Evolution of Ocean Heat Content Related to ENSO

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(Manuscript received 13 September 2018, in final form 28 February 2019)

ABSTRACT

As the strongest interannual perturbation to the climate system, El Niño–Southern Oscillation (ENSO) dominates the year-to-year variability of the ocean energy budget. Here we combine ocean observations, reanalyses, and surface flux data with Earth system model simulations to obtain estimates of the different terms affecting the redistribution of energy in the Earth system during ENSO events, including exchanges between ocean and atmosphere and among different ocean basins, and lateral and vertical rearrangements. This comprehensive inventory allows better understanding of the regional and global evolution of ocean heat related to ENSO and provides observational metrics to benchmark performance of climate models. Results confirm that there is a strong negative ocean heat content tendency (OHCT) in the tropical Pacific Ocean during El Niño, mainly through enhanced air–sea heat fluxes Q into the atmosphere driven by high sea surface temperatures. In addition to this diabatic component, there is an adiabatic redistribution of heat both laterally and vertically (0–100 and 100–300 m) in the tropical Pacific and Indian oceans that dominates the local OHCT. Heat is also transported and discharged from 20°S–5°N into off-equatorial regions within 5°–20°N during and after El Niño. OHCT and Q changes outside the tropical Pacific Ocean indicate the ENSO-driven atmospheric teleconnections and changes of ocean heat transport (i.e., Indonesian Throughflow). The tropical Atlantic and Indian Oceans warm during El Niño, partly offsetting the tropical Pacific cooling for the tropical oceans as a whole. While there are distinct regional OHCT changes, many compensate each other, resulting in a weak but robust net global ocean cooling during and after El Niño.

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-18-0607.s1>.

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DOI: 10.1175/JCLI-D-18-0607.1

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1. Introduction

El Niño–Southern Oscillation (ENSO) is the dominant mode of air–sea interaction in the climate system on interannual time scales (Trenberth et al. 2002, 2005). In the ocean, ENSO is associated with a strong redistribution of heat, especially in the Pacific Ocean (Wyrski 1975, 1985). The “delayed oscillator” and “recharge–discharge oscillator” paradigms for ENSO suggest that the equatorial ocean accumulates heat prior to El Niño (as a precursor to El Niño) and depletes it during and after El Niño by moving heat away from the equatorial region (Cane and Zebiak 1985; Zebiak and Cane 1987; Wyrski 1985; Jin 1997; Schopf and Suarez 1988). The delayed-oscillator paradigm emphasizes the important role of oceanic wave processes, and the recharge–discharge oscillator focuses more explicitly on the changes in upper ocean heat content. The recharge–discharge oscillator has also been suggested to set the time scale of ENSO (Jin 1997). Sun (1997) further suggests that the very existence of El Niño events may be due to the intensity of the tropical radiative heating.

The heat exchange suggested by the recharge–discharge oscillator or delayed oscillator is largely adiabatic because heat is advected within the ocean between the equatorial region and elsewhere. However, there is also a substantial diabatic heating component (Sun and Trenberth 1998; Sun 2000; Trenberth et al. 2002). During an El Niño event, heat is moved eastward in the equatorial Pacific Ocean and then poleward along the Americas and into the atmosphere, resulting in a discharge of the heat from the ocean (Trenberth et al. 2002). Subsequently then, there has to be a recharge of the heat within the ocean as a whole and then into the tropical ocean. Many theories of ENSO take less account of this diabatic component and the important details of exactly how the redistribution of heat occurs, both between the ocean and atmosphere, and also among the tropical oceans via the atmosphere (Mayer et al. 2014, 2016). A central purpose of this paper is to illuminate these aspects and thereby build a better framework for evaluating models and predicting ENSO evolution.

The accumulation of heat in the western tropical Pacific prior to El Niño events (during La Niña events) is accompanied by higher than normal sea levels associated with a deeper than normal thermocline, an indication of heat also being redistributed vertically in the Indo-Pacific basin. Meinen and McPhaden (2000) and McPhaden (2012) document how the ocean heat content (OHC) in the upper layers leads the central equatorial Pacific SST anomalies in the development of El Niño, and how that lead has shortened since 2000. There is a net heat increase in the upper ocean and a decrease

below 100-m depth in the Pacific Ocean during El Niño events (Roemmich and Gilson 2011; Mayer et al. 2014; Piecuch and Quinn 2016; Trenberth et al. 2014a, 2016; Wu et al. 2019). These opposing vertical changes are also evident in the global average of ocean temperature change (Roemmich and Gilson 2011; Cheng et al. 2015; Trenberth et al. 2016). Hence, the net vertically integrated heat content change is relatively small owing to these offsetting effects, and this aspect has yet to be quantified with multivariate observational records and will be quantified in this study.

As well as in the tropical Pacific Ocean where ENSO originates, impacts occur on a global scale with ENSO via teleconnections (Trenberth et al. 1998, 2002, 2014a,b; Trenberth and Fasullo 2017; McPhaden et al. 2006; Liu and Alexander 2007; Yang and DelSole 2012; Cazenave et al. 2014; Sprintall et al. 2014; Forget and Ponte 2015; Yeh et al. 2018). The tropical atmospheric heating can significantly regulate the large-scale atmospheric circulation and produce a global signature (Alexander and Scott 2002; Trenberth et al. 2002; Wang 2004; Capotondi et al. 2015; Fasullo et al. 2018). These teleconnections are mainly driven by the ocean heat loss to the atmosphere in the tropics (via evaporation): warming oceans moisten and heat the air so that convection breaks out and large-scale convergence ensues, leading to increases of precipitation P in the tropical central and eastern Pacific Ocean (Trenberth 2011). As well as driving anomalous regional overturning circulations such as the Walker cell, the anomalous latent heating and associated upper-tropospheric atmospheric divergence create a planetary-scale Rossby wave source that alters the jet stream and storm tracks, especially in the winter hemisphere (Alexander and Scott 2002; Trenberth 2011).

ENSO also regulates the surface winds and the general ocean circulation. In the tropics, the currents are significantly changed related to ENSO (Johnson et al. 2000; Roemmich and Gilson 2011; Wu et al. 2019). The Indonesian Throughflow (ITF) appears to be stronger during La Niña, influencing the heat budget in both Pacific and Indian Oceans on interannual time scales (Sprintall et al. 2009; Q. Liu et al. 2015; Mayer et al. 2014, 2018). The surface wind changes related to ENSO also drive the changes in upper ocean circulation and then local ocean heat transport, including in the Atlantic and Indian Oceans (Klein et al. 1999; Yeh et al. 2018; Latif and Grötzner 2000).

The rich regional ocean heat changes impede the development of a complete picture on ENSO-related ocean heat discharge and uptake. There has been some confusion over just how net ocean heat changes are related to ENSO: does the ocean as a whole gain or

lose heat during ENSO events and what mechanisms are involved? Some observational (Roemmich and Gilson 2011; Loeb et al. 2012; Mayer et al. 2014), reanalysis-based (Trenberth et al. 2014a; Wu et al. 2019), and model-based (Fasullo and Nerem 2016; Mayer et al. 2016; Saint-Lu et al. 2016) studies have shown that the ocean loses heat when the surface layer is anomalously warm (El Niño) and gains heat when the surface layer is anomalously cool (La Niña), so that the OHC has a tendency (OHCT) negatively correlated with Niño-3.4 sea surface temperature (SST) anomalies. In contrast, a few other studies have reached the opposite conclusion: they suggest that ocean gains heat during El Niño (Wong et al. 2006; Piecuch and Quinn 2016; Johnson and Birnbaum 2017).

There is also the issue of how much OHC change is simply a redistribution within the ocean, especially from equatorial to off-equatorial regions, as suggested by simple theories, such as the delayed oscillator and recharge–discharge oscillator (Wyrtki 1975, 1985; Jin 1997), versus a diabatic heat loss into the atmosphere largely through evaporative cooling of the ocean but a moistening and ultimately heating the atmosphere via precipitation P . The latter goes along with the observed increases in global mean surface temperature (GMST) (Trenberth et al. 2002) during and at the end of El Niño events, which in turn allows the heat to be dispersed and radiated to space. This suggests that ENSO plays a key role as a regulator of heat build-up in the Pacific and that there is a true discharge and recharge of the OHC during ENSO. Although there appeared to be little or no change in the net radiation at the top of the atmosphere (TOA) through the ERBE period, a period that includes the 1986–87 El Niño and the 1988–89 La Niña (Sun and Trenberth 1998, Sun 2000), this is likely the exception, rather than generally true.

The lack of consensus on ENSO's role in Earth's energy budget suggests that there is substantial uncertainty in the net OHC change due to ENSO. The uncertainty apparently stems mainly from ocean observations and their analysis into gridded products, as well as the domain considered, the interannual time scales (Abraham et al. 2013; Boyer et al. 2016), and the large variability from one El Niño event to another (Trenberth et al. 2014a; Timmermann et al. 2018). In particular, consideration of only part of the full domain by some studies clearly omits some of the energy flows that are involved.

This study examines the regional and global ocean heat evolution related to ENSO using ocean observations, reanalysis, and model datasets (e.g., how the heat is redistributed within the ocean and how much net heat changes in the ocean). The data and methods (section 2) and the global pattern of ENSO variability for SST,

OHC, OHCT, surface fluxes, and P (section 3) provide the basis for a holistic understanding of the ENSO-related OHC changes. The latter especially requires improved knowledge of OHC variations in the core tropical Pacific Ocean. It also requires knowledge of remote OHC changes outside the tropics associated with ENSO teleconnections, and how these local and remote changes contribute to the global imbalance. Therefore, section 4 focuses on the OHC changes in the tropics, and in section 5 OHC changes outside the tropics are examined and discussed. Net OHC changes for the global ocean are also considered and compared with other independent observations in section 5. A summary is provided in section 6.

