On the Energy Exchange between Tropical Ocean Basins Related to ENSO*

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ABSTRACT
Vast amounts of energy are exchanged between the ocean, atmosphere, and space in association with El Niño–Southern Oscillation (ENSO). This study examines energy budgets of all tropical (30°S–30°N) ocean basins and the atmosphere separately using different, largely independent oceanic and atmospheric reanalyses to depict anomalous energy flows associated with ENSO in a consistent framework. It is found that variability of area-averaged ocean heat content (OHC) in the tropical Pacific to a large extent is modulated by energy flow through the ocean surface. While redistribution of OHC within the tropical Pacific is an integral part of ENSO dynamics, variability of ocean heat transport out of the tropical Pacific region is found to be mostly small. Noteworthy contributions arise from the Indonesian Throughflow (ITF), which is anticorrelated with ENSO at a few months lag, and from anomalous oceanic poleward heat export during the La Niña events in 1999 and 2008. Regression analysis reveals that atmospheric energy transport and radiation at the top of the atmosphere (Rad\_TOA) almost perfectly balance the OHC changes and ITF variability associated with ENSO. Only a small fraction of El Niño–related heat lost by the Pacific Ocean through anomalous air–sea fluxes is radiated to space immediately, whereas the major part of the energy is transported away by the atmosphere. Ample changes in tropical atmospheric circulation lead to enhanced surface fluxes and, consequently, to an increase of OHC in the tropical Atlantic and Indian Ocean that almost fully compensates for tropical Pacific OHC loss. This signature of energy redistribution is robust across the employed datasets for all three tropical ocean basins and explains the small ENSO signal in global mean Rad\_TOA.

1. Introduction
The primary mode of global climate variability, El Niño–Southern Oscillation (ENSO), modulates the energy budget of both the ocean and the atmosphere in various ways. Ocean heat content (OHC) in the Pacific is redistributed within the basin (e.g., Roemmich and Gilson 2011) and altered via anomalous surface energy exchanges.

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pathways of energy through the ENSO cycle including both the atmosphere and the ocean is lacking (Trenberth and Fasullo 2012). The limitations of ocean data are one reason for this circumstance. While investigation of atmospheric energy budget variability associated with ENSO has made substantial advances because of the availability of global atmospheric reanalyses, ocean datasets for a long time have been much more limited in spatial and temporal coverage [e.g., data from the Tropical Ocean and Global Atmosphere (TOGA) Tropical Atmosphere Ocean (TAO) project array], and hence most observation-based quantitative studies were focused on the equatorial Pacific (e.g., Meinen and McPhaden 2000; Hasegawa and Hanawa 2003; Clarke et al. 2007).

The advent of new ocean reanalyses in recent years and advances in ocean data coverage due to the introduction of Argo (a global array of currently more than 3500 temperature and salinity profiling floats) in the 2000s should facilitate this attempt but data quality remains an issue. Balmaseda et al. (2013b) have shown that Ocean Reanalysis System 4 (ORAS4; Balmaseda et al. 2013a) consistently reflects the impact of major volcanic eruptions and the earth’s energy imbalance in global accumulated OHC. However, results of Loeb et al. (2012) and Trenberth et al. (2014) suggest that current ocean datasets practically fail to satisfy global constraints such as replication of anomalous radiation at the top of the atmosphere (RadTOA) measured from satellites in observed OHC variability. Uncertainty and resulting noise in ocean datasets is still too large to resolve small global mean anomalies of OHC tendencies (OHCT) on the order of ±1 W m\(^{-2}\) on interannual time scales. However, Loeb et al. (2012) found ENSO signals in global mean net radiation at the top of the atmosphere (TOA) on the order of ±0.5 W m\(^{-2}\) (corresponding to ±0.25 PW).

