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2 **Global ocean heat content redistribution during the 1998-**

3 **2012 Interdecadal Pacific Oscillation negative phase**

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Abstract

Previous studies have linked the slowdown in global surface temperature warming since 2000 to a negative Interdecadal Pacific Oscillation (IPO) phase. Here, we investigate the changes in ocean heat content (OHC) during this period. We compare two ensembles of coupled model experiments with either zero or observed prescribed tropical Pacific wind stress interannual anomalies. This successfully constrains the global surface temperature, sea level pressure and OHC patterns associated with the IPO phase transition around 1998. The negative IPO phase (1998 to 2012) is associated with a global ocean heat redistribution. The anomalously cold tropical Pacific Ocean leads to an increased oceanic uptake in this region, and a global OHC increase of 4×10^{22} J. The cold equatorial Pacific also forces mid-latitude wind changes through atmospheric teleconnections, leading to an enhanced wind-driven heat transport convergence at 40°N and 40°S. Enhanced Pacific easterlies also yield an enhanced heat transport to the Indian Ocean via the Indonesian throughflow. As a result, the anomalous Pacific heat uptake is entirely exported towards the North Pacific (~50%), Indian (~30%) and Southern (~20%) Oceans. A significant fraction of this heat is released back to the atmosphere in the North Pacific and Indian basins, and transported across 31°S in the Indian Ocean. Overall, OHC increases most in the Southern Ocean (~60% of global changes) and northern Pacific (~40%), with negligible changes in the Indian and Atlantic basins. These results point to the major importance of oceanic circulation in re-distributing the Pacific heat uptake globally during negative IPO phases.

Keywords: Decadal climate variability, Pacific Ocean, global warming, ocean heat content, air-sea interactions

1. Introduction

The continued increase in atmospheric greenhouse gas (GHG) concentrations since the Industrial Revolution induces a positive imbalance in the top-of-atmosphere (TOA) planetary energy budget (e.g. Hansen et al., 2005; Trenberth et al., 2009). The overwhelming majority of this excess energy is absorbed by the oceans, as they have a dominant heat capacity (Bindoff et al., 2013; Trenberth et al., 2014). Some of this heat also warms the atmosphere, leading to a steady rise in global mean surface temperature. In the 15-year period from around 1998 to 2012, however, the global mean surface temperature displayed a relatively weak warming rate compared to the previous decades (Easterling et al., 2009). This reduction in the rate of surface warming has been called the *hiatus* or *slowdown* in the global surface warming (Yan et al., 2016; Medhaug et al., 2017). Several recent studies largely attribute this change in the warming rate to natural climate variability, and more specifically to a prolonged abnormally cold sea surface temperature (SST) in the tropical Pacific Ocean (Meehl et al., 2011; Kosaka and Xie, 2013; Fyfe and Gillett, 2014; Clement and DiNezio, 2014; Trenberth, 2015; Dai et al., 2015).

The tropical Pacific Ocean is home to the El Niño Southern Oscillation (ENSO; e.g. McPhaden et al., 2006), the dominant climate mode on earth at interannual timescales (e.g. Deser et al., 2010). The ENSO negative phase, La Niña, is associated with a cold central and eastern Pacific, reduced convection in the central Pacific, and abnormally strong easterlies. The changes in tropical convection and associated convective heating during ENSO events induce remote changes in sea level pressure (SLP), surface winds and SST through atmospheric teleconnections, i.e. planetary waves towards the mid-latitudes channelled along mean-state dependent waveguides (e.g. Horel and Wallace 1981; Trenberth et al., 1998). This creates in particular the horseshoe-shaped cold SST anomaly pattern associated with warm anomalies in the northwestern and southwestern Pacific sectors during La Niña events. The

low-frequency tail of ENSO variations is associated with decadal-scale SST anomalies in the tropical Pacific, and the associated global SST pattern is called the Interdecadal Pacific Oscillation (IPO; Zhang et al., 1997; Power et al., 1999).

The IPO shifted from a warm to a cold phase around 1998, yielding a cooling trend of central and eastern Pacific SST, associated with an unprecedented intensification of the Pacific Walker circulation (Kociuba and Power, 2015; Choi et al., 2016) and strengthening of Pacific trade winds during the 20-year period beginning around 1992 (Balmaseda et al., 2013; England et al., 2014; de Boissésou et al., 2014). Several studies identified this negative IPO phase during the 1998-2012 period as an important driver of the hiatus, due to the associated internal rearrangement of heat within the climate system. A positive imbalance of $\sim 0.5\text{--}1\text{ W m}^2$ in the TOA planetary energy budget was indeed observed during this period (Loeb et al., 2012; Allan et al., 2014; Johnson et al., 2016), although volcanic activity and anthropogenic aerosols may have contributed to a small reduction in absorbed solar radiation (Solomon et al., 2011; Santer et al., 2014; Ridley et al., 2014; Smith et al., 2015). This observed imbalance implies that the total heat in the climate system steadily increased during the hiatus, and a smaller fraction of this heat was used to warm the atmosphere. Several studies suggested that the abnormally cold Pacific favored an enhanced oceanic heat uptake during this period (Meehl et al., 2011; Kosaka and Xie, 2013; Douville et al., 2015; England et al., 2014; Watanabe et al., 2014), with increased downward heat flux into the ocean in the tropical Pacific, yielding a larger global OHC increase in the subsurface during the negative IPO phase. Drijfhout et al. (2014) found that the ocean heat uptake was also modified in other basins, such as the Southern and subtropical Indian Oceans or the subpolar North Atlantic. Some of these studies used *pacemaker* experiments specifying observed anomalies of either SST (Kosaka and Xie, 2013; Douville et al., 2015) or wind (England et al., 2014; Watanabe et al., 2014; Douville et al., 2015) over the tropical or central/eastern Pacific. This

setup results in more realistic global SST patterns due to the remote impacts of Pacific SST anomalies through teleconnections, and a reduced rate of global surface temperature warming over the last two decades (e.g. Kosaka and Xie, 2013).