2. Data and methods

a. Observations and analyses

Beginning in 1979, an ocean mooring was established at 110°W on the equator with a thermistor string to monitor the upper 400 m or so of the ocean, and this was expanded into the Tropical Atmosphere Ocean (TAO)–TRITON array of moored buoys, mainly after 1992. Ship observations—for example, expendable bathythermograph (XBT) and conductivity–temperature–depth (CTD)—are mainly in midlatitudes and along frequently traveled maritime routes since the 1970s. After 2000, these vital observations were complemented by the growing number of Argo ocean profiler observations (Riser et al. 2016; von Schuckmann et al. 2016). The knowledge gained has enabled an improved estimate of historical ocean subsurface temperature fields back in time for several decades (Cheng et al. 2017a), providing an opportunity to revisit the OHCT–ENSO relationships during several El Niño events.

High-quality observations of the radiative imbalance at the TOA, especially after March 2000 (Loeb et al. 2009, 2018), together with atmospheric reanalysis, can be used to derive estimates of Earth's net surface energy flux (e.g., Trenberth et al. 2001; Trenberth and Fasullo 2008, 2017), which recently have been extended backward to 1985 (Allan et al. 2014; C. Liu et al. 2015; Liu et al. 2017). These results provide a valuable complement to in situ measurements to enable us to better understand the ocean heat budget on global and local scales. The energy and water budget analyses can be assessed by how well closure is achieved and are ideally suited to analysis of large-scale variability, because errors from the method decrease with spatial averaging, unlike those for in situ parameterized fluxes.

Therefore, this study mainly analyzes the ENSO-related OHC changes using the improved records of ocean estimates (Cheng and Zhu 2016; Cheng et al.

2017a) and surface flux data after 1985. In addition, state-of-the-art ocean reanalyses (Balmaseda et al. 2013; Zuo et al. 2018) and the Community Earth System Model Large Ensemble project (Kay et al. 2015) are also used in this study. Previous research has indicated that the atmospheric and oceanic reanalyses (Mayer et al. 2014) and some models (Mayer et al. 2016; Fasullo and Nerem 2016) are capable of simulating ENSO-related OHC changes. Ocean observation products contain substantial errors from month to month (Trenberth et al. 2016) because of the sparseness of observations and inadequate analysis methods. For instance, ocean reanalyses are sensitive to model bias and observation system changes (i.e., Palmer et al. 2017), surface fluxes are affected by errors in atmospheric reanalysis (i.e., Trenberth and Fasullo 2018), and model simulations are impacted by model bias. But all of the datasets have their unique strengths and by combining the four types of products and using physical conservation constraints, a comprehensive understanding of ENSO variability can be achieved.

b. Ocean subsurface products

Ocean observational temperature products from the Institute of Atmospheric Physics (IAP) are used in this study as a monthly dataset available for the upper 2000 m (Cheng and Zhu 2016; Cheng et al. 2017a). The IAP product is an ocean objective analysis dataset, with $1^\circ \times 1^\circ$ spatial resolution spanning from 1940 to now. The IAP product has advantages in both instrumental error reduction (required to ensure high-quality in situ observations) and a gap-filling method (to provide a homogenous product with complete global ocean coverage). This product merges all of the available ocean subsurface temperature observations from a variety of instruments, and has also been carefully evaluated by using high-quality Argo-period data since 2005 [see Cheng et al. (2017a) for more information]. The OHC for the 0–100-, 100–300-, 300–2000-, and 0–2000-m depth layers are examined, and detailed calculations can be found in Cheng et al. (2017a). In this study, OHC refers to the anomaly of OHC after removing a monthly climatology within 1981–2010.

We also analyze the first time derivative of OHC (OHCT); however, spurious and large month-to-month noise exists in all observational estimates of OHC (Trenberth et al. 2016), with changes that violate energy conservation. Trenberth et al. (2016) showed that month-to-month variation of OHCT ranges from 6.4 to 11.8 W m^{-2} for five ocean observation products, much larger than the standard deviation of net TOA radiative fluctuations estimated from CERES: 0.64 W m^{-2} . Although IAP data in Trenberth et al. (2016) had the

smallest month-to-month variation of 6.4 W m^{-2} , it remains too large by a factor of 10. This is because noise is amplified when calculating the first derivative, which is a high-pass filter that effectively weights the power spectrum by the frequency. Therefore, a smoother or filter is required to reduce noise before computing tendencies. However, any smoother/filter may also change the variance of the signal of interest and modify its temporal variability. To reduce the noise of observations and correctly capture the OHCT variation on interannual scales (i.e., periods from 1.5 to 7 yr), OHCT is calculated as follows: 1) A high-pass filter with cutoff frequency of $1/102$ (period of 8.5 yr) is applied to the monthly OHC time series to remove the decadal and multidecadal variability. A Lanczos filter is applied (Emery and Thomson 2001, 533–539). 2) Three-month running mean OHC time series are calculated to be consistent with the oceanic Niño index (introduced below in section 2c); 3) a centered difference is used to calculate OHCT based on monthly OHC time series; and 4) a low-pass filter with cut-off frequency of $1/18$ is applied to the OHCT time series to remove the variability with periods of less than 18 months. In the online supplemental material we construct idealized OHC time series to test the accuracy of this tendency calculation method, and the current method accurately captures the temporal variability and the amplitude of OHCT variation with periods of 1.5–7 years. Therefore, any regression and lead/lag analysis based on OHCT is deemed to be reliable. However, as high-frequency noise dampens the magnitude of correlation, an additional 13-month running mean is applied before any correlation is calculated.

Ocean reanalysis data are also analyzed, including Ocean Reanalysis System 5 (ORAS5; Zuo et al. 2018). This ocean reanalysis integrates different sources of information: in situ temperature and salinity profiles, altimeter-derived sea level, sea surface temperature analysis, and ocean-model forced by atmospheric surface fluxes. ORAS5 spans the period 1979 to the present, and uses a $1/4^\circ$ ($\sim 25 \text{ km}$ at the equator) ocean model with 75 vertical levels, including the sea ice model LIM2. Surface fluxes are taken from ERA-Interim (1979–2014) and operational ECMWF data thereafter.

We focus on the 1985–2016 period, consistent with available surface flux data (discussed below), and the build-up of TAO since early 1980s. The Argo network greatly extends the ocean data coverage after 2005 and increases the confidence of OHC from both ocean observation product and reanalysis (Roemmich et al. 2015; Boyer et al. 2016; Riser et al. 2016; Cheng et al. 2017a,b). Therefore, where possible, we provide results for both ocean observations and reanalysis for three periods: 1985–2016, 1985–2004, and 2005–16. Differences in the

results demonstrate the influence of both data quality and ENSO interevent variability. Another consideration is the major volcanic eruption of Mount Pinatubo in June 1991, after which significant ocean heat decreases are found (Fasullo and Nerem 2016). Because we are seeking relationships with ENSO, we exclude the ENSO events during 1991–93 in our (regression) analysis for all observational datasets. When doing correlations, the volcano eruption periods are not removed, but slightly offsetting the start point to avoid these periods leads to only minor differences.

c. Ocean surface products

This study uses updated SST data from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST) (Rayner et al. 2003). An ENSO index is used: the oceanic Niño index (ONI), which is the 3-month running-mean detrended (a linear trend) temperature anomalies in the Niño-3.4 region (5°N–5°S, 120°–170°W). GMST is based on data from GISTEMP Team (2018). Global Precipitation Climatology Project (GPCP) data are used for P , which is available on monthly means from 1979 to 2018 on a 2.5° global grid (Adler et al. 2003).

The two sea surface heat flux products used were derived from TOA net radiative imbalance observations and the atmospheric energy convergences from atmospheric reanalysis (i.e., computed using a “residual method”). The two products include that of Trenberth and Fasullo (2018) (TF hereinafter) and DeepC from Reading University (Liu et al. 2017). The surface heat flux estimates based on the residual method are shown to be superior to reanalysis forecast estimates for large-scale analyses because the latter are model-derived with specified SSTs and have incomplete forcings (especially aerosol effects) (Trenberth et al. 2001; Trenberth and Fasullo 2008, 2018; Trenberth and Fasullo 2013; C. Liu et al. 2015; Liu et al. 2017). Notably, the convergences from which the surface fluxes are computed are based on the atmospheric state analyses and not the forecast fields used to force ocean reanalyses. While this approach avoids the large error in reanalysis surface fluxes, it is known that the atmospheric energy convergences have spurious variations in ocean-to-land energy transports (Trenberth and Fasullo 2013) that could potentially impact the accuracy of surface flux estimates. The TF product uses ERA-Interim, which it estimates results in smaller error than other reanalyses. Liu et al. (2017) adjusted their results by constraining mean heat flux over land (0.08 W m⁻² over 1985–2012). Another error in the computation of atmospheric convergence comes from neglecting moisture enthalpy fluxes (Mayer et al. 2017), which could bias the regional data especially in the tropics. Both the TF and DeepC computations take

this error into account (Trenberth and Fasullo 2018). Negative values correspond to ocean cooling.