Here we employ three ocean reanalyses with different data assimilation approaches [ORAS4, the Hadley Centre third-generation Ensemble-Based Predictions of Climate Changes and Their Impacts (ENSEMBLES) project data (EN3) version 2a (herein HEN3; Ingleby and Huddleston 2007), and the Ensemble Coupled Data Assimilation System (ECDA; Zhang et al. 2007)] and two third-generation atmospheric reanalyses [European Centre for Medium-Range Weather Forecasts (ECMF) Interim Re-Analysis (ERA-Interim, hereafter ERA-I; Dee et al. 2011) and Modern-Era Retrospective Analysis for Research and Applications (MERRA; Rienecker et al. 2011)] from 1979 to 2012 to explore the energy budgets of the three tropical ocean basins and the tropical atmosphere in one consistent framework. Using these datasets and state-of-the-art budget methods we will show that on the scale of individual basins the signal-to-noise ratio in OHC data is sufficiently high to obtain robust signals of anomalous energy exchanges and transports within as well as between ocean basins in association with ENSO. Specifically, we will show that OHC changes in the tropical Pacific associated with ENSO are much larger than the response of Rad\(_{\text{TOA}}\) to ENSO. This apparent discrepancy can be explained by compensating OHC responses to ENSO in other tropical ocean basins that are of opposite sign compared to the tropical Pacific OHC response to ENSO. Atmospheric circulation and associated surface flux changes are responsible for these apparent energy exchanges across tropical ocean basins.

2. Data and methods

We investigate the variability of the energy budgets of atmosphere and ocean in a vertically integrated framework.

In an atmospheric column, net energy flux through the atmospheric lower boundary (\(F_S\); positive downward) is balanced by the local tendency of total atmospheric energy (AET), the divergence of horizontal atmospheric transports (DIVFA), and net radiation transport out of the column (DIVFA), and net radiation at TOA (throughout this article this means the sum of all radiation components):

\[
F_S = -\text{AET} - \text{DIVFA} + \text{Rad}_{\text{TOA}}. \tag{1}
\]

We compute \(F_S\) as a residual from the right-hand side of Eq. (1). The energy tendency AET is calculated from analyzed ERA-I fields, although the variability of AET in the tropics is small and almost negligible on interannual time scales. The divergence of atmospheric energy transport (DIVFA) is computed with the direct method from analyzed atmospheric state quantities as described in Mayer and Haimberger (2012) employing data from ERA-I and MERRA. Radiation at TOA is known to be biased in ERA-I forecasts (see, e.g., Berrisford et al. 2011), but Mayer and Haimberger (2012) showed that monthly anomalies agree well with independent satellite data from Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996). Hence, we mainly employ Rad\(_{\text{TOA}}\) from ERA-I because of the much longer period covered, but all computations are also performed with CERES data to confirm the results at least for the shorter commonly covered period (here March 2000–February 2012). We also tested Rad\(_{\text{TOA}}\) from MERRA for 1979–2012, but it was much less homogeneous in time and thus excluded from this study.

In an oceanic column, extending from the sea surface to the ocean bottom, net surface flux \(F_S\) is balanced by the local tendency of ocean heat (more precisely: enthalpy) content (OHCT) and the divergence of horizontal ocean
heat (more precisely: enthalpy) transport (DIVFO) out of the column (variability of other forms of energy in the ocean is negligible in this context):

\[ F_S = \text{OHCT} + \text{DIVFO}. \]

(2)

For the evaluation of OHCT in Eq. (2) we use three state-of-the-art oceanic datasets covering 1979–2012. ORAS4 employs an ocean model for data assimilation and ERA-I surface fluxes as boundary conditions. As we here aim to obtain independent estimates of atmospheric–oceanic energy exchanges, we additionally employ HEN3 and ECDA. The datasets partly differ in input data, but also in their data assimilation approaches. HEN3 represents an objective analysis purely from in situ data and hence is fully independent of atmospheric data, and the ECDA dataset represents a coupled data assimilation approach using the fully coupled Climate Model, version 2.1 (CM2.1).

The monthly-mean OHCT in Eq. (2) would ideally be obtained from the difference of OHC snapshots at 0000 UTC on the first day of two consecutive months to be perfectly consistent with the monthly averages of the other terms in Eqs. (1) and (2). However, generally no ocean temperature snap shots are provided with ocean datasets and thus OHCT is obtained from centered differences of monthly-mean values of OHC. On the time scales considered here, the introduced error is small. OHC is computed as

\[ \text{OHC} = c_p \rho_0 \int_0^{\text{depth}} \theta(z) \, dz, \]

(3)

with the heat capacity of seawater \( c_p = 3990 \text{ J kg}^{-1} \text{K}^{-1} \), the density of seawater \( \rho_0 = 1026 \text{ kg m}^{-3} \) (both values are not independent of temperature, but their product tends to be; Trenberth and Fasullo 2008), and ocean potential temperature \( \theta \). The lower integration bound is the sea surface height above a reference level. Tests have shown that variations of sea surface height on interannual time scales are almost entirely due to steric effects and thus can be neglected when using constant density. This is a common approximation in physical oceanography (see, e.g., Trenberth and Fasullo 2008). The upper integration bound (depth) ideally represents the ocean bottom.