A burning question is thus: where did the extra heat pumped into the Pacific during the hiatus decade go? This question is not easy to tackle from observations alone. Deep ocean observations are indeed too sparse to track the OHC changes, even with improved recent estimates (Cheng et al., 2017). The deployment of Argo floats in the early 2000s greatly improved the sampling (Lyman, 2012; Abraham et al., 2013), but the current configuration of Argo floats only reliably samples the upper 2000m of the ocean. In addition, observed changes are a combination of the effects of global warming and natural decadal variability, so that it is not easy to isolate the heat redistribution related to negative IPO phases from observations only (Liu et al., 2016a). There is hence so far little consensus on the location of OHC changes associated with the IPO. Meehl et al. (2011) point to the deep ocean, below 750m in Atlantic and the Southern Ocean, and below 300m in the Pacific and the Indian Ocean. England et al. (2014) suggest an increased OHC in the upper 350m of the tropical Pacific. Some studies suggest that a large fraction of the Pacific heat uptake has been transferred to the Indian Ocean via the Indonesian Throughflow (Lee et al., 2015; Nieves et al., 2015; Liu et al., 2016a). However, other studies combining model and observations have pointed out that the La Niña-related Pacific and Indian OHC increases in upper layers are not enough to account for the early 21st century warming slowdown, and that OHC increases at deeper layers in the Atlantic and Southern Ocean account for most of the sequestration (Chen and Tung, 2014; Drijfhout et al., 2014; Roemmich et al., 2015; Llovel and Terray, 2016; Oka and Watanabe, 2017). Cheng et al. (2017) also verified that such a warming in the Atlantic and Southern Ocean has occurred in observations. Liu et al. (2016a, 2016b) have however argued that the deep Atlantic and Southern Ocean observed warming are due to

anthropogenic climate change, and not specifically to a negative IPO phase, a point which is contested by Chen and Tung (2016). The relative contributions of various ocean basins to OHC changes during the recent negative IPO phase and the associated processes thus remain controversial.

This study aims at quantifying the OHC changes in each basin during the early 21st century hiatus period, and the associated processes, partitioning the relative contributions from surface heat uptake and lateral redistribution by the oceanic circulation. We use a coupled model with a *pacemaker* setup in the tropical Pacific to constrain the global IPO pattern through atmospheric teleconnections, as in previous studies. We will show that the resulting coupled model global surface temperature, SLP and OHC variations match those related to the IPO in observations. The global anomalous response associated with the IPO is obtained as the difference with a twin simulation with identical radiative forcings in which the *climatological* tropical Pacific wind stress is imposed. We then use a heat budget approach to examine the global OHC changes and their causes.

The paper is structured as follows: section 2 discusses the modelling setup and methodology; section 3 describes the global signals remotely driven by the strengthened easterlies and the resulting cold anomaly in the tropical Pacific; section 4 evaluates the associated ocean heat content changes, and analyses the related surface heat uptake and oceanic heat transport. We summarize and discuss the implications of our results in Section 5.

2. Methods and Data

2.1 Coupled model

We use the coupled IPSL-CM5A-LR global climate model (Dufresne et al., 2013). The atmospheric component is LMDZ5A (Hourdin et al., 2013), with a resolution of $3.75^\circ \times 1.875^\circ$ with 39 vertical levels. The land surface component is ORCHIDEE (Krinner et al., 2005), with the same resolution as the atmosphere. The ocean component is NEMOv3.2 (Madec, 2008) coupled to the LIM2 sea ice model (Fichefet and Maqueda, 1997). It uses a resolution of about 2° , with a refinement to 0.5° meridional resolution near the equator, and 31 vertical levels.

IPSL-CM5A-LR simulates the El Niño–Southern Oscillation (ENSO) amplitude, structure, and spectrum reasonably (Bellenger et al., 2014), although it displays an incorrect seasonal phase locking. It simulates well the main characteristics of the North Pacific variability (Fleming and Anchukaitis, 2016), the links between ENSO and IPO (Nidheesh et al., 2017), the tropical intra-seasonal variability (Maury et al., 2013; Hung et al., 2013) and the North Atlantic variability (Cattiaux et al., 2013; Gastineau et al., 2013). However, like many global climate models, IPSL-CM5A-LR has a double intertropical convergence zone (ITCZ) bias and a too strong equatorial upwelling in the Pacific (Dufresne et al. 2013). Compared to other Coupled Model Intercomparison Project Phase 5 (CMIP5; Taylor et al., 2012) climate models, IPSL-CM5A-LR also has relatively large transient and equilibrium climate sensitivities to CO₂ doubling (Forster et al., 2013). This response is mainly due to a large positive low-cloud feedback (Brient and Bony, 2013; Andrews et al., 2012).

2.2 Experimental protocol

We perform two ensembles of partially-coupled simulations using IPSL-CM5A-LR from 1979 to 2014. In the two ensembles, we use the standard CMIP5 GHG, aerosol and ozone forcings (Taylor et al., 2012), following the historical experiments through 2005 and

RCP8.5 from 2006 to 2014. Volcanic stratospheric aerosol forcings are applied by prescribing the aerosol optical depth, as given in Sato et al. (1993; 2012 updated version). 30 members are run for each of the two ensembles. Each ensemble was initialised in 1979 from a 30-member ensemble of fully-coupled historical simulations starting in 1940, described in more detail in Frankignoul et al. (2017).

