LETTERS

Observational evidence for an ocean heat pump induced by tropical cyclones

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Ocean mixing affects global climate and the marine biosphere because it is linked to the ocean's ability to store and transport heat[1] and nutrients[2]. Observations have constrained the magnitude of upper ocean mixing associated with certain processes[3,4], but mixing rates measured directly[3,5] are significantly lower than those inferred from budget analyses[6], suggesting that other processes may play an important role. The winds associated with tropical cyclones are known to lead to localized mixing of the upper ocean[7–9], but the hypothesis that tropical cyclones are important mixing agents at the global scale[10] has not been tested. Here we calculate the effect of tropical cyclones on surface ocean temperatures by comparing surface temperatures before and after storm passage, and use these results to calculate the vertical mixing induced by tropical cyclone activity. Our results indicate that tropical cyclones are responsible for significant cooling and vertical mixing of the surface ocean in tropical regions. Assuming that all the heat that is mixed downwards is balanced by heat transport towards the poles, we calculate that approximately 15 per cent of peak ocean heat transport may be associated with the vertical mixing induced by tropical cyclones. Furthermore, our analyses show that the magnitude of this mixing is strongly related to sea surface temperature, indicating that future changes in tropical sea surface temperatures may have significant effects on ocean circulation and ocean heat transport that are not currently accounted for in climate models.

The ocean circulation is complex, time-dependent and three-dimensional, but it is generally accepted that the meridional overturning circulation (MOC) is a key climate feature because it is linked with ocean heat transport (OHT), and the MOC is itself maintained by mechanical diapycnal mixing[1,11]. Below the near-surface mixed layer and above the main thermocline base, this mixing is usually driven by shear instability[4,12]. Internal waves, driven either by winds or tides, are the main sources of shear[1]. As these waves break, they generate most of the upper ocean's diapycnal mixing[4]. Winds drive at least half of the total mixing[1], with some studies giving a leading role to wind-driven mixing in the Southern Ocean[13] and some preferring a tropical location[14]. Both sources of mixing may be important. Budget analyses require 16 and 20 Sv ($1\,\mathrm{Sv} = 10^6\,\mathrm{m^3\,s^{-1}}$) of diapycnal cross-thermocline upwelling to occur within the tropical-to-midlatitude Pacific[15] and Atlantic[16], respectively, in addition to substantial Southern Ocean mixing. Furthermore, tropical upper ocean mixing is considered especially important to climate because simulations and theory agree that this region of strong stratification is the most efficient place for mixing to drive strong heat transport[14,17]—climate is especially sensitive to mixing variations in this region[18].

The magnitude and causes of upper ocean mixing are not well established, and the spread of values between many low measured values and those required by macroscopic balance requirements remains large[1]. Theory and microstructure-based measurements agree in that they both suggest an extremely weak low latitude mixing coefficient (0.005–$0.05\,\mathrm{cm^2\,s^{-1}}$), but the values are generally higher and range by a factor of 30 in subtropics and midlatitudes[3]. Upper ocean tracer release experiments generally agree with these estimates, showing a vertical mixing coefficient of $\sim$0.1–0.2 $\mathrm{cm^2\,s^{-1}}$ (ref. 5) in most locations but also with substantial variability[19]. Other approaches using large-scale energy budget requirements[6] indicate time-mean and basin-wide diffusivities at the higher end (0.1–1.5 $\mathrm{cm^2\,s^{-1}}$). Upper ocean background diffusivities of $\sim 0.1\,\mathrm{cm^2\,s^{-1}}$ are required in global ocean models. Reconciling these different approaches and results is difficult, but differences in approach between the methods provide insights. Upper ocean studies that have only been carried out for brief intervals (days to weeks) or that have sampled the oceans sparsely (the majority of microstructure and dye tracer studies) tend to show smaller values. Budget studies that integrate over years and over entire water masses tend to show higher values, suggesting that differences between these approaches might be due to spatially or temporally localized phenomena. Consequently, transient and localized strong sources of internal waves may provide a resolution to remaining discrepancies. Recent results indicate that transient, localized mixing—even in the tropics—may contribute significantly to the MOC, as long as the mixing happens near ocean boundaries[20].

