, which is again attributable to the "dilution" effect as discussed in the previous section. The temperature anomalies also converge for increasing depth, when the footprint is greater than 3°. Thus, we use

a cross-track length scale of ~ 8° and depth scale of 1200 m to quantify the TC-induced upper ocean thermal response. The average temperature anomalies for these footprint length scales are +0.039 °C for Hurricanes and -0.0125 °C for TS/TD.

The annual contribution of the net TC-induced changes in global OHC is calculated by multiplying the net ocean temperature changes with yearly averaged track lengths.

- <sup>10</sup> This method is based on our results that the averaged ocean thermal change over all storms between post-storm (after recovery) and pre-storm conditions is 0.039 °C and -0.0125 °C for Hurricanes and TS/TD, respectively. The positive values indicate heat gained by the ocean. These estimates correspond to global annual flux contribution of ~ 5.98 W m<sup>-2</sup> (0.75 PW) for Hurricanes and ~ -1.90 W m<sup>-2</sup> (-0.41 PW) for TS/TDs, where the positive values represent accepted post-store and post-time flux represent.
- <sup>15</sup> where the positive values represent oceanic heat convergence and negative flux represents the net oceanic heat loss. This implies that after hurricanes, the ocean keeps on warming, and recovers the storm-induced enthalpy flux during the storm passage.

These findings indicate the total TC contribution to the global ocean heat convergence is estimated to be 0.34 PW annually in 2004–2012 periods, which reflects a net <sup>20</sup> ocean heat gain from the atmosphere due to all storms.

Our results suggest that weak storms (TS/TD) tend to cool the ocean, while hurricanes tend to warm the ocean, when considering both storm-induced and post-storm fluxes. The difference in the ocean response may be due to the relatively weak vertical ocean mixing and surface cooling induced by TS/TD compared to hurricanes. For

<sup>25</sup> TS/TD, the OHC change is likely driven by the storm-induced enthalpy fluxes during passage. Because weaker storms typically cause less vertical mixing and thus less significant cold wakes following storms, there will be less post-storm heat flux into the ocean during the wake recovery stage. Conversely, strong events (hurricanes) induce more vertical mixing and surface cooling, which leads to more heat flux into the ocean



during the recovery stage and thus net oceanic heat convergence. This finding is generally remarkably consistent with recent results analyzing satellite altimetry (Mei et al., 2013), who show positive oceanic heat convergence for hurricanes (~ 0.37 PW globally). And also this result is consistent with those presented in Sriver and Huber (2007) and Jansen et al. (2010).

It is important to note that this estimate averages the post-storm temperature between 4 and 20 days after storm passage. As a test of this assumption, we can also define the post-storm restoration period to be when the OHC change is to zero. In other words, post-storm warming balances the storm-induced enthalpy flux. Our estimates suggest that the OHC restoring period for hurricanes is less than 20 days but more than 20 days for TS/TD.

#### 3.3 Uncertainties of the estimates

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In the previous two sections, we estimate the annual air–sea heat fluxes during 0– 3 days after storm passage and OHC changes during 4–20 days after storm passage relative to pre-storm conditions. Here we use a bootstrap technique to constrain the error bars and characterize the uncertainties of our heat flux and OHC estimates. Beginning with the total number of Argo float samples (4410 pairs), we randomly choose 90% of pairs and repeat our air–sea heat flux and anomalous OHC calculations, as described in the previous sections. We repeat the calculation 200 times.

- <sup>20</sup> In Fig. 7b and d, 200 estimates of air-sea heat fluxes presented as function of horizontal footprint size of the TC-affected regions (distance to the storm center). Most of these bootstrap estimates exhibit similar patterns with those shown in Fig. 1, supporting the robustness of our estimates. We choose an error bar of one standard deviation near 8° to quantify the uncertainty of our estimate. This uncertainty measure <sup>25</sup> is equal to  $\pm 0.85 \,\mathrm{W m}^{-2}$  for TS/TD and  $\pm 1.50 \,\mathrm{W m}^{-2}$  for Hurricanes, which is equiv-
- alent to ~ 20 and ~ 25 % of the fluxes for TS/TD and hurricanes, respectively. Or, in terms of global annual heat flux, this uncertainty equates to  $\pm 0.20$  PW for TS/TD and  $\pm 0.21$  PW for Hurricanes, which is equivalent to ~ 20 and ~ 25 % of the total estimates



for TS/TD and Hurricanes, respectively. Including these uncertainties, our estimates of air–sea heat flux during the TC forcing stage (0–3 days relative to storm passage) are:  $1.05 \pm 0.20$  PW ( $4.8 \pm 0.85$  Wm<sup>-2</sup>) for TS/TD and  $0.82 \pm 0.21$  PW ( $6.25 \pm 1.5$  Wm<sup>-2</sup>) for hurricanes.