We perform two ensembles, with a pacemaker approach designed to constrain the Pacific interannual variability. The first ensemble, W-CLIM, is meant as a reference representative of a neutral IPO phase. The second ensemble, W-FULL, is constrained to follow the observed IPO evolution. In these simulations, the wind stress is prescribed in the ocean model in the tropical Pacific, while coupling operates normally over the rest of the world. The surface wind stress is prescribed within the tropical Pacific from 20°S to 20°N, from the Indonesian archipelago to the west coast of Central and South America, with a buffer zone between 15° and 25° latitude to avoid discontinuities, as shown by the black contours in Fig. 1a. The surface turbulent heat fluxes are unchanged in the model, so that the sensible and evaporative fluxes are calculated using the interactive model wind. We prefer this approach to nudging SST anomalies because it yields a conservative ocean-atmosphere heat exchange (Douville et al., 2015).

The coupled model long-term wind stress climatology is applied in the tropical Pacific in W-CLIM. This long-term climatology is estimated as the 1979–2014 average of 6 members from the IPSL-CM5A-LR historical CMIP5 ensemble (1979 to 2005) and 4 members from RCP8.5 CMIP5 ensemble (2006 to 2014), smoothed (Lanczos filter, 30-day cutoff period) to remove noise. As the mixed layer dynamics is parameterized using a turbulent kinetic energy closure scheme (Madec, 2008), the high frequency variability of the wind stress linked to weather noise is important to estimate correctly the mixing within the oceanic mixed layer, as mentioned in Lee et al. (2015). High frequency winds indeed induce

a vertical shear, which in turn generates turbulence and deepens the mixed layer. Therefore, we add randomly scrambled high-frequency (3–30 day band-passed) wind stress anomalies from the ERA-Interim reanalysis (Dee et al., 2011) to the model climatological values. Figure 1b shows the zonal wind stress model climatology averaged over the Niño 4 region (160°E–150W°, 6°S–6°N). Though the trade winds are weaker in the model, the seasonal cycle agrees with ERA-Interim, with maximum values in January, and minima in boreal spring and fall.

The second experiment, referred to as W-FULL, uses the daily wind stress anomalies from ERA-Interim from 1979–2014 added to the IPSL-CM5A-LR model climatology. The daily wind stress anomalies were obtained by subtracting the daily ERA-Interim 1979–2014 climatology.

2.3 Ocean heat content and heat transport

OHC is calculated from:

$$OHC = \int \rho_0 C_p T dz \quad (1)$$

where C_p is the seawater specific heat capacity. T is the potential temperature field and ρ_0 is the seawater density (i.e. 1025 kg m⁻³). z denotes depth. We decompose the OHC variations using a simple heat budget:

$$\Delta OHC = \bar{F} \Delta t + \left(\int \rho_0 C_p \nabla (\overline{\mathbf{u} \cdot T}) dz \right) \Delta t + R \quad (2)$$

where $\mathbf{u}$ is the current, and F , the local surface heat flux anomaly at the ocean surface, positive downward. Δt denotes the length of the time period considered. ΔOHC is the OHC variation over the time period. The overline designates monthly average. The first term on the right hand side is the surface heat uptake, which includes the effect of air/sea heat exchanges. This term is calculated from the heat fluxes on the oceanic grid. The second term is the ocean heat transport divergence and represents the influence of the ocean circulation. This term is

calculated from the monthly vertically-integrated ocean heat transport calculated on-line by the ocean model, including the eddy-induced (bolus) velocity calculated by the Gent and McWilliams (1990) parameterization, but without accounting for the ocean transport due to diffusive lateral mixing (outputs not available). The last term, R , is calculated as a residual and accounts for the contributions from diffusive lateral mixing and changing sea ice heat content. We will show in Section 4.3 that this term is small for the time period we consider.

OHC is an extensive variable and is sensitive to the area of the oceanic grid cell. To avoid any influence of the grid cell area, the maps shown in the following illustrate the OHC per unit area (J m^{-2}), while the OHC integrated over the different basins is in J.

2.4 Observations and reanalysis data

We use the SLP and 2m surface air temperature data from the ERA-Interim reanalysis (Dee et al., 2011) from the European Center for Medium-Range Weather Forecasts (ECMWF). For SST, we mainly use the monthly NOAA Optimal Interpolation SST Analysis, version 2 (OISST) (Reynolds et al., 2002); we also examine HadISST1 (Rayner et al., 2003) after removal of the global warming signal using an optimal filter based on linear inverse modeling from Frankignoul et al. (2017). We take monthly sea surface height (SSH) from the Archiving, Validation and Interpretation of Satellite Oceanographic (AVISO) dataset (Nerem et al., 2010), gridded ocean fields from the Ocean Reanalysis System 4 (ORAS4; Balmaseda et al. 2013), and surface heat fluxes from the TropFlux product (Kumar et al. 2012).

2.5 Climate indices

We use the Niño 3.4 index as a standard metric for ENSO: the SST anomaly over 120° – 170° W and 6° S– 6° N. The east–west thermocline tilt is given by the average 6° S– 6° N

20°C isotherm depth in 120°E–170°W minus that in 150°–80°W. In order to analyse the IPO, we use the SST tripole index (TPI, Henley et al. 2015), defined as the central equatorial Pacific SST (170°E–90°W, 10°S–10°N) minus the average of northwest (140°E–145°W, 25°–45°N) and southwest (150°E–160°W, 15°–50°S) Pacific SST, smoothed using a 1-3-4-3-1 filter (Trenberth et al., 2007). The equatorial Pacific zonal SLP gradient, defined as the area-averaged SLP in the eastern (160°–80°W, 6°S–6°N) minus that in the western equatorial Pacific (80°–160°E, 6°S–6°N) is used as a proxy of the Walker circulation (Vecchi et al. 2006; Kociuba and Power 2015). To evaluate the atmospheric circulation changes in the mid-latitudes we also use the averaged SLP over the North Pacific (as in the North Pacific Index, Trenberth and Hurrell 1994) and the Amundsen Sea, defined over 160°E–140°W, 30°–65°N and 170°E–70°W, 60°–75°S respectively. Lastly, we use the normalized difference between 35°N and 65°N of the zonally averaged sea-level pressure (SLP) in 80°W–30°E as a NAO index (Li et al., 2013).