Tropical cyclones (hereafter 'cyclones') are transient events, and their high wind speeds generate strong, near-inertial internal waves, making them efficient upper ocean mixers[7]. This diapycnal mixing is evidenced by the prominent cold wakes seen along, and to the right of, storm tracks, with sea surface temperature (SST) depressed by up to 8 °C (Supplementary Fig. S1). The wakes are primarily (>75%) due to upward mixing of cold water caused by the breaking of inertial waves that entrain water from the mixed-layer base[7–9]. This mixing is known to transport heat downwards[10,21] and nutrients upwards[2]. Cold wakes last days to weeks, and surface conditions are restored to climatologically normal conditions through surface fluxes[10]. The downward-mixed heat anomaly persists, and the full vertical column experiences a net warming[10,21].

This downward pumping of warm water by cyclones represents a heat convergence—that is, a net increase in ocean heat content (OHC). In steady state, such heat convergence should be balanced by a lateral transport of ocean heat out of the storm-affected region. Emanuel[10] used observations of cyclone cold wakes, together with model simulations using a simple coupled, mixed layer ocean model, to estimate cyclone-induced OHT for 1996. His estimate is 1.4 petawatts ($1\,\mathrm{PW} = 10^{15}\,\mathrm{W}$) with a total uncertainty of ± 0.7 PW, which is sufficient to account for the peak observed OHT in the subtropics. Emanuel[22] also hypothesized that the integrated intensity of cyclone events might be linked to the mean climate state by the potential relationship between cyclone activity and SST[23,24]. Thus, feedbacks could exist between cyclones, ocean heat convergence and transport, profoundly altering the behaviour of the climate system with higher-than-modern temperatures.

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We test the key elements of this hypothesis by comparing surface temperatures before and after storm passage, using reanalysis data. The methodology builds on the techniques described in ref. 24 (see Methods). We find that cyclones regularly cool the tropical oceanic mixed layer (Fig. 1a), generate strong vertical mixing (Fig. 1b), and pump downward the heat lost from the mixed layer (Fig. 1c). These processes are a strong function of tropical SST (Fig. 2). The tropical SSTs are depressed by up to several degrees climatologically where tropical cyclone activity is greatest (Fig. 1a). This temperature anomaly pattern is found in multiple surface temperature data sets (Supplementary Fig. 2). Phrased as a $W\,m^{-2}$ forcing, the upper ocean is cooled by ~1–3 $W\,m^{-2}$ (Fig. 1a) annually, assuming that the SST anomaly penetrates to 50 m depth. Interestingly, the locations for this heat input are well correlated with observed OHC anomalies, suggesting a positive correlation between upper ocean heat content changes, cyclone activity, and cyclone-induced heat pumping (Fig. 1c), as required by Emanuel's hypothesis.

Using observed, spatially resolved, cyclone-season averaged, ocean vertical temperature profiles[25], we estimate the depth of cyclone-induced mixing (Supplementary Fig. S3). This mixing depth length scale is used to express the cyclone-induced cooling as an effective vertical diffusivity (Fig. 1b) (see Methods). We use diapycnal, vertical and diathermal diffusivity interchangeably because they are generally equivalent in the regions of interest[1,6]. This map of annualized vertical diffusivity clearly shows maxima in regions with substantial cyclone activity, and much of the mixing occurs near western boundaries. The values vary from close to the minimum values observed from direct measurements (~0.05 $cm^2\,s^{-1}$) and up to much higher values (>1 $cm^2\,s^{-1}$) in the western Pacific. Nearly all tracer and microstructure studies have avoided measuring cyclone-influenced conditions. The one published measurement for cyclone-influenced conditions is 10 times the undisturbed conditions measured at that location[19], and agrees with our calculation. The lack of cyclone-induced mixing near the Equator is consistent with weak mixing noted there[3].