d. LENS

The Community Earth System Model (CESM) Large Ensemble project (LENS; Kay et al. 2015) explores the influence of internal variability on detecting climate change in a 40-member ensemble simulation of climate. The “control” is a series of runs with pre-industrial atmospheric greenhouse gas concentrations and no volcanism. Also included in LENS is an extended control run of nearly 2000 years. A key strength of the ensemble is the fairly realistic spatial representation of ENSO in the model (Fasullo et al. 2016; Fasullo and Nerem 2016), but the model Niño-3.4 SST variance amplitude is about 40% too large (Trenberth et al. 2015). Also ENSO asymmetry is not well simulated in the LENS and many climate models (Zhang and Sun 2014; Sun et al. 2016). These enable budgets to be explored, however. Here we use the extended control run to quantify ENSO’s impact on energy flows and thereby exclude the influence of external forcing, and especially volcanic eruptions, on our analysis.

e. Data processing and statistical significance

The 90% confidence interval is uniformly applied here in determining statistical significance. Linear regression coefficients are calculated by the ordinary least squares method. For correlations between variables related to ENSO, autocorrelation is taken into account according to Trenberth (1984) based on Davis (1976). The autocorrelation is definitely variable with different time series, but generally, for the 384 months from 1985 to 2016, the effective degrees of freedom are typically about 50–150, suggesting statistical significance at the 10% level if the correlations exceed 0.2–0.3.

f. Ocean energy budget

We use $d\text{OHC}/dt$ (OHCT) as a key metric for examining the ENSO-related OHC change based on the local ocean heat budget equation:

$$\text{OHCT} = Q - \nabla \cdot F_O, \quad (1)$$

where OHCT can be attributed to an air–sea plus ice–sea net heat exchange component Q that is diabatic and an oceanic heat convergence $-\nabla \cdot F_O$ that is related to ocean heat redistribution and advection. The simplified notation here belies the fact that a vertical integral has to be performed. Because the three-dimensional ocean velocity fields are less well constrained by observations than is temperature, we compute it as a residual while making extensive use of OHC. Because we use the top

FIG. 1. Time series of ONI, GMST, OHC within 0–100 and 0–2000 m (IAP is solid line; ORAS5 is dashed), and P (GPCP) in the equatorial Pacific Ocean within 5°N – 5°S , 120°E – 80°W after 1980. The red and blue shading respectively mark the El Niño and La Niña events with $|\text{ONI}| > 0.5$ K. For all of these time series, a high-pass filter with cutoff frequency of $1/102$ (period of 8.5 yr) is applied before further analyses. To gain better visibility, time series except OHCT are smoothed by a 13-month running smoother weighted by $(1, 6, 19, 42, 71, 96, 106, 96, 71, 42, 19, 6, 1)/576$ (Trenberth et al. 2007).

2000 m of the ocean, the vertical heat transport through the bottom of the layer on interannual scales is very small and is deemed to be negligible on the considered time scales; for instance, the magnitude of interannual temperature fluctuations within 1500–2000 m is smaller by at least a factor of 10 than near the sea surface (based on IAP and ORAS5 data).

3. Global pattern of ENSO variability links to OHC

Broad relationships are first explored using time series (Fig. 1). GMST closely follows ONI change with a delay of 2–3 months (Trenberth et al. 2002). Thus, based on GMST, a mini-global warming appears soon after the peak of each El Niño (e.g., 2015/16) whereas a mini-global cooling occurs during La Niña (e.g., 2017). However, 2017 still was the third-warmest year on record despite the unabated long-term global surface warming (GMST data).

OHC in the 0–100- and 0–2000-m layers in the equatorial Pacific (averaged over 5°N – 5°S , 120°E – 80°W) (Fig. 1) has significant correlation with ENSO. Changes of OHC at 0–100 m are very similar to SST in the Niño-3.4 region (ONI), because they mainly reveal the changes within surface mixed layer above the main thermocline. However, OHC at 0–2000 m leads ONI by ~ 5 months on average from 1985 to 2016. This equatorial OHC was thought to be a major source of predictability for ENSO (Wyrki 1985; Cane and Zebeik 1985; Meinen and McPhaden 2000; Kug et al. 2010; McPhaden 2012) before the Argo network was established and there were insufficient observations to monitor subsurface OHC changes (McPhaden et al. 2001). Instead, the warm water volume (WWV) above the 20°C isotherm within 5°N – 5°S and 120°E – 80°W was widely used as an index of OHC after the build-up of the TAO array (Meinen and McPhaden 2000). Also used is sea level change since 1993 (Kug et al. 2010; Jin 1997).

McPhaden (2012) suggested that WWV leads ONI by 6 (2–3) months with maximum correlation of 0.68 (0.75) from 1980 to 1999 (from 2000 to 2010). We find that equatorial OHC at 0–2000 m leads ONI by 6–7 months from 1980 to 1999 (peak correlation 0.84); this lead reduces to 3 months after 2000 (peak correlation 0.85). It remains to be seen whether this increased correlation is due to improved estimation of subsurface temperature in IAP/ORAS5 with regard to the TAO-derived WWV used by McPhaden (2012) or instead occurs because the 0–2000-m OHC is a better ENSO predictor. In any case this should be explored in the future, taking account of the statistical significance of the difference. The reduction of the lead time between OHC and ONI after 2000 is due to the shift from more east Pacific El Niño events in the 1980s and 1990s to more central Pacific El Niño events more recently (Ashok et al. 2007; Kao and Yu 2009).

Estimated OHCT (Fig. 1) shows that the equatorial Pacific begins losing heat during the developing stage of El Niño, and reaches a minimum shortly after the El Niño peak with a magnitude from -10 to -80 W m^{-2} depending on its strength. Note that P is greatly increased over the equatorial Pacific during El Niño and shows strong negative correlations with OHCT (Fig. 1). The maximum regional P anomaly during 1997–98 El Niño is $>4 \text{ mm day}^{-1}$, corresponding to a latent heat release of $>120 \text{ W m}^{-2}$ (but note that P change is due to both atmospheric lateral moisture convergence and local evaporation). Indeed, the increased P and net OHC loss during El Niño imply an evaporative energy exchange from ocean to atmosphere. Note that P has particular relevance to the changes of atmospheric circulation because it signals latent heating in the upper troposphere and thus atmospheric divergence and hence teleconnections (Trenberth et al. 1998, 2002). Strong P in the tropical Pacific during El Niño indicates a significant perturbation to the atmospheric circulation and a global impact.

The global multivariate pattern of ENSO variability is estimated for SST, OHC, P , OHCT, and surface flux (Figs. 2–4). In each case, a linear regression is calculated at each grid point of ONI onto the detrended anomalies of each of the other variables. The spatial pattern of SST related to El Niño is familiar (e.g., Klein et al. 1999; Trenberth et al. 2002; Alexander et al. 2002) and characterized by anomalous warming in the central and east Pacific, extending to the west coast of North and South America, and cooling in the far west Pacific extending to the middle latitudes in the north and south central Pacific (30° – 40° N, 30° – 40° S) (Fig. 2). The North Pacific and Atlantic changes are consistent with Pacific–North America (PNA) teleconnections (Alexander et al.

2002), depending on the seasonal cycle and air–sea interactions both within and beyond the North Pacific Ocean (Alexander et al. 2002; Trenberth et al. 2014b). In the tropical Atlantic Ocean, the regression features a transition from a cooling to a warming state with El Niño development (30° S– 30° N) (Fig. 2), as found with earlier data (Trenberth et al. 2002). A cold–warm–cold pattern stretches from the southwest South Pacific to the South Atlantic during El Niño, sometimes referred to as the Antarctic dipole (ADP) mode (Yuan 2004). In the Indian Ocean the signature of the Indian Ocean dipole appears contemporaneously with the ENSO signal.

OHC anomalies for the upper 2000 m during ENSO (right column of Fig. 2; IAP data) show a very different pattern from SST because OHC integrates subsurface signals (i.e., changes of the main thermocline). Similar results are found for ORAS5 data (see Fig. S1 in the online supplemental material). A zonal gradient in anomalous OHC evolves with El Niño with a divide at around 170° E on the equator at zero lag, farther east than SST (divide at $\sim 150^{\circ}$ E), as also identified in Roemmich and Gilson (2011) using ~ 7 years of Argo data (but with no lead/lag analysis). There is a strong and broad decrease of OHC in the west Pacific that extends into the South and North Pacific to $\sim 20^{\circ}$ S and $\sim 20^{\circ}$ N. This contrasting pattern in the Pacific Ocean persists from a lag from -8 to $+8$ months and peaks at lag 0. The reverse zonal gradient (warm in the west and cold in the east) is evident in the Indian Ocean during El Niño. The OHC change in the tropical Atlantic Ocean (30° S– 30° N) is similar to SST changes, showing a shift from a colder state (lag = -8 , -4) to a warmer state (lag = 4 , 8), similar to Mayer et al. (2014) using ORAS4 data.