2.6 Trend analysis and statistical testing

In time series plots, the ensemble mean is displayed, and the ensemble spread is depicted using 95% confidence intervals for the ensemble mean. We will first focus on the 21-year period of strengthening Pacific trade winds (Balmaseda et al., 2013; de Boissésou et al., 2014; England et al., 2014) to investigate if W-FULL successfully describes the associated transition from a positive to negative IPO global pattern. Unless stated otherwise, trends are reported for annual means. The level of statistical significance of simulated trends is evaluated using two-tailed Student *t*-tests, using the values of all members. Correlations between observations and simulations are based on linearly detrended values, and use the ensemble means. The significance level of observed trends and correlations between observations and simulations is calculated using a *t*-test, adjusting the standard error and the

degrees of freedom following a first order autoregressive process (Bretherton et al., 1999; Santer et al., 2000). As the TPI is smoothed using a low-pass filter, we estimate the level of significance of the TPI trend using the adjusted standard error from an ARMA(1,1) model (Foster and Rahmstorf, 2011). Trend and correlation values are given in boldface if they are statistically significant at the 95% confidence level.

3. Climatic impacts of the increasing tropical Pacific wind stress

3.1 Pacific sea surface temperature

The observed wind stress anomaly in the Niño 4 region (Fig. 1c) displays a strengthening trend of $-1.65 \cdot 10^{-2}$ Pa decade⁻¹ over 1992–2012 consistent with previous studies (Balmaseda et al., 2013; de Boissésou et al., 2014; England et al., 2014). It also displays clear ENSO-related interannual variability, to which the east-west tilt in the thermocline (Fig. 1d, **0.98** correlation with ORAS4 reanalysis) and SST anomalies in the central Pacific (Fig. 1e, **0.91** correlation with OISST) respond. W-FULL is able to reproduce observed ENSO SST and thermocline variations in response to the prescribed observed wind stress anomalies. This is consistent with the well-known dominant role of wind anomalies for the thermocline depth at the equator and for the ocean temperature (Sarachik and Cane, 2010), while the different amplitudes might be due to the heat flux feedback, which depends on the coupled model used (Bellenger et al., 2014). At longer timescales, W-FULL reproduces the increasing east-west thermocline tilt and cooling SST trend in the central Pacific associated with strengthening easterlies (see Table 1). Overall, W-FULL reproduces the observed tropical Pacific decadal variability, capturing the transition from the more predominant El-Niños in the 1990s to the more predominant La Niñas in the 2000s.

Conversely, imposing the wind stress climatology in W-CLIM (Figs. 1d and 1e, red line) yields no interannual variability in thermocline tilt. As a result there is no interannual variability in SST, except for the cooling following the Pinatubo (1991) and El Chichón (1982) volcanic eruptions. W-CLIM also reveals a warming trend associated with the imposed evolution of GHG (Table 1), that is difficult to see in W-FULL due to the larger interannual and decadal variations.

W-FULL hence reproduces Pacific ENSO variations and their decadal modulation, including the transition from a warm phase in the 1990s to a cold phase in the 2000s. W-FULL also reproduces the IPO time evolution (Fig. 1f), with the observed transition from the positive to negative IPO around 1998, again with an overestimated amplitude. In the following sections, we will investigate the associated atmospheric changes.

3.2 Surface air temperature

W-CLIM (Fig. 2a) features a broad-scale positive 1992–2012 2m air temperature trend, with larger amplitudes over land and reduced warming in the Southern Ocean and subpolar North Atlantic, consistent with the global warming pattern simulated by most models (e.g. Stouffer et al., 1989). In contrast, W-FULL (Fig. 2b) features a prominent cooling across the tropical Pacific, and enhanced warming near the Kuroshio extension and around 40°S, consistent with the trend toward a negative IPO. Comparison with the ERA-Interim 2m temperature trend (Fig. 2c) indicates that W-FULL largely captures the observed negative IPO-like trend pattern, with an overestimated amplitude and an equatorial cooling that extends too far westward, consistent with the cold tongue bias in that model (Bellenger et al. 2014).

Figure 2d displays annual global mean 2m air temperature time series, with the 1992–2012 trend values in Table 1. The 1992–2012 ensemble mean trend is significantly smaller in

W-FULL than W-CLIM, and much closer to ERA-Interim. Imposing observed wind stress anomalies in the Pacific leads to a more realistic rate of global warming, as in previous similar experiments (England et al. 2014; Watanabe et al. 2014; Douville et al., 2015; Thoma et al. 2015). The W-FULL surface temperature trend is still greater than observations, reflecting the large IPSL-CM5A-LR climate sensitivity (Forster et al. 2013). Overall, the tropical Pacific changes yield a global temperature response that is broadly consistent with the observed changes from 1992 to 2012. Next, we explore the corresponding atmospheric circulation changes.