It would certainly be perfectly consistent to integrate OHC to the bottom of the ocean everywhere, but in general, data quality decreases with depth as the number of measurements decreases with depth. Hence, integration depth is a compromise between data quality and physical consistency. We tested several values for the integration depth (300, 400, 700, and 2000 m, and the ocean bottom) and the agreement among the datasets in terms of variability was best for a depth of 300 m. Moreover, the satisfaction of global constraints such as the replication of \( \text{Rad}_{\text{TOA}} \) anomalies in OHCT was best, albeit certainly not perfect, for an integration depth of 300 m. As will be shown below, for depths greater than 300 m the increase of noise in the data outweighs the increase in the signal. This is especially true for the linear relation between OHCT and ENSO, which we will focus on in the following sections.

Regarding DIVFO, the main focus of this paper is on ocean heat exports out of the three tropical ocean basins. We can estimate these exports (i.e., area-integrated DIVFO) from the directly computed ocean heat transports across the lateral boundaries of the basins. These are available from ORAS4 and we employ full-depth ocean heat transport across selected ocean cross sections: Pacific (1979–2012) and Atlantic (1979–2009) ocean heat transport at 30°N and 30°S, Indian Ocean heat transport across 30°S (1979–2009), and heat transport by the Indonesian Throughflow (ITF; 1979–2012). At least at the considered cross sections, ocean heat transport variability below 300 m is small, and especially below 700 m it is very close to zero and in contrast to OHCT the noise does not increase with depth (see Fig. S1 in the supplementary material).

An alternative way to evaluate DIVFO is to compute it indirectly from \( F_S \) and OHCT (as described in Trenberth and Fasullo 2008). In our case, the indirect estimate yields six different DIVFO estimates (every possible combination of employed OHC and \( F_S \) data). The major disadvantage of this method is the accumulation of uncertainties from all employed fields (DIVFA, \( \text{Rad}_{\text{TOA}} \), \( \text{AET} \), and OHCT) in DIVFO. As a consequence the global mean value of DIVFO and especially its anomalies are not equal zero, which should be the case when neglecting contributions from river discharge and calving glaciers. Moreover, Eq. (2) holds only if OHCT is integrated to the ocean bottom. If OHCT is integrated only to a depth of 300 m (to avoid noise from below 300 m), a heat flux term across a depth of 300 m has to be added in Eq. (2). Since this flux is not known it contributes to the uncertainty of the indirect DIVFO estimate. Tests showed that indirectly estimated DIVFO is not distinguishable from noise on the scale of tropical ocean basins (30°N–30°S), independent of integration depth (see Fig. S2 in the supplementary material), but the signal-to-noise ratio is much better when considering smaller regions with no spatially compensating anomaly structures (e.g., equatorial Pacific). Overall, we found the direct approach to yield better results on a basin-integrated scale, where anomaly signals are generally weak because of compensating processes within the respective basins.
All anomaly fields and series shown in the present article represent monthly anomalies with the annual cycle removed and are 13-point filtered in time (Trenberth et al. 2007). Fields have been detrended before correlation and regression analysis.