Comparison with integral budget-based values[6] reveals remarkably close agreement, both in terms of inferred global, tropical mean diffusivity and their sharp increases over the warmest ocean water masses (Fig. 2a). The close agreement of these data sets can be parsimoniously interpreted to indicate that much or possibly all of the upper ocean mixing observed in these analyses[6] is attributable to cyclone activity. These results support the conjecture that tropical uppermost ocean mixing is quite weak except during the occasional cyclone mixing event[12]. This could have serious implications for climate models, which represent this mixing as a constant 'background' diffusivity, and also for other analyses dependent on vertical mixing, such as models for primary production.

We calculate cyclone-induced OHC changes by assuming that surface temperature anomalies indicate uniform cooling to 50 m depth, and that this heat lost from the mixed layer is pumped downward. We represent the annualized increase in OHC as the steady-state cyclone-induced global OHT, assuming that all heat converged must, in steady state, become a poleward transport. This value for OHT estimated from reanalysis is 0.26 PW, with peak values greater than 0.50 PW (Supplementary Fig. S4). Our results indicate that ~15% of peak global OHT may currently be directly related to cyclone-derived mixing. This value is comparable to the total transport in the Atlantic Ocean past 50° N (ref. 26). These values are

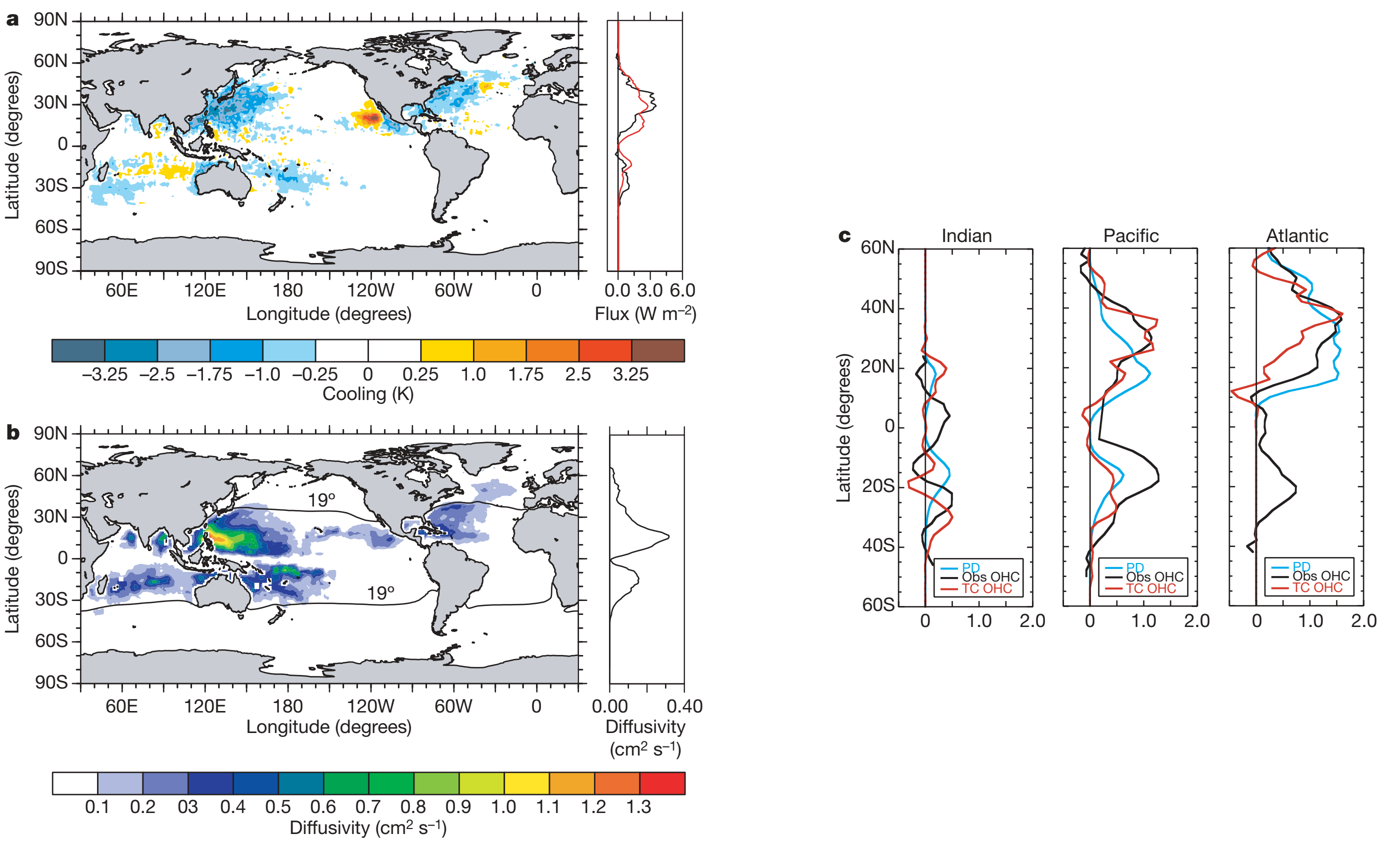

Figure 1 | Maps of tropical cyclone effects on the upper ocean. **a,** Average cooling for the annually accumulated, cyclone-induced temperature anomalies derived from ERA-40 2 metre air temperature (2MT; 1982–2001). Inset, the zonally averaged fluxes needed to restore the heat anomalies over one year, assuming the anomalies are 50 m deep (black line, 2MT; red line, SST; see Methods and Supplementary Information). **b,** Annualized average of vertical diffusivity attributable to cyclone mixing (see Methods). Contour represents the 19 °C isotherm, the margin of the 'warm water sphere'. Inset, the zonally averaged diffusivity. **c,** Annually averaged zonal-mean cyclone power dissipation (PD; blue curves), observed OHC anomalies from IGOSS (Obs OHC; black curves), and tropical-cyclone-induced OHC anomalies (TC OHC; red curves) for the period 1994 to 2001. Observed OHC curves are divided by $4 \times 10^8\,J\,m^{-2}$. Dissipation and cyclone OHC are normalized by the respective observed OHC for each basin.