3.3 Large-scale atmospheric circulation changes

Figure 3 shows the 1992–2012 SLP trend to illustrate the large-scale atmospheric circulation changes during that period. W-CLIM SLP trends (Fig. 3a) are weak, with lower SLP in the tropics as the SST warms and an increase of SLP at 40° over the northern and southern mid-latitudes as the Hadley circulation expands poleward under climate change (Lu et al. 2008). There is also a SLP decrease over the Arctic induced by the large surface warming in that region (see Fig. 2a). W-FULL (Fig. 3b) displays much larger SLP trends than W-CLIM, with a SLP increase in the eastern tropical Pacific and decrease in the western Pacific and Indian Oceans, consistent with an enhanced Walker circulation. These changes are broadly consistent with ERA-Interim (Fig. 3c), though the W-FULL positive SLP anomaly in the eastern equatorial is overestimated and extends too far westward. The equatorial Pacific zonal SLP gradient, Δ SLP (Fig. 3d), illustrates that the Walker circulation in W-FULL (Fig. 3d) reproduces the observed variations of the Pacific easterlies (correlation of **0.81**) and its 1992–2012 strengthening trend, albeit with larger amplitude. Therefore, constraining the wind stress yields realistic SST variations in the equatorial Pacific (Fig. 1e) that in turn constrain the equatorial Pacific Walker circulation.

The W-FULL extratropical Pacific SLP changes (Fig. 3b) display large anticyclonic anomalies over the Aleutian Low and the southern Pacific Ocean at 40°S, while cyclonic anomalies are simulated over the Amundsen Sea. Such SLP anomalies are also found in ERA-Interim (Fig. 3c; also see the North Pacific and Amundsen Sea Low indices in Table 1). These SLP changes are consistent with those expected from ENSO teleconnections. El Niños indeed yield a deeper Aleutian Low, negative SLP anomalies at ~30-50°S in the Pacific and south of Australia (e.g. Sterl et al. 2007) and a weaker than usual Amundsen low (e.g. Fogt et al. 2012; Turner et al. 2013; Clem and Renwick, 2015; Meehl et al., 2016), with roughly opposite signals during La Niñas. The transition from a positive (more Niños) to a negative (more Niñas) IPO phase hence yields the signals in Figs. 3b and 3c. It also explains the significant trends that are similar for observations and W-FULL for the averaged SLP over the North Pacific and Amundsen Sea (Table 1).

Over the Atlantic, W-FULL does not simulate the observed extratropical SLP changes, as illustrated by the NAO index trend in Table 1. The 1992–2012 W-FULL trend shows a SLP dipole over the North Atlantic centered over the western part of the basin that projects onto a positive NAO pattern, which is consistent with the response to La Niña conditions (see Gastineau et al., 2013; their Fig. 15). The observations however display an opposite trend, with a weak and insignificant trend toward a negative NAO-like pattern. The ensemble mean displayed for the model filters out internal atmospheric variability and reveals more clearly the response from the changing IPO, while the internal variability over the Atlantic Ocean dominates in observations.

The above analysis suggests that constraining the ENSO phase in the tropical Pacific reproduces most of the observed large-scale atmospheric circulation trends in the tropical Indo-Pacific and extratropical Pacific. Those changes are consistent with the expected ENSO

teleconnections when going from a positive to a negative IPO phase. In the following section, we examine the OHC changes associated with these large-scale atmospheric changes.

4. Evaluation of the ocean heat content changes

In this section, we will first illustrate the global OHC changes attributable to the IPO over 1979-2014. We will then focus on the 1998-2012 negative IPO phase, and first show that the global SST anomalies and OHC changes during this period are consistent with observed ones. We will then illustrate the geographical distribution of the OHC changes attributable to the 1998-2012 negative IPO phase, and their origin (surface heat uptake or oceanic transport). We will end by detailing the mechanisms responsible for the heat uptake and redistribution by oceanic transport.

4.1 Global energy balance and ocean heat content evolution

We first diagnose and quantify the global OHC changes and their causes. W-FULL and W-CLIM both display a stable TOA radiative imbalances from 1995 onward of about 1.2 W m^{-2} (not shown), when accounting from energy nonconservation in the atmosphere by removing the value from the TAO imbalance calculated over the whole preindustrial control simulation, as in Hobbs et al. (2016). This value is larger than the present-day observed estimate, which ranges between 0.5 and 1 W m^{-2} (Loeb et al. 2012; Allan et al. 2014; Johnson et al., 2016), and reflects the large climate sensitivity of IPSL-CM5A-LR (Brient and Bony 2013; Andrews et al. 2012). The monthly globally-integrated OHC increases throughout 1995-2014 in both W-FULL and W-CLIM due to this TOA imbalance (Fig. 4a).

The globally-averaged difference between the two simulations, which filters out the influence of the common radiative forcing in each simulation, only retains the changes associated with the IPO. Assuming that the climate variations simulated in W-FULL are a

linear superposition of the global warming simulated in W-CLIM and the IPO-related changes, taking the difference W-FULL minus W-CLIM should also remove the changes induced by the large climate sensitivity of the model used. The average monthly heat flux into the ocean (positive downward) in Fig. 4b displays sustained positive anomalies after 1998, and variations from positive to negative anomalies before 1998. This is related to a larger TOA radiative imbalance of 0.11 (0.02) W m^{-2} in W-FULL during 1998-2012 (1979-1997) linked to the more frequent La Niña after 1998, as found in observations (Allan et al., 2014). As a result, the accumulated heat into the global ocean (orange curve on Fig. 4c) displays a steady rise from 1998 onward. The global OHC variations (black curve on Fig. 4c) closely track those of the integrated heat fluxes, which confirms that the heat budget is closed at the global scale. As 1998-2012 experiences most of the IPO-related OHC increase, hereafter we will focus on this period to track the OHC changes associated with the IPO negative phase. We will discuss the OHC changes occurring before 1998 in section 5.