Similarly, the 200 estimates of TC-induced OHC changes are shown in Fig. 8. We choose the uncertainty to be equivalent to the standard deviation of the average temperature change for spatial extent of 8° and 1200 m, consistent with the heat flux estimate. This uncertainty equates to ±0.0063 °C for TS/TD and ±0.0143 °C for hurricanes, which represents ~ 50 and 36 % of the estimated OHC changes for TS/TD and hurri canes, respectively. Considering this uncertainty, our estimates of TC-induced thermal changes are: -0.0125 ± 0.0063 °C for TS/TD and 0.0390 ± 0.0143 °C for hurricanes. Equivalently, these estimates correspond to global annual heat flux of -0.41 ± 0.21 PW (-1.90 ± 0.96 W m<sup>-2</sup>) for all TS/TDs, and 0.75 ± 0.25 PW (5.98 ± 2.1 W m<sup>-2</sup>) for all hurricanes, where the positive values denote a net oceanic heat convergence.

#### 15 4 Conclusion and discussion

We examine TCs' contribution to global annual air-sea heat flux and net OHC changes by using the ARGO observing system. We find that during the storm passage, the ocean generally experiences a net heat loss to the atmosphere through storm-induced enthalpy fluxes. Our observational results suggest that TCs contribute 11.5 W m<sup>-2</sup> (1.87 PW) heat in TC-affected regions annually from the ocean to the atmosphere within 0–3 days after storm passage. Of this total, weak storm (TS/TD) contribute 4.80 W m<sup>-2</sup> (1.05 PW) and strong storms (hurricanes) account for the rest. The uncertainty of our estimate is about 20 % for TS/TD and 25 % for hurricanes.

Recent in-situ, remotely sensed and reanalyzed air–sea heat flux products (Smith et al., 2011) have faced challenges in closing the ocean heat budget. These analyses show a net global oceanic heat gain of  $20-30 \text{ Wm}^{-2}$  (Josey et al., 1999), while the global mean net heat flux is ~  $0.5 \text{ Wm}^{-2}$  from observed variations in OHC. Our obser-



vational results suggest that TCs may provide a potential mechanism (heat flux in high wind regime) for filling this gap.

After storm passage, ocean conditions in TC-affected regions experience a recovery process to at least partially restore upper ocean conditions pre-storm or climatological values through enhanced air–sea fluxes leading to ocean heat convergence. This recovery stage lasts much longer than the forcing stage during the storm passage. We estimate the net changes in a time scale of 4–20 days relative to pre-storm conditions, which implicitly includes fluxes during the forced stage. On this time scale, around ~ 0.75 PW (5.98 W m<sup>-2</sup>) of heat is transferred annually from atmosphere to the ocean for hurricanes, which represents a net ocean heat gain after storms. However, TS/TD exhibit an opposite response, ~ –0.41 PW (1.90 W m<sup>-2</sup>), representing a net ocean heat loss for weaker events. We estimate the uncertainty to be about 50 % of our estimates for TS/TD and 35 % for Hurricanes. The opposite sign of net OHC changes after storm (4–20 days) for weak and strong storms implies the impact of these events on the upper

ocean is sensitive to the intensity. This result also suggests that additional atmospheric heating due to anthropogenic warming may potentially increase the rate of TC-induced ocean heat uptake, since research suggests the number of strong TCs may increase with continued warming (Bender et al., 2010; Knutson et al., 2010).

To assess the TC contribution to historical trends in the ocean heat uptake, we calculate the total air–sea heat flux in each year from 1970 to 2010, by assuming that each TS/TD transfers  $6.25 \text{ W m}^{-2}$  and each hurricane transfers  $4.8 \text{ W m}^{-2}$  heat from the TCaffected region to the atmosphere during 0–3 days after storm. The annual heat flux is shown in Fig. 9 in blue. The figure shows a maximum atmospheric heating ~  $12 \times 10^{22} \text{ J}$ during 1996–1997 and a generally larger signal between 1988 and 1998, which is due

to more TC activity during these years. As suggested in Trenberth and Fasullo, (2007), the large El Nino activity during these years (3 between 1990–1995 and a large event in 1997–1998) may be at least partially responsible this boost in activity in key TC regions (e.g. west Pacific).