Global patterns of OHCT (Fig. 3 for IAP) are determined by both surface fluxes Q , which are related to the net heat exchange between ocean and atmosphere, and ocean heat convergence [Eq. (1)]. ORAS5 shows consistent OHCT patterns (Fig. S1). Positive OHCT is evident in the east Pacific and coastal areas along the west of North and South America before El Niño (with a peak value $>70 \text{ W m}^{-2}$ per K in the tropics; Fig. 3, left column) and is responsible for the positive OHC anomalies before El Niño (Fig. 2). Negative OHCT exists in the west Pacific before El Niño, responsible for the negative OHC anomalies during El Niño. A similar—but reverse—zonal dipole OHCT pattern is also found in the Indian Ocean. The dipole structures in both the Indian and Pacific Oceans change sign after El Niño events. In the Pacific, the increase in tendency appears first near the equator (Fig. 3; OHCT at lag = -8), followed by changes in the western part of the basin (Fig. 3; OHCT at lag = -4 , 0). This is why OHC

FIG. 2. Regression (left) between SST and ONI (HadISST data; K K^{-1}) and (right) between OHC and ONI (IAP data; $10^8 \text{ J m}^{-2} \text{ K}^{-1}$) in each $1^\circ \times 1^\circ$ grid after 1985. Different lags are shown: -8, -4, 0, 4, and 8 months (positive means after El Niño). The stipples indicate signals that are insignificant (at 90% confidence level).

signals (Fig. 2) are higher at lag = 0 than lag = -4 and 4 in the tropical Pacific and Indian Oceans. In the next section, we examine the zonal/meridional mean changes to investigate the observed OHCT variability.

Net surface flux regressions show patterns distinct from those of OHCT in both the Pacific and Indian Oceans (Fig. 3, right column). An ocean cooling in the central and east Pacific before El Niño becomes

FIG. 3. As in Fig. 2, but (left) between OHCT and ONI (IAP data; $\text{W m}^{-2} \text{K}^{-1}$) and (right) between surface flux and ONI (DeepC data; $\text{W m}^{-2} \text{K}^{-1}$). This figure is intended to illustrate the geographical pattern of ENSO impacts on global ocean heat content change at different ENSO evolution periods.

FIG. 4. As in Fig. 2, but between P (GPCP; $\text{mm day}^{-1} \text{K}^{-1}$) and ONI.

meridionally broader with time, and it extends into the northeast and southeast Pacific (30° – 60°N , 30° – 50°S) during El Niño. In addition, a weak but broad-scale warming influence, driven by surface fluxes, is found in the Indian and Atlantic Oceans (less than $5 \text{ W m}^{-2} \text{K}^{-1}$), except for some areas in the middle equatorial Indian Ocean and northwest and southwest Atlantic Ocean

(Fig. 3). The North Atlantic pattern is likely related to the PNA teleconnection as well as effects from the Walker-type circulations in the tropics that result in the Atlantic Ocean cooling around 40°N and warming in the lower and higher latitudes (20° and 50°N) (Fig. 3) (Alexander et al. 2002).

The net surface energy flux consists of shortwave and longwave radiation and sensible and latent heat turbulent fluxes, plus other small terms like P enthalpy (Trenberth 2011). The latent heat flux is associated with the water cycle because it is manifested by evaporation. The evaporation is then closely linked to the P in the tropics (Trenberth 2011). El Niño is associated with a strong increase of P (Fig. 4) in the central and east Pacific. The P maximum occurs north of the equator, in contrast with the net surface heat in Fig. 3, which is centered at the equator. Note that P and surface flux are not necessarily collocated because P is also associated with the atmospheric circulation changes. The maximum increase of P is $>2 \text{ mm day}^{-1}$ per 1-K increase of ONI (equivalent to $>58 \text{ W m}^{-2}$ in latent energy release). Reduced P in the tropical west Pacific extends into the middle and high latitudes of the northwest and southwest Pacific, also consistent with surface flux (Fig. 3). There is reduced P in the tropical Atlantic (Fig. 4) but with an increase in the midlatitudes (magnitude less than 0.3 mm day^{-1} per 1-K ONI change). The change in the global P pattern indicates a strong influence on the atmospheric circulation by ENSO (Oort and Yienger 1996; Alexander et al. 2002; McPhaden et al. 2006).

4. Ocean heat evolution in the tropics

This section focuses on OHC changes in the tropics where ENSO events originate and we present results from different perspectives highlighting various spatial features including meridional gradients (section 4a), zonal evolutions (section 4b), changes as a function of depth (section 4c), net changes in the tropics (section 4d), and a synthesis and discussion (section 4e).

a. Meridional ocean heat evolution

A Hovmöller diagram of zonal mean OHCT in the Pacific Ocean for the upper 2000 m from 1985 to 2016 (Fig. 5a) and lagged cross correlations of ONI with zonal mean OHCT as a function of latitude (Fig. 6a) show a clear ENSO-related heat redistribution inside and outside the equatorial zone. The southern and equatorial tropical ocean (20°S – 5°N) accumulates heat during La Niña and before El Niño ($\text{OHCT} > 0$) and loses heat during/after El Niño ($\text{OHCT} < 0$) (Figs. 5a and 6a). Opposing OHCT values are evident around the 5°N latitude line (5° – 30°N vs 20°S – 5°N), indicating that heat

FIG. 5. Hovmöller diagrams of the zonal mean (top) OHCT (W m^{-2}) and (bottom) OHC (10^{20} J) at each 1° latitude band from 25°S to 25°N in the tropical Pacific basin from IAP data, compared with ONI (in green). The peaks of El Niño events since 1985 are marked with vertical dashed lines.

loss in the southern tropics during El Niño is discharged into the off-equatorial regions in the Northern Hemisphere. The striking divide near 5°N is collocated with the intertropical convergence zone (ITCZ) in the Pacific. The magnitude of this heat redistribution can reach $10\text{--}20 \text{ W m}^{-2}$ in the Pacific for each event, with a maximum of $>40 \text{ W m}^{-2}$ during the 1997–98 super El Niño event (Fig. 5). The heat discharge is relatively small during the 2015–16 super El Niño ($20\text{--}25 \text{ W m}^{-2}$), which is also seen in Mayer et al. (2018).

The corresponding zonal mean OHC anomalies (Fig. 5b) show that the ocean within $20^\circ\text{S}\text{--}5^\circ\text{N}$ is in a state of warming before El Niño and also peaks before El Niño ($\text{OHC} > 0$), and then it stays in an anomalous cold state after El Niño ($\text{OHC} < 0$). Cross-correlations of ONI with OHC (Fig. 6b) show that during 1985–2016 equatorial OHC ($5^\circ\text{S}\text{--}5^\circ\text{N}$) leads ONI by 3–7 months on average (Fig. 6b) over both EP and CP El Niños. But North Pacific ($5^\circ\text{--}15^\circ\text{N}$) OHC leads ONI by 1–6 months, slightly behind the equatorial change (Fig. 6b), implying a tropical driver of OHC variability. This result provides a more complete picture of OHC change in the tropics: the ocean builds up heat prior to El Niño within $20^\circ\text{S}\text{--}5^\circ\text{N}$ as a precursor, and

loses heat during and after El Niño, some of which is discharged into the North Pacific, which is also shown in Meinen and McPhaden (2000).

The cross-correlation between net surface flux Q with ONI (Fig. 6d) shows a pattern distinct from OHCT (Fig. 6a). Surface flux anomalies result in ocean cooling in the tropical Indo-Pacific basin (within $10^\circ\text{S}\text{--}10^\circ\text{N}$) during El Niño, coincident with the sea surface warming (Fig. 6c) and P increase (Fig. 6e) for the same zonal bands. Hence high SST during El Niño drives the surface flux release via evaporation, and the associated increase in atmospheric water vapor boosts convection and P in the tropics. The propagation of SST and Q anomalies from the equator (i.e., lag = 0 near the equator) to higher latitudes, (i.e., lag = 10 at 20°N) during El Niño is also evident (which is more obvious for the Northern Hemisphere than the Southern Hemisphere), suggestive of the remote impact of ENSO into the subtropics via teleconnections.

The contrasting patterns between OHCT and Q indicate the dominance of ocean heat transport for the meridional ocean heat content change, suggesting net meridional heat transport anomalies due to ENSO:

FIG. 6. Lagged cross correlations of zonal mean (a) OHCT (IAP data), (b) OHC (IAP data), (c) SST (HadISST data), (d) surface flux (DeepC data), and (e) P (GPCP data) in the Pacific Ocean as a function of latitude. The correlations are calculated within the 1985–2016 period when all datasets are available. The 90% confidence interval of the correlation (contours) is ~ 0.3 (it depends on the effective degrees of freedom).

there is a net northward heat transport into the north (5° – 20° N) before and during El Niño, and reduced northward transport during La Niña. This is evident when using regression to identify the magnitude of OHCT and Q (not shown), rather than the correlations shown in Fig. 6. This is also evident in LENS simulations (online supplemental Fig. S2); the model nodal line is also centered at 5° N, and the meridional extension is reduced. Observations give direct

evidence for the heat recharge/discharge theory proposed decades ago (e.g., Wyrski 1985; Cane and Zebiak 1985; Zebiak 1989; Jin 1997) and also confirm the significant adiabatic component of ENSO's meridional structure.

b. Zonal ocean heat evolution

The evolution of meridional mean OHCT (0–2000 m) in the tropics (20° S– 20° N) with ENSO (Figs. 7 and 8a)

FIG. 7. Hovmöller diagram of meridional mean OHCT in the tropics (20°S–20°N) from IAP data after 1985. ONI is superimposed at 150°W since it is the averaged SST within 120°–170°W. The peaks of El Niño events since 1985 are marked with horizontal dashed lines. The three major basins are roughly separately by black vertical dashed lines.

from 1985 to 2016 reveals strong zonal gradients in OHCT in the Pacific (cooling in the west and warming in the east) in the lead up to El Niño (lead ONI by 5–10 months), with a predominance of eastward propagating signals in the Pacific basin, and slower western propagating signals in the Indian Ocean, except during the 2015/16. The local OHCT reduction in the west