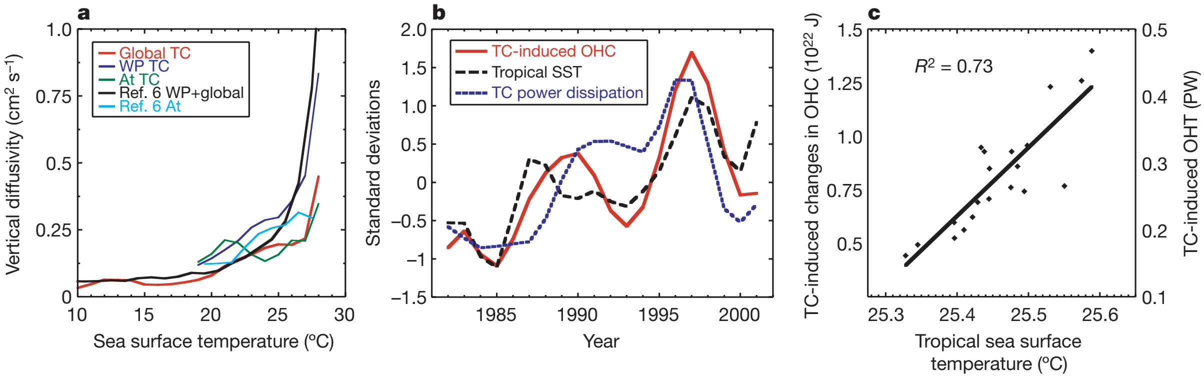

Figure 2 | Potential cyclone-induced climate interactions with vertical ocean mixing. a, Effective vertical diffusivity versus SST for global tropical-cyclone mixing (Global TC; red line), western north Pacific tropical-cyclone mixing (WP TC; dark blue line), and north Atlantic tropical-cyclone mixing (At TC; green line)—compared to global (WP+global; black line) and Atlantic (At; light blue line) diffusivity observations from ref. 6. **b**, Annual tropical-cyclone-induced OHC changes derived from ERA-40 2 metre air temperature (2MT) (solid, red curve), mean annual tropical SST from ERA-40 (dashed, black curve), and globally integrated tropical-cyclone power dissipation (PD) from ERA-40 near surface winds (dotted, blue curve). All data have been 5-yr low-pass-filtered, and are normalized by their respective standard deviations. **c**, Scatter plot of tropical-cyclone-induced OHC anomalies (left axis) and the equivalent OHT (right axis) derived from ERA-40 2MT and mean annual tropical SST from ERA-40. Both quantities are filtered as in Fig. 2b.

significantly smaller than early estimates[10], but are consistent with recent modelling work[27].

More important than this quantity's current mean value is its potential climate sensitivity, because previous studies observed correlations between increasing cyclone activity and rising SST[23,24]. Figure 2b shows the time series from 1982 to 2001 of globally integrated and annually averaged quantities as follows: cyclone-induced increases in OHC, tropical SST, and cyclone power dissipation (an integrated measure of cyclone wind intensity, hereafter referred to as dissipation). Tropical SST and globally integrated annual OHC/OHT correlate well, with $r^2 = 0.73$ (Fig. 2c). This agreement is consistent with theory[22], and suggests that cyclones may play a key role in climate dynamics. Close agreement also exists between integrated dissipation and cyclone-induced OHT ($r^2 = 0.57$, Supplementary Fig. S5), which is consistent with the hypothesis that increases in the intensity, frequency or duration of intense storms should lead to more extensive, vigorous mixing and more pronounced cold wakes. A 0.25 °C increase in mean annual tropical SST may lead to a 60% increase in global dissipation[24]—we show here that this SST increase may lead to a tripling of cyclone-induced global OHT (Fig. 2c).

As a consistency check on these dissipation calculations, we compute the power available for turbulent mixing and compare this with independent estimates. The average total reanalysis-derived dissipation observed through this interval (3.3×10^{19} J) corresponds to 1×10^{12} W expended on turbulent mixing. Using different methods, the global contribution of cyclones to the ocean's near-inertial spectral power range is calculated to be $\sim 7 \times 10^{11}$ W (ref. 28), consistent with our estimate. This is fully half of the total mixing required[1] to balance 30 Sv of deep-water formation, so it appears that mechanical stirring by cyclones may be responsible for about half of what is commonly called the thermohaline circulation. Furthermore, the results are consistent with—and provide a physical explanation for—water-mass budgets requiring tropical cross-thermocline mixing[15,16]. The importance of the cyclone-induced contribution has hitherto gone unnoticed in some previous studies[11], probably because wind fluctuations with timescales less than 5 days were not considered—hence the cyclone contribution was filtered out. A preliminary analysis indicates that even wind data sets with daily resolution that are used to drive ocean models under-represent the integrated wind intensity by a factor of 10, so no current simulations are likely to accurately capture this highly non-linear mixing process. This may partially explain a frequently noted tendency of ocean models to produce overly shallow mixed layer depths.

Our analysis suggests that changes in global cyclone frequency, duration and/or intensity are closely related to the amount of heat pumped into—and available to be subsequently transported by—the oceans. This relationship may have implications for changes in heat transport associated with past and future climate change. Extrapolation of our results suggests that future increases in tropical temperatures may result in increased dissipation, mixing, heat storage, and eventually heat transport. Moreover, this positive response in transport might feed back on climate by redistributing heat poleward, diminishing the Equator-to-pole temperature gradient, and raising global mean temperature[29]. We have provided some evidence that cyclone-induced mixing is a fundamental physical mechanism that may act to stabilize tropical temperatures, mix the upper ocean, and cause polar amplification of climate change. It is not included in the current conceptual or numerical models of the climate system. Better representation of cyclone winds and the associated mixing in climate models may help to explain the still-vexing questions posed by past climates[30].