3. Results

a. Tropical Pacific and Atlantic budget variability

We first consider time series of budget relevant terms over the Pacific and Atlantic oceans. Over the Pacific (Fig. 1a), the largest variability is found from DIVFA with stronger (weaker) atmospheric energy export and OHCT with heat loss (gain) during warm (cold) ENSO phases. This behavior is most conspicuous during the strong El Niño events in 1982/83, 1987, 1997/98, and 2009/10. Compared to DIVFA, the time series of OHCT exhibits considerable noise, but the ENSO-related heat discharge/recharge is clearly visible. Radiation at TOA over the tropical Pacific also exhibits noticeable variability associated with ENSO, with weaker (stronger) net energy input during El Niño (La Niña), which is a result of the strong ENSO-related anomalies of outgoing longwave radiation in the subtropics (Trenberth et al. 2010). Comparison of Rad_TOA from ERA-I and independent CERES data for the limited period of the satellite dataset shows very good agreement between the two series (Fig. 1a; correlation coefficient $r = 0.95$), suggesting Rad_TOA anomalies from ERA-I to be reliable also prior to 2000. Significant correlation is found between DIVFA, OHCT, and Rad_TOA and the Niño-3.4 index [N34, anomaly index (in units of kelvin) of SSTs area averaged over $5^\circ$S–$5^\circ$N, 170°–120°W; see Fig. S3a in the supplemental material].

Variance of the Pacific oceanic heat export (i.e., the sum of ITF heat transport and poleward ocean heat export; here defined as difference of northward transport across 30°N and 30°S) is generally small. However, the ITF heat transport shows a slight weakening following El Niño events, in accordance with England and Huang (2005), and it is significantly correlated with N34 at a 6-month lag ($r = -0.65$). Poleward ocean heat export shows no significant correlation with ENSO ($r = 0.22$).

![Fig. 1. Time series of budget-relevant fields area integrated over (a) the tropical Pacific (30°N–30°S) and (b) the Atlantic (PW). Shown are 13-point-filtered anomalies (annual cycle removed) with the respective long-term mean added. For the ERA-I curve, the CERES mean is added. The ITF heat transport is defined positive westward. The vertical light red and blue bars indicate warm and cold ENSO events, respectively, as defined by the National Oceanic and Atmospheric Administration.](image-url)
but nevertheless there can be seen some variability, for example positive anomalies during the 1999 and 2008 La Niña events.

Altogether, this suggests that Pacific Ocean heat loss and gain associated with ENSO is mainly balanced by atmospheric energy export and also Rad\textsubscript{TOA}. Ocean heat transport apparently plays a minor role in modulating basin-integrated OHC, with the strongest contribution from the ITF.

The behavior of DIVFA and OHCT in response to ENSO over the Atlantic Ocean is opposite compared to the Pacific (Fig. 1b), which is a result of the opposite response of the Atlantic Hadley cell to ENSO compared to the Pacific Hadley cell (see Klein et al. 1999). Atmospheric energy export is weaker (stronger) and the ocean gains (loses) heat during El Niño (La Niña). Both DIVFA and OHCT over the Atlantic are significantly correlated with N34 (see Fig. S3b in the supplementary material). Atlantic Ocean heat export varies weakly, and the correlation with N34 is very low ($r = -0.20$). Radiation at TOA from ERA-I also exhibits very small anomalies, in agreement with CERES. As for the Pacific, this suggests a balance between the divergence of atmospheric energy transport and Atlantic OHC. A quantitative assessment of the qualitative findings in this section will be given in sections 3d and 3e.

Despite the good qualitative agreement of the curves in Figs. 1a,b it is noted that the anomaly budget is generally not perfectly closed [i.e., the sum of the atmospheric and oceanic anomalies do not exactly add up to zero in the respective basins as would be required from the budget Eqs. (1) and (2); e.g., after the eruption of Mt. Pinatubo]. One reason for the observed discrepancies between atmospheric and oceanic budget anomalies is the absence of signals from major volcanic eruptions El Chichón (1982) and Mt. Pinatubo (1991) in Rad\textsubscript{TOA} from ERA-I, while ORAS4 clearly shows the impacts of these eruptions (see Fig. 1a; Trenberth et al. 2014). In contrast to Rad\textsubscript{TOA} from ERA-I, analyzed atmospheric fields in ERA-I such as stratospheric temperature do show the impact of volcanic eruptions (Simmons et al. 2014), which necessarily introduces inconsistencies between Rad\textsubscript{TOA} and other
atmospheric and oceanic fields. However, results in the following sections will show that these imbalances are uncorrelated with ENSO.