Pacific is as large as -20 W m^{-2} for every event (Fig. 7), and even reaches from -40 to -50 W m^{-2} during the 1997–98 super El Niño. This zonal seesaw pattern reverses quickly after the peak of El Niño (Figs. 7 and 8a). The warming in the west Pacific rises to $\sim 60 \text{ W m}^{-2}$ after the 1997–98 super El Niño, with cooling of approximately -40 W m^{-2} in the east. This OHCT evolution

FIG. 8. Lagged cross correlations of meridional mean (a) OHCT (IAP data), (b) OHC (IAP data), (c) SST (HadISST data), (d) surface flux (DeepC data), and (e) P (GPCP data) in the tropics (20°S – 20°N) as a function of longitude. The correlations are calculated within the 1985–2016 period when all datasets are available. The 90% confidence interval of the correlation (contours) is ~ 0.3 (based on the effective degrees of freedom). The green line indicates the division longitude of SST change at the equator. The range 20°S – 20°N is chosen because it is a clear boundary for the zonal mean OHCT variability in Figs. 5 and 6. Using 30°S – 30°N or 10°S – 10°N provides similar results (not shown).

results in a warmer ocean (higher OHC) than normal in the eastern tropical Pacific Ocean (east of 160°W) during El Niño; meanwhile, there is negative OHC in the west (see OHC in Fig. 8b). The Pacific OHC lags ONI by 0–1 month. The Indian Ocean also shows a similar seesaw pattern before and after El Niño but with opposite sign, with $\text{OHCT} > 0$ before and during the peak of El Niño and $\text{OHCT} < 0$ afterward in the west (Figs. 7 and 8a), probably related to Indian Ocean dipole events. But the magnitude is smaller than in the tropical Pacific Ocean, with a maximum OHCT of $\sim 20 \text{ W m}^{-2}$ in the west Indian Ocean during the 1997–98 super El Niño. However, the dipole during the 2015–16 super El Niño is

much weaker than in 1997–98 and many other events, as seen in ORASS in Mayer et al. (2018). The Indian OHC lags ONI by about 3 months.

Cross-correlations between Q and ONI (Fig. 8d) again reveal distinct patterns related to OHCT in the Indian and Pacific basins. Strong negative Q appears in the central and east Pacific (east of 160°E) during El Niño. Meanwhile, the west Pacific and east Indian Oceans receive heat within 60° – 160°E ($Q > 0$) (Fig. 8d). Although different from OHCT, Q anomalies in the tropical Indian and Pacific basins are consistent with SST and P anomalies (Figs. 8c–e). The sea surface warms, releasing heat into the atmosphere (negative Q),

and P is enhanced in the central and east Pacific Ocean and west Indian Ocean during El Niño; SST decreases in the west Pacific along with ocean warming by Q and decreased P . In particular, the SST, Q , and P seesaw pattern in the Pacific exhibits a node at $\sim 170^\circ\text{E}$ in the tropical Pacific, different from OHCT and OHC ($\sim 170^\circ\text{W}$). Additionally, there is a west–east propagation of the maximum correlations between lag = -20 and lag = 0 in Figs. 8c–e, also shown in divergence of atmospheric latent heat transport in Mayer et al. (2013). The consistency between SST, Q , and P anomalies originates from the diabatic effects of ENSO: higher SST in the central and west tropical Pacific and west Indian during El Niño, along with changed winds (not shown), drives higher evaporation, which cools the ocean, and this atmospheric moisture boosts convection and enhances P . These anomalies in turn drive atmospheric teleconnections and remote effects (Trenberth et al. 2002).

The contrasting patterns of OHCT and Q anomalies in the Indian and Pacific Oceans (Fig. 8) suggest that Q contributes to, but is not responsible for, the formation of the local ocean heat tendencies in tropical Indian/Pacific basins. Instead, changes of ocean heat convergence (advection) likely play a key role. Our results indicate that there is a net meridional ocean advection in the tropics: from the west (east) Pacific into the east (west) Pacific and from the east (west) Indian Ocean into the west (east) Indian Ocean before (after) El Niño. LENS simulations also confirm that anomalous heat convergence is the main reason for the formation of OHCT pattern (online supplemental Fig. S3).

In contrast with the Pacific and Indian Oceans, Atlantic OHCT shows weaker correlations with ENSO (Fig. 8a). Because of the geometry asymmetry across the equator in the Atlantic, the averages within 20°S – 20°N of zonal and meridional changes are aliased: meridional mean signals in the west (east) Atlantic are associated with changes in the north (south). The west (north) tropical Atlantic features positive OHCT 10–12 months before the El Niño peak (Fig. 8a), and the east (south) tropical Atlantic starts warming several months later than the west. A positive OHC anomaly during and after El Niño (lagging ONI by ~ 5 months) exists in the west Atlantic, and the east Atlantic OHC lags the west Atlantic by 5–10 months, meaning that SST changes in the tropical Atlantic are very coherent with OHC variability (Fig. 8c), consistent with the previous finding that the tropical North Atlantic SST is a maximum ~ 5 months after El Niño (e.g., Enfield and Mayer 1997; Trenberth et al. 2002).

Surface fluxes Q (Fig. 8d) warm the tropical Atlantic Ocean during El Niño (from -5 to 5 months of ONI

peak), along with reduced P over the entire tropical Atlantic bands (Fig. 8e). Comparison between OHCT and Q shows a broad coherence in the west Atlantic and an OHCT lag in the east. Lohmann and Latif (2007) and Mayer et al. (2014) showed that El Niño–related Atlantic SST warming occurs because of surface fluxes rather than ocean dynamics in the tropical and North Atlantic. This explains the consistency between OHCT and Q/P in the west Atlantic. But in the east Atlantic, the wind-driven ocean circulation changes play a more important role (Lohmann and Latif 2007), so that OHCT and Q/P are out of phase.

c. OHC change for different layers

OHC changes in the tropics involve subsurface changes (in particular thermocline variations). For 20°S – 20°N the temperature for the upper 100-m layer (Fig. 9) increases during El Niño (temperature tendency > 0) and quickly decreases after El Niño (temperature tendency < 0). For the regional average, temperature change within 100–300 m is always opposite to the upper 100-m change (Fig. 9) with a divide around 100 m, indicating a strong vertical heat redistribution between 0–100 m and 100–300 m related to ENSO. The major signals are limited to the upper 300 m for most ENSO events (Fig. 9) except for some strong cases in which signals can penetrate down to ~ 800 m (during the 1997–98 super El Niño). Therefore, now we use 0–100 m to indicate upper ocean changes and 100–300 m to indicate the subsurface changes. Using other layers such as 100–500 m or 100–2000 m shows similar results. We note that the global average temperature change at different layers shows a similar evolution as in the tropics (not shown), indicating the dominance of the tropics for the global mean variations at different layers by ENSO (Roemmich and Gilson 2011).

The large degree of consistency in the estimates of OHCT in the upper 100 m from IAP and ORAS5 [Figs. 9 and (online supplemental) S4] decreases in lower levels, and it is stronger during the 1997–98 El Niño, when the decrease OHCT is visible in ORAS5 below 100 m but not in IAP. The major difference of the two datasets during the 1997–98 El Niño comes from the South Pacific (30° – 5°S), where the ocean observations are sparse (not shown). It seems that ORAS5 is more realistic because it shows a net OHCT decrease for 0–300 and 0–2000 m after the El Niño peak, similar to other events (Fig. 9); however, IAP shows a maximum heat loss before the peak. Careful intercomparison is required in future to identify and solve this problem.

This vertical heat redistribution is linked to thermocline variations. A steeper thermocline (e.g., 20°C isotherm) along the equator (5°S – 5°N) is apparent during La Niña in both the Indian and Pacific Oceans

FIG. 9. (top) OHCT in the tropics (20°S – 20°N) within 0–100 (blue), 100–300 (green), 0–300 (purple), and 0–2000 (orange) m, compared with ONI (black). IAP data are shown as solid curves, and ORAS5 is shown as dashed curves. The same 13-point smoother is applied as in Fig. 1. The red and blue shading respectively mark the El Niño and La Niña events with $|\text{ONI}| > 0.5 \text{ K}$. (bottom) The temperature tendency in the tropics (20°S – 20°N) (K month^{-1}) as a function of depth and time from 1980 to 2016 (IAP data). The 100- and 300-m layers are marked with horizontal lines.

(Fig. 10), resulting in a positive temperature anomaly in the west Pacific and east Indian Oceans mainly below 100 m together with a negative temperature anomaly in the east Pacific and west Indian Oceans at the upper 100 m. The thermocline relaxes during El Niño and leads to changes that are opposite to those during La Niña (Fig. 10).