Figure 9 also shows the accumulated TC-induced net OHC changes in recovery stages (4–20 days) relative to pre-storm conditions, which shows the net affect of storms on OHC. We assume the net OHC effect is -1.9 Wm<sup>-2</sup> for TS/TD and 5.98 Wm<sup>-2</sup> for hurricanes (positive value shows a net heat gain by the ocean). Net
OHC changes show that TC-induced ocean heat convergence is increasing since 1970. This OHC change is likely due to the increase in the fraction of strong storms during the past 40 years (Knutson et al., 2010). The linear trend of TC-induced ocean heat uptake is about 0.046 × 10<sup>22</sup> J year<sup>-1</sup>, which is 11 % of global ocean heat uptake of the upper-most 2000 m during the past 55 years (~ 0.42 × 10<sup>22</sup> J year<sup>-1</sup>) (Levitus et al., 2012).

In summary, the ocean response to TCs is complex. It is not a simple surface cooling and subsurface warming everywhere in TC-affected regions. It is highly variable, with upwelling/divergent currents near the storm center and down-welling/convergent currents in the outer regions (see part I of this study for full discussion), entrainment in the

<sup>15</sup> mixed layer, inertial oscillation of vertical/horizontal currents (Price, 1983), and maybe other differences in the response due to TC characteristics such as translation speed (Emanuel, 2007). In this study, we use global Argo data to provide first-order estimates of the global air-sea heat fluxes during the storm passage and net changes in OHC. Our results imply that TCs are an important component in the ocean system, providing

<sup>20</sup> a link between variability in air-sea heat flux and ocean heat uptake.

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**Figure 1.** Estimates of air-sea heat flux within TC with different footprint domain sizes (horizontal and depth) for: **(a)** TS/TDs and **(b)** hurricanes. The results of 1200 m are highlighted in cyan.





Figure 2. The impacts of domain size on global annual heat transfer from the ocean to the atmosphere by TCs for: (a) TS/TDs and (b) hurricanes. The colors are the same to those in Fig. 1.

Interactive Discussion



**Figure 3.** Geographical pattern of air–sea heat flux caused by TCs. (a) Globally integrated net heat flux caused by TCs calculated using Argo float data ( $W m^{-2}$ ). (b) Zonally averaged TC-induced heat flux (red curve), compared with the annual climatology (1990–2010) of air–sea latent heat flux (black curve) derived from NCEP/NCAR reanalysis (Kalnay et al., 1996). (c) Net surface flux (positive upward) along storm tracks for climatological conditions during the period 1990–2010 ( $W m^{-2}$ ), derived from NCEP/NCAR reanalysis using storm tracks from 1990–2010. The plot represents the background air–sea flux contribution to the Argo analysis using the 20 year daily climatology. (d) Zonal average of the climatological net surface heat fluxes shown in c.





**Figure 4.** Frequencies of tropical cyclones per year affecting 1° by 1° grid boxes, when the TC-affected region is assumed to be **(a)**  $\pm$ 1° from the track center, and **(b)**  $\pm$ 8° from the track center.





**Figure 5.** Geographical distribution of air–sea heat flux caused by TCs. We assume each grid box can only be affected by 1 storm within a 20 day window.





**Figure 6.** Column averaged temperature anomalies within 4–20 days after tropical cyclones, relative to pre-storm conditions, as a function of the horizontal box size across the storm track (from  $\pm 1$  to  $\pm 15^{\circ}$ ) for: **(a)** TS/TD, and **(b)** hurricanes. The colors are different vertical size of the TC-affected box from 100 to 1900 m. The results for the box with 1200 m depth is highlighted in cyan.





**Figure 7.** 200 estimates on heat transport (in **a** and **c**) and air–sea heat flux (in **b** and **d**) based on 200 randomly selected samples of pairs. (a) and (b) are estimates under TS/TD conditions and (b) and (c) for hurricanes. The mean of 200 estimates is highlighted in blue for TS/TD and in red for hurricane. Error bars represent the standard error.





**Figure 8.** 200 estimates (in cyan) of 0–1200 m column averaged temperature as a function of distance across the storm track, based on randomly sampling 90% of the Argo pairs for: (a) TS/TD and (b) hurricanes respectively. The mean and standard deviations are highlighted as the red line and error bars.