b. Regression analysis

To explore the linear relationship between energy budget terms with ENSO more rigorously, we consider local regression coefficients of DIVFA, $F_s$, OHCT, and sea surface temperature with N34 at zero lag, presented in Figs. 2a–d, respectively. First we focus on the tropical Pacific energy budget variability with its strong ENSO imprint. Regression fields of DIVFA and $F_s$ both show anomalies of opposite sign over the western and eastern Pacific, respectively (see also Mayer et al. 2013). Note the high structural and quantitative agreement of the two fields indicating their tight relationship (i.e., anomalous atmospheric energy transports are largely driven by surface fluxes). Compared to $F_s$, regressed Pacific OHCT exhibits different structures, which represent the signature of ocean heat redistribution within the ocean (deepening thermocline and hence increase of ocean heat content in the eastern Pacific in association with El Niño and vice versa for La Niña), a necessary prerequisite for increased interaction of the ocean with the atmosphere via SST (Roemmich and Gilson 2011). Additionally, negative OHCT regression coefficients much stronger than the surface flux coefficients can be found along the equator at all Pacific longitudes, indicative of OHC changes due to Sverdrup transports in association with El Niño and La Niña (Jin 1997). This aspect will be discussed in section 3c.

Scatterplots of N34 versus tropical Pacific DIVFA anomalies from ERA-I and OHCT anomalies from ORAS4 are presented in Figs. 2e and 2f, respectively. They confirm the findings from Fig. 1a: The regression lines show similar slopes but of opposite sign suggesting that ocean heat feeds the atmospheric energy transport anomalies.

Tropical Atlantic local regression coefficients (Figs. 2a–d) are more uniform than those over the Pacific. Uniformly negative coefficients are found for DIVFA, which is a result of a negative correlation between Atlantic Hadley cell strength and ENSO (Wang 2005). As in the Pacific, net surface flux regression coefficients are again very similar to those of DIVFA (but of opposite sign). Ocean heat content tendency shows moderate positive coefficients nearly everywhere in the tropical Atlantic, consistent with the surface flux regression field (Fig. 2b), the well-known Atlantic SST maximum 5 months after El Niño (Enfield and Mayer 1997), and with Lohmann and Latif (2007), who found that El Niño–related Atlantic SST warming occurs because of surface fluxes rather than ocean dynamics. Scatterplots of area-averaged tropical Atlantic (30°S–30°N) DIVFA and OHCT versus N34 (Figs. 2g,h) again confirm findings from Fig. 1b; namely, negative (positive) DIVFA and positive (negative) OHCT anomalies in association with warm (cold) ENSO states.

Anomalous divergent atmospheric energy transports from the eastern Pacific to the Indo-Pacific warm pool and the Atlantic, represented by the vectors in Fig. 2a, indicate the anomalous energy transport by the anomalous Walker cell, suggesting an energy exchange between the tropical Pacific and Atlantic Oceans via total energy transport within the atmosphere. This picture will be corroborated in section 3d.

c. Budget of the equatorial Pacific

So far we have not considered estimates of DIVFO, but these are potentially as large as the other terms. As already found from Figs. 2a–c, structures of locally regressed OHCT are much stronger than those of $F_s$ and DIVFA, which must be due to DIVFO. To demonstrate the dominance of ocean heat transports along the equatorial Pacific compared to atmospheric transports, we present regression coefficients of $\text{Rad}_{\text{TOA}}$, DIVFA, and OHCT from different datasets with N34 at different lags in Fig. 3. Confidence intervals are estimated from the residual sum of squares, taking autocorrelation into account (following Oort and Yienger 1996). Atmospheric energy export is in phase with N34 and responds with $0.13 \pm 0.02 \text{PW K}^{-1}$ to ENSO. The $\text{Rad}_{\text{TOA}}$ response is negligible. The ensemble mean ocean heat content tendency response shows much stronger values of opposite sign ($-0.49 \pm 0.10 \text{PW K}^{-1}$ at lag 3). The difference between DIVFA and OHCT response (on
the order of 0.36 PW K⁻¹) indicates that OHC variations along the equator in association with ENSO must be mainly balanced by ocean heat transport, while smaller amounts of energy are exchanged at the surface. This is consistent with the recharge oscillator concept by Jin (1997), which predicts that equatorial OHC anomalies are dominated by anomalous Sverdrup transports. Westerly wind anomalies responsible for these transports are strongest during the ENSO-related SST peak (Clarke et al. 2007). Hence, during peak ENSO (peak of SST anomalies) equatorial OHC changes are strongest both because of ocean dynamics (Sverdrup transports) and surface fluxes (mainly evaporation, but also surface radiation), but DIVFO quantitatively dominates in this region. However, as will be demonstrated below, DIVFO anomalies related to ENSO almost vanish when averaged over 30°S–30°N.