However, it is a mistake to assume that the two layers are doing opposite things locally, as can be seen from the maps (Fig. 11). Spatially, the vertical heat redistribution is dominated by the changes in the tropical Pacific Ocean, with local OHCT anomalies larger than $45 \text{ Wm}^{-2} \text{ K}^{-1}$ ONI change (Fig. 11) in both the 0–100- and 100–300-m layers. Before El Niño, a broad ocean

warming of the upper 100 m in the central and eastern Pacific (east of 170°E) exists, with cooling in the west (west of 170°E)—similar to the SST pattern in Fig. 2. A similar zonal structure is found for 100–300 m, but with much more prominent signals in the west Pacific, and with a more eastward division of $\sim 150^{\circ}\text{W}$ than for 0–100 m (170°E). Therefore, the broader area of ocean warming for the upper 100 m in the east Pacific dominates the tropical/global OHCT for that layer, while the broader cooling area in the west Pacific dominates the tropical/global OHCT for 100–300 m (Figs. 9 and 11). Similar results are found for ORAS5 (Fig. S4). The reverse sign for the layers (Fig. 9) is only really evident from 170°E to 150°W . After El Niño, a similar pattern

FIG. 10. A cross section from 5°N to 5°S along the equator illustrates the temperature structure (°C) with depth for an El Niño in 2015 vs a La Niña in 2010 for November–January. The thermocline is the temperature gradient centered approximately on the 20°C isotherm, highlighted by the dark black line. The white spaces denote the landmasses. IAP data are used.

evolves for both 0–100 and 100–300 m with opposite signs when compared with those before El Niño.

d. Net OHC changes in the tropics

As there is large regional compensation, especially in the tropical zonal mean, among the three major ocean basins, and in the vertical, we explore the net OHC changes in the tropics within 20°S–20°N.

Both ORAS5 and IAP data show that the tropical Pacific Ocean cools strongly during and after El Niño (from −5 to 7 months) with a maximum of -0.29 ± 0.05 PW per 1 K ONI change (IAP 1985–2016), which is consistent among different periods and datasets (Fig. 12; see also online supplemental Figs. S5 and S7). When use is made of OHCT for the upper 300 m, a slightly smaller value results: -0.24 ± 0.05 PW K^{−1} (IAP). This cooling is mainly caused by enhanced surface fluxes Q : -0.26 ± 0.03 PW K^{−1} (DeepC 1985–2015), with TF and DeepC within 2000–15 showing similar results (Fig. 13; see also online supplemental Figs. S6 and S7). The Q is slightly smaller than the OHCT itself although their error bars overlap. This first indicates that the surface flux cooling is mainly responsible for the net tropical Pacific Ocean cooling, despite the strong local heat redistribution shown in previous sections, as found by Mayer et al. (2014). The small difference between OHCT and Q is likely linked to the heat exchanges between the tropics and off-tropics and also the uncertainty in observations. A comparison of OHCT/ Q during 1985–2004 with the 2005–16 period (Figs. S5 and S6) reveals a strong cooling in the El Niño period: OHCT/ Q leads ONI by

10–13 months for 2005–16 but by 17–21 months for 1985–2004, as previously noted by McPhaden (2012). The OHCT changes before El Niño (lag from −20 to −10) are much larger than surface flux anomalies (for all periods), indicating a dominant role of ocean dynamics. LENS shows similar Q changes in tropical Pacific but weaker OHCT change than observations (Figs. 12 and 13), indicating an error in ocean heat transport in model.

In the tropical Indian Ocean, both datasets show that surface flux anomalies have a warming effect during El Niño (e.g., 0.05 ± 0.02 PW K^{−1} for DeepC at lag = −1 for 1985–2015) and cooling after El Niño (e.g., -0.02 ± 0.01 PW K^{−1} for DeepC at lag = 10) (Fig. 13). However, the OHCT change is larger than suggested by Q anomalies alone: OHCT leads ONI by 5 months (2 months) for IAP (ORAS5) with a peak of 0.12 ± 0.02 (0.14 ± 0.04) PW K^{−1} for 1985–2016 (Fig. 12). After El Niño, OHCT decreases to -0.12 ± 0.02 (-0.15 ± 0.04) PW K^{−1} for IAP (ORAS5) at lag = 12 (10) months for 1985–2016. The difference between Q and OHCT suggests a major contribution of ocean heat exchange with either the southern Indian Ocean or tropical Pacific Ocean via ITF. Before El Niño, the stronger ITF transports warmer water into the tropical Indian Ocean, whereas after El Niño the ITF is reduced and cools the Indian Ocean by 0.1 – 0.5 PW K^{−1} (Chirokova and Webster 2006; Sprintall et al. 2009; Sprintall et al. 2014; Mayer et al. 2014; Lee et al. 2015; Q. Liu et al. 2015), and the ITF lags ONI by ~ 7 months (Q. Liu et al. 2015). However, there was a much stronger reduction of ITF

FIG. 11. As in Fig. 2, but between OHCT and ONI using IAP data for the (left) 0–100- and (right) 100–300-m layers.

during 2015/16 versus the 1997–98 El Niño event that probably contributed to the surprisingly weak Pacific OHC loss in 2015–16 (Mayer et al. 2018).

The tropical Atlantic Ocean also has a significant heat accumulation during El Niño (Fig. 12), with coherent Q evolution with ENSO (Fig. 13). The magnitude of

OHCT change is $\sim 0.09 \pm 0.02$ ($\sim 0.13 \pm 0.03$) PW K^{-1} for IAP (ORAS5) for 1985–2016 at lag = 0 (Fig. 12), and with a similar amount of Q change: $\sim 0.07 \pm 0.02$ ($\sim 0.08 \pm 0.02$) PW K^{-1} for DeepC (TF) for 1985–2015 (2000–16). Hence the surface flux plays a dominant role in the tropical Atlantic energy budget, as discussed in both the

FIG. 12. Regional OHCT changes related to ENSO. The tropics are within 20°S–20°N. Total OHCT in each box (PW) is regressed with ONI at lags within the interval from –20 to 20 months. Both IAP analysis and ORAS5 reanalysis are used and are shown for the 1985–2016 period. Similar results for LENS model simulations are shown as dashed orange lines (100-yr preindustrial runs). The shadings (or bars for LENS) show the 90% confidence interval.

previous section and previous studies (Latif and Grötzner 2000; Alexander and Scott 2002; Mayer et al. 2014). Anomalies in Q and OHCT changes for 1985–99 and 2000–2016 (Figs. S5 and S6) are mutually consistent within the range of uncertainty.

While warming in the tropical Atlantic and Indian Oceans during El Niño offsets somewhat the strong cooling trend in the tropical Pacific, for the global tropics (Fig. 12) there is a net tropical cooling of $-0.16 \pm 0.04 \text{ PW K}^{-1}$ at lag = 2 (IAP) to $-0.25 \pm 0.06 \text{ PW K}^{-1}$ at lag = 7 (ORAS5) 1985–2016. Wu et al. (2019) showed a consistent net tropical cooling of about -0.14 PW K^{-1} at lag = 8 based on another ocean reanalysis. Using 0–300 m instead of 0–2000 m results in similar but slightly less cooling: $-0.12 \pm 0.05 \text{ PW K}^{-1}$ (IAP). Surface evaporative cooling (Q) causes at least one-half of the OHCT change: $0.11 \pm 0.04 \text{ PW K}^{-1}$ (TF2000–2016) at lag = 1, $0.11 \pm 0.04 \text{ PW K}^{-1}$ (DeepC 1985–2015), and 0.10 PW K^{-1} in Wu et al. (2019). Before El Niño, increases in both the Indian and Pacific basins create a positive OHCT peak of 0.18 ± 0.04 (0.20 ± 0.06) PW K^{-1} for IAP lag = –16 (ORAS5 lag = –13) for 1985–2016, and the Q contribution is less than 0.1 PW K^{-1} . Therefore, for the whole tropics, OHCT transitions from positive before an El Niño to negative during and after an El Niño, and this is confirmed by

independent Q observations in Fig. 13, although the change of Q is somewhat smaller in magnitude before El Niño when compared with OHCT (Fig. 12). The difference between Q and OHCT in the tropics mainly stems from the Indian Ocean, probably because of strong heat exchanges between the Indian Ocean and the Southern and Pacific Oceans.

e. Discussion

The analyses of OHCT and related variables and their evolution in the tropics with ENSO demonstrate that both 1) the diabatic effect on net global OHC and 2) the adiabatic effect associated with horizontal and vertical ocean heat redistribution are important. Naturally the latter is more important locally but effects cancel as the domain is extended. The many processes involved are summarized in Fig. 14.

During the developing stage of an El Niño event, the westerly wind bursts trigger Kelvin waves and thermocline feedback, causing the shoaling of the main thermocline in the western Pacific and uplifting and expanding the western Pacific warm pool eastward (Wang 2004; Roemmich and Gilson 2011; Yin et al. 2018) (dashed green line in Fig. 14). This reduces heat storage in the western Pacific, especially below 100 m (Figs. 9, 10, 11, and 14). The trigger of an El Niño event is

FIG. 13. Regional surface flux changes related to ENSO. Total surface flux in each box (PW) is regressed with ONI at lags within the interval from -20 to 20 months. DeepC (green; 1985–2015) and TF (red; 2000–16) data are used. LENS results are shown as dashed orange lines. The shadings (or bars for LENS) show the 90% confidence interval.

still debated. Another nonlinear and diabatic view, proposed by Sun (1997), attributes the ENSO fluctuations to the existence of two unstable equilibria. At least two complementary mechanisms are involved. The first is strong eastward-flowing anomalous currents that carry warm surface waters from the western Pacific into the central Pacific (Johnson et al. 2000), resulting in net eastward advection of heat. The second is the eastward-propagating, downwelling equatorial Kelvin waves, which leave a depressed thermocline in their wake over thousands of kilometers in the east–west direction (Weisberg and Wang 1997; Boulanger and Menkes 1999; McPhaden and Yu 1999; McPhaden 2015). The depressed thermocline, along with altered local winds, reduces the normal upwelling of cool water to the surface, resulting in an increase of OHC in the central and eastern Pacific (Figs. 2 and 14), especially for the upper 100 m ($\text{OHC} > 0$; Figs. 9, 10, 11, and 14). Changes in the temperature of the upwelled waters in turn control the SST in the eastern equatorial Pacific (Fig. 2), reduce SST gradients along the equator, and create anomalous winds and winds that amplify the SST gradient and also affect upwelling. This is the Bjerknes feedback that maintains and strengthens El Niño events through increases of OHC in the east Pacific (mainly within 0–100 m) and decreases of OHC in the west Pacific during El Niño (mainly within 100–300 m), illustrated in Fig. 14.