4.2 SST and OHC changes during the period 1998–2012

Since we focus on the 1998-2012 IPO negative phase, we investigate the SST anomalies and OHC changes during this period. The 1998–2012 W-FULL minus W-CLIM SST anomalies (Fig. 5a) illustrate that this period corresponds to a typical negative IPO phase, with a pattern that agrees well with observational SST anomalies after removal of the global warming signal (Fig. 5b, see section 2.4 for details). The model and observations both display an anomalously cold tropical Pacific. The trademark horseshoe pattern associated with La Niña, with warm SST anomalies in the Kuroshio extension region and subtropical Southern Pacific, broadly agree in the model and observations. The too strong and westward cold anomalies in the tropical Pacific reflect the typical cold tongue bias of the model.

The OHC changes between early 1998 and late 2012 (hereafter, ΔOHC) in W-CLIM (Fig. 6a) display a relatively uniform increase, with heat content mostly increasing in the top 300m (not shown). OHC actually diminishes between 10°N and 20°N in the Pacific, presumably in response to a northward shift of the ITCZ. It also decreases in the eastern North Atlantic as the Atlantic Meridional Overturning circulation (AMOC) slows down (Drijfhout et al., 2012). ΔOHC in W-CLIM however bears little resemblance to the ORAS4 estimate (Fig. 6c). On the other hand, the W-FULL ΔOHC (Fig. 6b) displays a much better agreement with observations (Fig. 6c). In particular, OHC diminishes over the eastern tropical Pacific and increases over the western tropical Pacific, consistent with the anomalously strong easterlies during that period (e.g. England et al. 2014). While this is expected due to the prescribed wind stress anomalies in the tropical Pacific in W-FULL, W-FULL also better reproduces OHC changes in other regions. For instance, both W-FULL and ORAS4 display increasing OHC over the Kuroshio extension, the southwestern Pacific, and (to a lesser extent) the eastern tropical Indian Ocean and Gulf Stream extension regions. The better agreement of W-FULL than W-CLIM with the observed 1998-2012 OHC changes shows that our methodology constrains reasonably the OHC changes associated with the IPO, at least over the Indo-Pacific.

Hereafter, the 1998-2012 W-FULL minus W-CLIM OHC changes, i.e. the changes attributable to the negative IPO phase, are called $\Delta\text{OHC}_{\text{-IPO}}$. The negative IPO is associated with an OHC increase in the western tropical Pacific, and around 40° latitude in both hemispheres (Fig. 7a). On the other hand, there is a clear OHC decrease in the eastern tropical Pacific and at 60°S in the Pacific sector. The vertical distribution of $\Delta\text{OHC}_{\text{-IPO}}$ reveals that the most pronounced OHC changes are located in the upper 300m in the tropics (Fig. 7b). The intermediate ocean (300m-1200m) also contributes to $\Delta\text{OHC}_{\text{-IPO}}$ at mid and high-latitudes (Fig. 7c), with anomalies of the same sign as in the upper 300m. The OHC changes are

weaker below 1200m (Fig. 7d), and mostly located in the Southern Ocean, in contrast to Meehl et al. (2011) who pointed to larger heat content variations in the deep ocean during hiatus periods.

4.3 Regional heat budget for the IPO-induced OHC changes

We now investigate the contributions from surface fluxes, oceanic transport, and the residue, as in Eq. (2), for $\Delta\text{OHC}_{\text{IPO}}$. Figure 8 shows this regionally-integrated budget for the tropical Pacific, North Pacific (north of 31°N), Atlantic (including the Arctic), Southern (south of 31°S), and Indian Ocean basins (basin boundaries are depicted in Figs 7 and 8e). The residual in Eq. (2) (Fig. 8d) represents the contributions of parameterized diffusive lateral mixing and changing sea ice heat content. This residual is in general close to zero or much weaker than the dominant terms in the budget, so that OHC changes can be accounted for by the sum of the surface heat uptake and ocean heat flux transport convergence.

We find that the anomalous surface heat uptake during the 1998-2012 IPO negative phase (Fig. 8c) exclusively occurs in the tropical Pacific Ocean ($65 \pm 4 \times 10^{21}$ J), with other basins experiencing either negligible IPO-induced surface fluxes (Atlantic, Southern Ocean) or more moderate heat losses to the Atmosphere (Indian Ocean, North Pacific). Despite this major heat input from the atmosphere, the tropical Pacific OHC hardly changes over the 1998-2012 period (Fig. 8a). The tropical Pacific surface heat uptake is indeed entirely transferred to other basins. The ratio of the oceanic transports from the tropical Pacific (Fig. 8b and arrows in Fig. 8e) versus the surface heat uptake (Figs. 8c and 8e) shows that 53% of the heat received at the surface of the tropical Pacific is transferred to the North Pacific, 26% to the Indian Ocean, and 21% to the South Pacific. Overall, 65% of the global $\Delta\text{OHC}_{\text{IPO}}$ occurs in the Southern Ocean ($26 \pm 11 \times 10^{21}$ J, as illustrated in Fig. 8a), where the surface heat loss is negligible (Fig. 8c). The ratio of oceanic heat transport (arrows in Fig. 8e) versus the

southern ocean $\Delta\text{OHC}_{\text{IPO}}$ illustrates that 49% of the OHC increase is due to the heat transport across 31°S from the tropical Pacific, 32% is due to the heat transport from the Indian ocean, and 19% is due to the transport from the Atlantic Oceans. Despite receiving the largest heat transport from the tropical Pacific, the North Pacific experiences a smaller OHC increase than the Southern Ocean (40% of the global OHC increase), due to heat losses to the atmosphere ($13 \pm 3 \times 10^{21}$ J, see Fig. 8c and 8e). In the Indian Ocean, about half of the input of heat from the tropical Pacific is evacuated southward to the Southern Ocean via the Indonesian throughflow (Fig. 8e), with a relatively large heat quantity ($17 \pm 2 \times 10^{21}$ J) released back to the atmosphere. As a result, the Indian Ocean experiences little $\Delta\text{OHC}_{\text{IPO}}$. In the Atlantic, the IPO induces little changes in either atmospheric heat uptake or transport, with no significant $\Delta\text{OHC}_{\text{IPO}}$ as a result, as will be discussed in section 5.