d. Quantification of energy exchanges between the tropical basins

Tropical Pacific regression coefficients of Rad₉OA, DIVFA, and OHCT from different datasets with N34 at different lags are presented in Fig. 4a. At zero lag, the Pacific atmosphere exports more energy than normal on the order of 0.20 ± 0.04 PW K⁻¹ during warm ENSO events. This is the extra energy driving global teleconnections (Trenberth et al. 2002a). At the same time, area-averaged Rad₉OA responds comparatively weakly, on the order of −0.06 ± 0.04 PW K⁻¹, in agreement with results from CERES for the shorter period covered (not shown). The sum of these two values is in excellent agreement with associated OHC changes of −0.26 ± 0.07 PW K⁻¹. OHCT regression coefficients from the three employed datasets agree very well despite the different assimilation approaches, probably as a result of the relatively dense ocean observation system in this region. Most convincingly, the combined DIVFA and Rad₉OA regression coefficients from ERA-I at zero lag (0.26 PW K⁻¹) are perfectly balanced by the OHCT regression coefficient obtained from the fully independent HEN3 dataset (−0.26 PW K⁻¹). This indirectly indicates that the contribution of DIVFO variability to the Pacific energy balance variability should be close to zero. We also computed OHCT regression coefficients for an integration depth of 700 m and found the signal to be increased by 10% on the expense of 50% increased noise (−0.29 ± 0.11 PW K⁻¹; see Fig. S4 in the supplementary material).

Additionally, we also performed a lagged regression analysis of ocean heat transport by the ITF and Pacific poleward heat export computed directly from

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**Fig. 4.** As in Fig. 3, but for (a) tropical Pacific (30°S–30°N), (b) tropical Atlantic (30°S–30°N), (c) Indian Ocean (30°S–30°N), and (d) all tropical ocean basins (30°S–30°N, 360° of longitude). The yellow dot in (a) represents the regression coefficient of DIVFA at lag 0; that is, the same value as the yellow dot in Fig. 2e.
temperature and velocity fields of ORAS4, presented in Fig. 5. While poleward ocean heat export shows insignificant response to ENSO at all lags, the response of heat transport through the ITF is small at zero lag, but considerable at positive lags ($-0.10 \pm 0.03$ PW K$^{-1}$ at a 6-month lag). Thus, the energy balance around the 6-month lag has to include also the ITF heat transport to be reasonably closed. The sum of the regression coefficients of DIVFA, Rad$_{TOA}$, OHCT, and ITF heat transport is close to zero at all lags (not shown). It is important to note that this balance does not necessarily hold for individual ENSO events, but it holds for the average response of the energy budget to ENSO.

We stress again the vital importance of OHC redistribution within the Pacific due to ocean heat transports but that strong mutual cancellation of DIVFO anomalies leads to a small residual signal in area-averaged DIVFO.

Regression curves for the tropical Atlantic ($30^\circ$S-30$^\circ$N) are presented in Fig. 4b. Compared to the Pacific Ocean, the response of Atlantic OHCT and DIVFA has a similar magnitude (larger regression coefficients in an area-specific sense, smaller coefficients in the area integral because of the smaller basin size), but opposite sign ($+0.12 \pm 0.04$ and $-0.14 \pm 0.04$ PW K$^{-1}$, respectively). The Rad$_{TOA}$ response is very small and Atlantic Ocean heat export computed directly from ORAS4 shows no significant ENSO response at lags less than one year (Fig. 5). Thus, we conclude that in the tropical Atlantic ENSO-related variability in DIVFA is almost entirely balanced by OHCT. Note also that no net surface flux data have been employed to obtain this result [$F_3$ is eliminated when subtracting Eqs. (1) and (2)].