Besides the zonal and vertical Pacific heat redistribution, there is also heat recharge/discharge into and away from the equatorial regions (Jin 1997). Equatorial OHC peaks several months before the peak of El Niño (Figs. 1 and 6) and then decreases ($\text{OHCT} < 0$ within from -6 to 10 months of the El Niño peak). Therefore, there is a net heat transport from the equatorial Pacific (20°S – 5°N) into the off-equatorial regions (5° – 20°N) during El Niño (Figs. 5 and 6), as illustrated in Fig. 14 (solid green arrow). This character is also seen in the second mode of the 20°C isotherm changes (Meinen and McPhaden 2000).

Why does the zonal mean OHCT indicate only a northward discharge of heat during El Niño? The geometry difference between the North and South Pacific may play some role, but a key aspect is that the ITCZ resides between 5° and 10°N . The latter is the meteorological equator and forms the location of the monsoon trough, so that it is the southeast trades that matter in the equatorial zone. This is why the central El Niño region is from 20°S to 5°N in the earlier analysis. OHCT evolution in Fig. 3 shows that the ENSO-related signals are not symmetric across the equator: the west Pacific changes are more prominent in the north, but the signals in the east Pacific extend well westward on the equator and are more prominent in the south. When calculating zonal means, changes south of 5°N are dominated by east

FIG. 14. Schematic illustration of the ocean heat change during El Niño in the tropics within 20°S and 20°N. The three slices show zonal mean temperature change regressed against ONI from the sea surface to 400 m (right side), meridional mean temperature change regressed against ONI over the global tropical belt (front), and mean temperature within 0–2000 m (corresponds to OHC for 0–2000 m, shown in Fig. 2) in each 1° by 1° grid (top). All fields are normalized by their maximum values, so the units are kelvins and all values are within the interval from –1 to 1 K. The green arrows indicate the anomalous heat flow within the ocean: a dashed arrow indicates its occurrence before El Niño, and a solid arrow indicates the anomalies during El Niño. The change of atmospheric meridional circulation (anomalous Walker-type circulations) in the tropics is also shown, with green curved arrows indicating the net surface flux changes (an arrow pointing into the ocean corresponds to the ocean gaining heat).

Pacific changes, but north of 5°N they are dominated by west Pacific changes. Besides, this might also relate to Sverdrup transports associated with equatorial wind stress curl anomalies (Jin 1997; Clarke et al. 2007). Also there is stronger seasonality in the Northern Hemisphere than in the Southern Hemisphere and ENSO is phase-locked with the annual cycle (Trenberth 1997).

Although we find that ocean dynamics are a major physical driver of OHC variability in the tropical Pacific/Indian Ocean, there are no direct observations for ocean heat convergence because of insufficient data on current velocity. Moreover, there are only nine El Niño events after 1985, although the different character of ENSO accounts for less in this study. LENS results also suggest that heat convergence (Figs. S2 and S3) dominates the ocean heat redistribution within the Pacific Ocean, although the LENS ENSO is too periodic compared with observations, resulting in more symmetric evolution of all fields before and after El Niño.

Despite the strong local heat redistributions in the Pacific Ocean, there is also strong net ocean cooling in the tropical Pacific with El Niño, mainly due to heat loss (evaporation) into the atmosphere (Figs. 12–14), and this forms the diabatic component of ENSO. During

El Niño, the high SST in the central and eastern tropical Pacific (Figs. 2, 6, and 8) increase the net heat loss into the atmosphere (mainly via evaporation) (Figs. 2, 6, and 8), setting the stage for anomalous convection over the central Pacific. Heavy P (evident in Figs. 1, 6, and 8), shifts the upward branch of the Walker circulation from the Maritime Continent to the central Pacific, and then the rest of the overturning atmospheric circulation is shifted around (illustrated in Fig. 14) (Lau and Yang 2002). The anomalous diabatic heating, both a positive latent heating anomaly in the central Pacific and a deficit of latent heating in western tropical Pacific, is the driver of the global teleconnections during ENSO.

The impact of ENSO on the tropical Atlantic and Indian Oceans is mainly from an “atmospheric bridge” (Trenberth et al. 1998; Alexander et al. 2002), as summarized in Fig. 14. Anomalous subsidence over the equatorial Atlantic is produced (Enfield and Mayer 1997; Klein et al. 1999; Mayer et al. 2014) with less P (Figs. 4, 8, and 14). The weakened local Hadley cell reduces surface evaporation in the tropical Atlantic Ocean ($Q > 0$) and increases tropical Atlantic SSTs and OHC (Figs. 2, 3, 5, 6, and 14). This also sets the stage for the year after the El Niño being exceptionally active for Atlantic hurricanes (Trenberth et al. 2018).

The winds in the tropical Indian Ocean are also coupled with ENSO (Fig. 14). Weakened trade winds in the tropical Pacific before El Niño contribute to weakening westerly winds in the Indian Ocean, which weakens the upwelling in the west Indian Ocean (thermocline feedback). Thus, the Indian Ocean shows warming in the west ($\text{OHC} > 0$) and cooling in the east ($\text{OHC} < 0$) during El Niño. This Indian OHC change (Figs. 2 and 3) is an imprint of the evolution of positive Indian Ocean dipole events, which tend occur during El Niño events if the Indian Ocean background conditions are favorable (Ummenhofer et al. 2017; Mayer et al. 2018). Anomalous subsidence over the equatorial Indian Ocean also provides more surface downwelling solar radiation and lighter winds, which reduce evaporation (positive surface heat fluxes in Fig. 13), to drive a net SST and OHC increase (Figs. 12 and 13), which is referred to as the Indian Ocean basin mode (Klein et al. 1999; Yu and Rienecker 1999; Du et al. 2009; Xie et al. 2016; Webster et al. 1999). The lead/lag difference in net Q and OHCT change highlights the reduced ITF during and after El Niño (Figs. 12 and 13).

5. Remote OHC change related to ENSO

How does ocean heat change outside the tropics, and how do regional ocean changes coordinate to determine the global change? Detailed analyses of the teleconnection patterns are outside the scope of this study [but see Alexander et al. (2002) and other studies].

a. OHC change outside the tropics

In the North Pacific Ocean (20° – 60°N), it appears that a small net ocean warming occurs about 5 to 10 months before El Niño (0.05 – 0.08 PW K^{-1}) (Fig. 12), with a similar change of the surface flux (0.04 – 0.08 PW K^{-1}) (Fig. 13). During El Niño, surface flux cools the ocean (from -0.05 to -0.02 PW K^{-1} at lag = 3; Fig. 13), and although the overall OHCT change is insignificant (Fig. 12), there are large regional changes with a characteristic pattern that arises from strong atmospheric teleconnections between the tropical and North Pacific Oceans (i.e., Trenberth et al. 1998; Alexander et al. 2002). Strong cyclonic flow around an anomalously deep Aleutian low during El Niño events cools the central North Pacific and warms the water along the west coast of North America. This fingerprint can be seen clearly in OHCT/OHC and SST patterns (Figs. 2 and 3) in the North Pacific and is also characteristic of the Pacific decadal oscillation (PDO). These local changes compensate and produce a small negative net surface flux in the North Pacific during El Niño that is compensated by net northward heat

transport from tropical Pacific to the North Pacific (Fig. 6).

While there is some symmetry between the North and South Pacific for the top 100 m, the OHCT is somewhat different for 100–300 m (Fig. 11) and the basin shape differences come into play. In the South Pacific (70° – 20°S) overall, OHCT observations suggest a transition from an ocean cooling (from -0.1 to -0.02 PW K^{-1}) 6 months before El Niño to a warming after El Niño (0.12 – 0.17 PW K^{-1} at lag within 5–15 months) for 1985–2016 (Fig. 12), corresponding to a similar evolution of Q with weaker magnitude: from -0.07 to -0.03 PW K^{-1} before El Niño and from -0.01 to 0.05 PW K^{-1} after El Niño (Fig. 13). As different data/periods show large variability in OHCT change in these regions, chaotic aspects come into play and, as in the North Pacific, there is large regional cancellation of opposing local changes (shown in Figs. 2 and 3) and also the observation density is lower here than in the north. LENS shows almost no change of OHC in the South Pacific Ocean and different phase of when Q compared with observations, marking a major difference between observations/reanalysis and model simulations.

The south Indian Ocean (20° – 65°S) shows a weak OHCT increase after El Niño (0.02 – 0.10 PW K^{-1}) (Fig. S5) at lag = 10 but Q -induced warming during El Niño (0.05 – 0.11 PW K^{-1}) near lag 0 (Fig. S6). This mismatch is consistent with the different spatial patterns (Fig. 3) and indicates the dominant role of ocean transport. This region is linked to the tropical Indian, Atlantic, and Pacific Oceans, and is worthy of more scrutiny to clarify its heat budget. Also the exact choice of latitudinal bounds for our diagnostics can affect results. The North and South Atlantic (20° – $65^{\circ}\text{N}/20^{\circ}$ – 65°S) show an insignificant and small ocean heat change related to El Niño (Fig. S5, for IAP, ORAS5, and LENS), consistent with insignificant Q anomalies (Fig. S6), but these results mask some significant local patterns (Figs. 2 and 3) whose signals cancel.

b. Global interannual OHC changes

As ENSO-related OHC signals at different locations (Figs. 12 and 13; see also Figs. S5–S7) compensate each other when doing a large area average, the signal-to-noise ratio is substantially reduced. Is there a robust ENSO signature in the global OHC record? Here we compare global OHCT time series from multiple observational records—an ocean objective analysis (IAP), reanalysis (ORAS5), and ocean surface flux (TF and DeepC) observations in Fig. 15—with ONI.