As expected from the surface temperature changes, the integrated $\Delta\text{OHC}_{\text{IPO}}$ over the different basins diminishes in the upper 50m (Fig. 9a), except for the Southern Ocean. The dominant OHC increase occurs within 50-300m in the tropical Pacific (Fig. 9b). However, the OHC decreases below 300m in the tropical Pacific, producing a negligible OHC anomaly over the total water column. Conversely, the Southern Ocean and North Pacific OHC increase at all depths below 50m. This suggests that there is a global heat redistribution within the ocean, as previously found (Meehl et al., 2011; Chen and Tung, 2014; Liu et al. 2016a; Liu et al. 2016b), but here our experiments find a larger role of the Southern Ocean, with an OHC increase that remains important below 1200m.

Figure 10 shows the spatial distributions of the heat budget terms. $\Delta\text{OHC}_{\text{IPO}}$ (Fig. 10a) is repeated from Fig. 7a, while Fig. 10b and 10c are the contributions from surface fluxes and ocean heat transport. Table 2 quantifies these contributions to the global and regional $\Delta\text{OHC}_{\text{IPO}}$ using pattern regression: 68% of the global OHC changes attributable to the IPO are driven by the oceanic transport. The tropical Pacific is the only region with a dominant

effect of the surface heat uptake (72%). This surface heat uptake occurs mostly over the western tropical Pacific (Fig. 10b), due to cold IPO-induced SST anomalies there (Fig. 5a), which favor a downward surface net heat flux anomaly. Table 3 further illustrates that the enhanced downward heat flux is mostly due to the reduced upward surface latent heat flux and the increased downward surface shortwave radiation. The increased downward surface shortwave radiation is weaker in clear-sky conditions, and is consistent with the reduced total cloud cover. The Pacific heat uptake during negative IPO phases is expected to occur in the central and eastern Pacific (e.g. Kosaka and Xie, 2013). The dominant heat uptake in the western Pacific in our simulations is probably related to the cold tongue bias, which induces IPO-related SST anomalies which are too strong and shifted westward (Fig. 5).

In the North Pacific, Indian, and Atlantic Oceans, ocean dynamics are the dominant contributor to $\Delta \text{OHC}_{\text{IPO}}$, with surface fluxes to the atmosphere acting as a negative feedback (Fig. 10, Table 2). In particular, ocean heat transport changes explain the two bands of $\Delta \text{OHC}_{\text{IPO}}$ increase around 40° latitude in both hemispheres. This anomalous oceanic heat transport convergence is mainly due to the eulerian transport, with only a small contribution from the eddy bolus transport (not shown). In the following section, we will focus on the mechanisms that explain the main IPO-induced ocean heat flux convergence seen in Figs. 8 and 10c.

4.4 Mechanisms of IPO-induced ocean heat transport

The W-FULL minus W-CLIM SLP difference has a characteristic pattern of La Niña or negative IPO teleconnections during 1998-2012 (Fig. 11a), with positive SLP anomalies in the Aleutian Low region and around 50°S in the Pacific sector of the Southern Ocean (e.g. Trenberth and Caron, 2000) and negative SLP anomalies over the Amundsen Sea (Turner et al. 2013). As a result, the W-FULL minus W-CLIM wind stresses (Figs. 11a and 12a) display

positive anomalies around 55°S and 55°N, which can be interpreted as the poleward shift of westerlies characteristic of La Niña conditions (e.g. Lu et al. 2008).

To investigate the associated Ekman response, we compute the Ekman transport and pumping (Gill, 1982). The Ekman response to the anomalous wind stress indeed reveals increased poleward transport at 31°N and 31°S at the edges of the domains used in Fig. 8 (see Fig 11b). The horizontal Ekman meridional transport converges around 40° latitude in both hemispheres (Figs 11b-c). The negative SLP anomaly over the Amundsen Sea, typical of negative IPO conditions (Fogt et al 2012; Clem and Renwick, 2015; Meehl et al. 2016), also yields Ekman flow divergence at 60°S in the Pacific basin. $\Delta \text{OHC}_{\text{IPO}}$ (Fig. 11b) is consistent with the Ekman response, with downward Ekman pumping associated with positive $\Delta \text{OHC}_{\text{IPO}}$ especially at 40°S, between South of Australia and South America and at 40°N in the North Pacific, and decreased OHC associated with upward Ekman pumping in the Amundsen Sea.

The influence of these processes is clearly seen in the anomalous W-FULL zonal-mean temperature change (Fig. 12b) that is dominated by the Pacific Ocean anomalies. The positive temperature anomalies at 40°S and 40°N penetrate downward into the ocean where the mean slope of the isopycnals is large (not shown), which enhances the heat penetration into the deeper layers through the mean flow. We also note an intensification of the shallow subtropical cells between 20°S and 20°N in the upper 400m of the ocean (see contours in Fig. 12b) as described in England et al. (2014), but these only lead to shallow temperature anomalies. Such anomalies are associated with the large OHC increase found at 50-300m in the tropical Pacific (see Fig. 9). A positive anomaly of the global meridional streamfunction is also found between 65°S and 50°S, as the Deacon cell intensifies and shifts southwards.