Analogous analysis for the Indian Ocean (Fig. 4c and Fig. S3c in the supplementary material) suggests a similar balance between OHCT and DIVFA, plus a small contribution from Rad$_{TOA}$, but the fields are less consistent than in the Atlantic case ($+0.08 \pm 0.05$, $-0.06 \pm 0.03$, and $+0.02 \pm 0.01$ PW K$^{-1}$, respectively). The OHCT regression coefficients from the different datasets differ considerably and the Indian Ocean heat transport across $30^\circ$S appears to compensate ITF transport anomalies to some degree (not shown), but it is unclear how robust this result is. Hence, uncertainties appear higher for the Indian Ocean compared to the other tropical basins.

Finally, we consider the sum of the three tropical ocean basins (Fig. 4d). At zero lag, we find almost perfect mutual compensation of DIVFA anomalies over the three basins. Hence, the positive response of poleward divergent atmospheric energy transports from the central and eastern tropical Pacific region as indicated by Fig. 2a is compensated by a negative response of poleward atmospheric energy transports over the Indo-Pacific warm pool and the Atlantic. There is, within uncertainty bounds, no net atmospheric energy export anomaly from the tropics during the peak ENSO phase. This confirms the results of Mayer and Haimberger (2012) finding only weak correlation of zonal mean tropical atmospheric energy export with ENSO. OHC changes in the Indian and Atlantic Oceans for the most part compensate Pacific OHC changes. Only for lags around 6 months, there is a net atmospheric energy export anomaly from the tropics and also radiative energy loss at TOA (in quantitative agreement with CERES, not shown). The atmospheric energy export a few months after the El Niño peak phase probably is a result of increasing SSTs in the Atlantic and Indian Oceans reinvigorating the regional Hadley cells. However, these signals are relatively weak and there is considerable uncertainty (significant correlation only for MERRA DIVFA, but not ERA-I DIVFA, see Fig. S3d in the supplementary material), but Rad$_{TOA}$, DIVFA, and OHCT still agree within (large) error bounds.

e. Constrained regression

In section 3d we found that the sum of regression coefficients of those terms in Eqs. (1) and (2), which are significantly different from zero, is close to but not exactly zero when considering basin-averaged series. This closure is remarkable and it is certainly desirable to obtain budget closure without any constraints. However, in this section we illustrate how to obtain numerically exact budget closure by computing the basin-averaged regression coefficients under the strong constraint of adding up to zero. Similar to the approach of Bollmeyer and Hense (2014) to close the atmospheric energy budget, the idea here is to minimize
Table 1. Regression coefficients of all budget terms at zero lag for tropical Pacific from ordinary least squares (OLS) and constrained OLS with 95% confidence intervals (PW K\(^{-1}\)). Boldface values are significantly different from zero. Uncertainties for the constrained OLS results are estimated with a block bootstrap method (length of the blocks is set to 6 months).

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<td>(-\text{Rad}_{\text{TOA}})</td>
<td>(0.06 \pm 0.04)</td>
<td>(0.07 \pm 0.04)</td>
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<tr>
<td>(\text{DIVFA})</td>
<td>(0.20 \pm 0.04)</td>
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<td>(\text{AET})</td>
<td>(-0.01 \pm 0.02)</td>
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<td>(\text{OHCT (0–300 m)})</td>
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<td>(\text{OHCT (0 m–bottom)})</td>
<td>(-0.28 \pm 0.13)</td>
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<td>(-\text{ITF})</td>
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4. Discussion and conclusions

Ample changes in circulation and associated changes in clouds, precipitation, and radiation in response to ENSO (as described in section 1 and references given therein) lead to an indirect redistribution of OHC across tropical ocean basins via the atmosphere. Most of the extra heat released from (stored in) the tropical Pacific during El Niño (La Niña) indirectly leads to increasing (decreasing) OHC in the other tropical basins. During peak ENSO conditions, tropical Atlantic OHC changes account for about 45% of Pacific OHC changes and Indian OHC changes account for about 30%, although the tropical Atlantic and Indian Oceans constitute only about 12% and 11% of the total ocean area, respectively. Accompanied by altered regional Hadley cells over the respective basins and anomalous zonal circulations between the basins, horizontal atmospheric energy transports (DIVFA) show variations quantitatively consistent with OHC changes, especially over the Atlantic.