METHODS

Data. We primarily show results for two-metre air temperature (2MT) from the European Centre for Medium-Range Weather Forecasts Reanalysis Project (ERA-40) after 1981. Additional results from other data sources are discussed in the Supplementary Information. We explicitly assume that the cyclone-induced temperature depression is due entirely to vertical mixing (see Supplementary Information). We use cyclone tracks from the 'best track' data sets (see ref. 24 for details) and SSTs at 6 h intervals over a 6° × 6° footprint centred on the storm's eye.

Calculation of cyclone-induced temperature anomalies. To calculate anomalies, final temperatures are taken at each location 3 days after storm passage for 2MT fields and 7 days after for SST and skin temperature (SKT) fields; final values are subtracted from the initial temperature conditions 3 days before the storm for all fields. Separate anomaly timescales are used for measuring 2MT, SST and SKT because the characteristic variability of each variable within the reanalysis data sets varies. SST data after 1981 within reanalysis are based on weekly averages, so that a longer timescale is needed in order to capture the cyclone-induced anomaly compared with 2MT. On the other hand, 2MT is more variable owing to atmospheric processes, and we use a shorter timescale for reanalysis fields for 2MT anomalies (6 days) with respect to SST anomalies (10 days).

Calculation of diffusivity attributable to cyclone mixing. We calculate the effective vertical diffusivity, k_v, as $k_v = L^2/\tau$, where L is a depth scale over which mixing occurs and τ is a characteristic timescale over which the mixed layer deepening and entrainment occur. We use 'vertical' here, but diapycnal or diathermal would be equivalent statements, as described earlier. To calculate the effective vertical diffusivity attributable to cyclone mixing (Fig. 1b), we begin by assuming that all mixing in a given year is achieved during the single largest cooling event, that is, the largest cyclone-induced temperature anomaly at each location is indicative of all the mixing at that location during a given year. This assumption appears reasonable, given the small diffusivities measured in tropical regions[3] in undisturbed conditions, and is consistent with the observation that all effective mixing in these locations is attributable to a few strong mixing events per year[12].

We estimate L using storm-season-averaged vertical temperature profiles ($\partial T/\partial z$) from ref. 25 in tandem with the SST anomalies (ΔT) from ERA-40. This technique assumes that the anomaly measured at the surface reflects well-mixed conditions down to some depth. We define L at each location as the level from which upwelling needs to occur in order to achieve the observed surface temperature anomaly, $L = \Delta T(\partial z/\partial T)$. We estimate that τ is $\sim$24 h, based on analysing individual storm track anomalies using satellite-based estimates with high temporal resolution (available at www.ssmi.com).

Calculation of cyclone-induced changes in OHC. The vertically integrated heat anomaly, Q, is calculated, adapting the formalism of Emanuel[10], as:

$$Q = \iiint F\rho C \Delta T \mathrm{d}h \mathrm{d}W \mathrm{d}S$$

where F is the fraction of heat transported downward from the oceanic mixed layer, ρ and C are respectively the density and heat capacity of sea water, ΔT and dh are respectively the magnitude and the depth of the temperature anomaly, and dW and dS are respectively the cross-track length and the along-track length of the storm wake.

For all calculations, ρ and C are held constant and equal to 1,020 kg m^{-3} and 3,900 J kg^{-1} °C^{-1}, respectively. To simplify the depth of the vertical mixing (dh), we assume the depths of all heat anomalies are constant and equal to 50 m. This value for dh is likely to be an underestimate for strong storms, for which vigorous vertical motions have been observed and modelled down to depths of 200 m. We assume $F = 1$, and thus, all heat lost from the oceanic mixed layer is transported downward and ultimately poleward.

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