For OHC and OHCT, the two ocean datasets (IAP and ORAS5) show substantial differences before 2005, when Argo data became available globally, but more

FIG. 15. Global OHC change on the interannual scale. Shown are OHC and OHCT in the upper 2000 m from 1993 to 2016 from two data sources: IAP ocean analysis (solid curves) and ORAS5 reanalysis (dashed curves). Also shown are surface flux data using the residual method: TF data (dashed curves) from 2000 to 2016 and DeepC (solid curves) from 1993 to 2015. ONI is depicted with a black curve. OHCT is in watts per meter squared, averaged over the global surface ($5.1 \times 10^{14} \text{ m}^2$). The red and blue shading respectively mark the El Niño and La Niña events with $|\text{ONI}| > 0.5 \text{ K}$. The gray shading marks the period before 2005, i.e., before the Argo network is established.

similar after 2005. IAP data also show larger magnitude changes before 2005 than ORAS5. Cheng et al. (2017a) show that the signal-to-noise ratio of the global interannual OHC change is larger than 2 only after 2005. Therefore, for global OHC or OHCT, we focus on the post-2005 period. There is a decrease of OHC (OHCT < 0) during and after El Niño and an increase during La Niña (OHCT > 0). Regressions among observed OHCT (0–2000 m) and ONI after 2005 show a significant OHC decrease during El Niño for both ocean datasets (Fig. 16), with a maximum OHCT ranging from $-0.29 \pm 0.07 \text{ PW K}^{-1}$ (IAP) to $-0.08 \pm 0.08 \text{ PW K}^{-1}$ (ORAS5). Using OHCT at the upper 300 m gives similar results ($-0.15 \pm 0.08 \text{ PW K}^{-1}$ for IAP), as most of the ENSO impacts are at the upper 300 m (section 4c). Wu et al. (2019) suggested a global OHCT change of $\sim 0.12 \text{ PW K}^{-1}$ for 1992–2011 based on another ocean reanalysis product, similar to ORAS5. Because there are only three El Niño events after 2005, it is extremely difficult to reliably quantify either the global ENSO-induced OHC/OHCT changes or the agreement/disagreement among datasets for this short period. Therefore, we recommend revisiting the ENSO-related OHC changes in the future revealed by different

products, for example IAP and ORAS5 used in this study and also other analysis or reanalysis products (e.g., Roemmich and Gilson 2011; Trenberth et al. 2016; Wu et al. 2019).

Globally, the net surface flux should be approximately equal to OHCT (if vertical heat exchange at 2000-m depth is negligible) because the horizontal ocean heat transport cancels. TF and DeepC data show excellent consistency during their overlap period after 2000 (Fig. 15). After 2005, surface flux data are also reasonably consistent with OHCT (Fig. 15), especially ORAS5, confirming a robust ocean cooling during and after El Niño. Figure 16 provides the regression between surface flux and ONI for different datasets in different periods: DeepC 2000–15 and 1985–99 and TF 2000–16. All of the regressions show a buildup of ocean heat 5–15 months prior to the El Niño event, and significant cooling during and after El Niño starting 5 months before. The cooling effect from -0.10 to -0.13 PW K^{-1} persists for at least 7 months after the El Niño peak (shown in surface flux data in Fig. 15b). For the 1985–99 period, DeepC data show stronger ocean heat change related to ENSO and longer periods of fluctuation, highlighting the possible uniqueness of the 1997–98 event.

We also check the results in a climate model (LENS) with 100 years of a control simulation, although there are many insufficiencies: for example, LENS fails to exhibit the observed diversity of El Niño (i.e., CP). OHCT and surface flux in LENS data also suggest an ocean heat loss starting from 4 months before the El Niño ($\sim 0.06 \text{ PW K}^{-1}$) and a maximum cooling of $\sim -0.09 \text{ PW K}^{-1}$ at 6–8 months after the El Niño. LENS simulations are consistent with observations and reanalyses, revealing a robust ocean cooling in the ocean.

Therefore, we conclude that the entire ocean is warming during La Niña (from -15 to -5 months before the peak of El Niño) and cooling during and after El Niño (from -4 to 12 months with respect to the peak of El Niño), but the magnitude of the change varies.

6. Conclusions

This study uses direct ocean observations (ocean objective analyzed product), ocean reanalyses, an Earth system model, and surface flux data (derived from TOA observation and atmospheric reanalysis) to understand the local and global ocean heat evolution related to ENSO. Two ocean heat changes related to ENSO are involved: 1) adiabatic (heat redistribution within the ocean) and 2) diabatic (net ocean heat exchanges with the atmosphere) changes. In the tropical Pacific Ocean, ENSO is associated with strong adiabatic heat redistribution in the zonal, meridional, and vertical

FIG. 16. Regression between ONI and (a) global OHCT (ORAS5, IAP within 2005–16, and LENS data) and (b) global ocean surface flux (DeepC for the 2000–15 and 1985–99 periods, TF for 2000–16, and LENS data) over different periods. The shadings (or bars for LENS) show the 90% confidence interval.

directions, yielding regional OHC changes of opposite sign, and it is mainly when integrated over larger domains that the net diabatic cooling signal in the Pacific clearly emerges during El Niño events. However, the warming of the tropical Indian and Atlantic Oceans during the El Niño events largely offsets the Pacific cooling. The reinforcing and competing net heat changes in three major tropical oceans results in a net cooling over the global tropical bands, which peaks after the mature phase of El Niño events.

We also documented the changes outside the tropics during ENSO, the bridges to the North Pacific and Atlantic Ocean via the atmosphere, while the link to the Indian Ocean occurs both through the atmosphere and the ITF changes. The patterns of change are distinctive but quite complex, and also vary in time. The results here are limited in time by the available data, and the full spectrum of behavior is clearly much more extensive than we can document.

Here we examined the 0–2000-m OHC changes, but the results in this study agree with many previous studies in that most of the variability is confined to the upper 500 m (Fig. 9). Because there are definitely signals below 500 m for some events, this study chooses a very deep vertical layer to fully contain the ENSO signal. OHC changes using a smaller depth (0–300 m) show slightly smaller regression coefficients although some of the signals are lost and there are somewhat smaller uncertainty bounds. In the future, when more data are available, we recommend that the vertical structure of ENSO should be explored further.

We show that the global ocean (and Earth system) cools during and after El Niño due to strong tropical Pacific cooling, driven by SST anomalies that induce net surface flux increases and thus evaporative cooling of the ocean. The Argo network, fully established after

2005, helps to greatly increase the accuracy of the global OHC estimate and to bring all the different estimates into better agreement. It is important to recognize that there are large compensating effects, both in the vertical and laterally, and thus the domain over which any integration occurs can distort results. This is an important factor in accounting for the different views of ENSO's energetic role summarized in the introduction. Use of annual means can also distort results and the perspectives from both OHC and OHCT are valuable. The evidence suggests large variability across different ENSO events. The implication is that the current ocean observation system must be maintained and a dense upper ocean observing system is required to better resolve the net ocean heat changes. These are important also for adequate ENSO forecasts because the disposition of OHC can affect subsequent coupled atmosphere–ocean evolution.

In addition, we note that the key results in this study are robust across all employed datasets, and several of the datasets are completely independent from each other (e.g., IAP and the surface flux data; IAP/ORAS5 and LENS simulations). LENS results differ from the observations in some local changes, such as in the South and North Pacific Ocean. This confirms earlier studies that showed that the details of the ENSO-related OHC redistribution and exchanges in the model differ from what is found from observational data (Mayer et al. 2016). This discrepancy is now corroborated with more data. Maintaining these observational, reanalysis, and model efforts is key to extending our understanding of ENSO and other climate variability.

It will be valuable to carry out a full intercomparison of ENSO-related variations for OHC in different ocean reanalyses and statistical reconstructions, and for surface flux in different products using direct or indirect

methods. Even better would be assessing the ENSO–energy cycle using the full range of ocean–atmosphere historical datasets. This work is a first step in that direction in that it identifies key diagnostics that can be applied to multiple datasets.

Acknowledgments. This study is supported by National Key R&D Program of China (Grants 2017YFA0603202 and 2016YFC1401705). Author M. Mayer is funded by Austrian Science Fund project P28818; J. Fasullo was supported by NSF Award ID AGS-1243125 and DOE Award ID DE-SC0012711. The National Center for Atmospheric Research is sponsored by the National Science Foundation. ONI is provided at https://origin.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_v5.php. The data used in this study can be found in the following links: IAP data (<http://159.226.119.60/cheng/>), TF surface flux (available per request), DeepC data (<http://www.met.reading.ac.uk/~sgs02rpa/research/DEEP-C.html>), HadISST (<https://www.metoffice.gov.uk/hadobs/hadisst/>), and ORAS5 (<https://www.ecmwf.int/en/elibrary/18519-ocean5-ecmwf-ocean-reanalysis-system-and-its-real-time-analysis-component>).

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