Contribution of the ITF ocean heat transport variability to tropical Pacific OHC changes is not negligible and has to be included to reasonably close the tropical Pacific energy budget. Poleward oceanic heat export from the tropical Pacific shows little ENSO correlation. Hence, while anomalous ocean heat transports in association with ENSO are large within the tropical Pacific (see section 3c), the largest part of basin-averaged OHC changes is realized through surface fluxes and not through lateral ocean heat transports out of that region. A similar conclusion can be drawn for the tropical Atlantic, with the addition that there ENSO-related ocean heat transports are relatively small also within the basin (in contrast to the Pacific).

The small regression coefficient of Pacific poleward ocean heat export does not necessarily exclude that ENSO-related export anomalies may happen occasionally. Inspections of the time series in Fig. 1a show positive ocean heat export anomalies out of the tropical Pacific in association with the strong La Niña events in 1999 and 2008. Whether these anomalies are a random occurrence, related to the nonlinearities of strong ENSO events, or reflect the interaction with any other mode of the climate system (such as the Pacific decadal oscillation) is an interesting question but beyond the scope of this paper.

Variability of tropical Pacific \(\text{Rad}_{\text{TOA}}\) is correlated significantly with ENSO \((r = -0.57\) at a 6-month lag), but according to our calculations only about 20% of the anomalous tropical Pacific OHC is exchanged at TOA, mainly in the Pacific subtropics, and ENSO-related \(\text{Rad}_{\text{TOA}}\) anomalies over the other basins are negligible. Hence, ENSO-related variability of atmospheric energy export from the tropical Atlantic region is energetically...
strongly coupled to tropical Atlantic OHC (via net surface fluxes), while net radiation at TOA plays a negligible role in this balance. Overall, our results explain the comparatively small global mean TOA signal of ENSO, found by Loeb et al. (2012).

It is remarkable how distinct and robust obtained results are despite known shortcomings in the data (e.g., the missing signature of volcanic eruptions in RadTOA from ERA-I). Tests also showed that splitting background climatologies for computation of anomalies to reduce temporal data inhomogeneities as carried out by Mayer et al. (2013) resulted in slightly higher correlations.

Regression and correlation, albeit highly significant, of course do not prove causality, but our results are consistent with the physical understanding of the response of the tropical ocean–atmosphere system to ENSO. The presented pathways of energy are summarized in a schematic exemplary for the Pacific and Atlantic basins in Fig. 6. We also note that the presented results describe the first-order linear response of the energy budgets to an average canonical ENSO event and naturally neglect nonlinear effects of strong events, asymmetries between warm and cold events, as well as the different flavors of single ENSO events. Yet, budget closure for single events proves elusive with current datasets.

Global climate models, by construction, satisfy global mean consistency requirements such as replication of anomalous TOA fluxes in OHCT on interannual time scales (Wong et al. 2006), while current oceanic reanalysis datasets practically fail to do so (Loeb et al. 2012; Trenberth et al. 2014). As can be seen from our results, this is due to compensating processes in association with climate variability, which only leave a small global mean residual signal. However, here we demonstrate that, when carefully selecting regions to maximize signal-to-noise ratio, largely independent observation-based datasets of the atmosphere and ocean exhibit very good agreement. Thus, these results provide a new quantitative benchmark for the replication of ENSO-related energy flows and exchanges in coupled climate models. It is intended to be applied to model output from phase 5 of the Coupled Model Intercomparison Project (CMIP5) in the near future.

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FIG. 6. Schematic of anomalous energy fluxes within and between tropical atmosphere and oceans. Depicted is anomalous divergence of atmospheric energy transport; anomalous fluxes at the surface [including net radiation at the surface (Rad$_S$), latent heat flux (LH), and sensible heat flux (SH)] and at TOA; and anomalous heat transport by the ITF (yellow arrows), as well as anomalous SSTs and OHCT as a response to El Niño (based on regression with N34 at zero lag). Note the strong west–east compensation of OHCT in the Pacific, which is due to zonal oceanic heat transports. White arrows in the atmosphere indicate circulation anomalies in the vertical–equatorial plane associated with the Walker cells.