Llovel and Terray (2016) found a similar process when analysing Argo data and atmospheric reanalysis over the 2005-2014 period, but they could not attribute these observed

changes specifically to either climate change or natural variability. Using our experimental setup, we can quantify those two influences. Figure 12a shows the zonal-mean wind stress anomalies associated with the negative IPO phase (black line) and the anomalies simulated in W-CLIM (black dashed line). In W-CLIM, increasing greenhouse gases and decreasing stratospheric ozone yield positive wind stress anomalies around 50°S (Shindell and Schmidt, 2004). The IPO-related (i.e. W-FULL minus W-CLIM) anomalies are similar, but they are, however, larger and 10° further south. Similarly, we evaluate that the Southern Ocean heat storage occurs at a rate of $61.9 \pm 5.6 \text{ } 10^{21} \text{ J decade}^{-1}$ in W-FULL over the negative IPO period, while it occurs at a rate of $44.5 \pm 5.1 \text{ } 10^{21} \text{ J decade}^{-1}$ over the same period in W-CLIM. For comparison, the average rate for the Southern Ocean heat storage is $33.1 \pm 2.5 \text{ } (39.2 \pm 2.9) \text{ } 10^{21} \text{ J decade}^{-1}$ over 1979-2014 in W-CLIM (W-FULL, respectively). Therefore, external forcings have caused an accelerated Southern Ocean OHC increase during the negative IPO period (increased rate of $23.1 \pm 4.3 \%$), but we find that these changes are smaller than those induced by tropical SSTs through atmospheric teleconnections (increase of $48.1 \pm 8.6 \%$).

Overall, our results therefore suggest that, during negative IPO phases, ENSO teleconnections to the extratropical Pacific drive increased poleward Ekman heat transport through 31°S and N. The associated Ekman pumping induces heat subduction at 40° S and N in the intermediate water formation regions of both hemispheres. The heat subducted in the Southern Ocean is larger than that in the northern Pacific, reflecting the larger meridional temperature gradient and the larger oceanic area in the Southern Hemisphere.

5. Summary and discussion

5.1. Summary

In this study, we examine the OHC changes over the 1998-2012 period, during which global mean surface temperature displayed a reduced warming rate, which has been attributed to oceanic heat redistribution and enhanced heat uptake in the Pacific (Meehl et al., 2011; Kosaka and Xie, 2013; Fyfe and Gillett, 2014; Clement and DiNezio, 2014; Trenberth, 2015; Dai et al., 2015). To test this hypothesis, we prescribe the 1979–2014 tropical Pacific (20°N–20°S) surface wind stress in two ensembles with identical GHG, aerosol, and ozone forcing: one with climatological wind stress (W-CLIM), and the other with wind stress anomalies from ERA-Interim (W-FULL). As already explored in previous studies (England et al. 2014; Watanabe et al. 2014; Douville et al. 2015; Delworth et al. 2015; Thoma et al. 2015; Saenko et al. 2016), we find that this strategy is successful in constraining ENSO variations in the tropical Pacific and their decadal modulation, with an El Niño residual signature during 1979-1997 and a La Niña signature during 1998-2012. Atmospheric teleconnections also produce global air temperature and SLP patterns characteristic of a transition from a positive to a negative IPO phase. As a result, W-FULL also simulates the observed reduced rate in global surface warming during 1992–2012 (Gleisner et al. 2015).

The SST and OHC anomaly patterns during the negative IPO phase (1998-2012) are similar to the observed ones, which allows us to use our simulations to investigate OHC changes during this period. The W-CLIM minus W-FULL ensemble mean indeed isolates the effects of enhanced tropical Pacific wind stresses associated with the negative IPO phase. The IPO negative phase leads to a global heat redistribution, in accordance with previous studies (Meehl et al. 2011, 2013; England et al. 2014). Almost all the anomalous oceanic heat uptake occurs in the tropical Pacific. Part of this heat is exported to the Indian Ocean through the enhanced throughflow, but yields no OHC increase due to heat release to the atmosphere and enhanced southward transport across 31°S in this basin. The remainder of the tropical Pacific heat uptake is transported polewards within the ocean. The largest OHC increase

occurs at all depths in the Southern and North Pacific Oceans. The OHC increase is weaker in the North Pacific as much of the heat is released back to the atmosphere in this region. There is a strong heat convergence around 40°N and 40°S due to the teleconnections associated with the cold tropical Pacific SST, producing enhanced poleward Ekman transport across 31°N and 31°S, and Ekman convergence at 40° latitude in both hemispheres, as found by Oke and England (2004). As a result, the OHC changes are strongly related to the Ekman pumping patterns associated with IPO teleconnections. Recent simulations have also shown a southward transport of ocean heat from the tropical Pacific during the 1998–2012 warming slowdown (Oka and Watanabe 2017). Our experimental protocol allows us to attribute most of this wind-driven OHC change to IPO-related teleconnections.

5.2. Discussion

It is important to acknowledge some limitations in our modeling. First, as mentioned previously, IPSL-CM5A-LR, like many other models, displays SST anomalies in the tropical Pacific that extend too far westward, which is linked to the equatorial Pacific cold tongue bias (Li and Xie, 2014). This probably explains why the heat uptake occurs predominantly in the western Pacific in our study, whereas it is generally expected to occur mostly in the central/eastern Pacific (e.g. Tokinaga and Xie, 2012).

We have seen in our results that the W-FULL minus W-CLIM anomalous heat uptake occurred almost exclusively during the 1998-2012 negative IPO period, with little sign of anomalous heat uptake during the 1979-1997 positive IPO period. Why is there such an asymmetry? As expected, the negative IPO is characterized by a negative SST anomaly and the positive IPO by a positive SST anomaly in the tropical Pacific (Table 3). As a result, the longwave, sensible and latent surface heat fluxes have opposite signs during these two periods. However, the shortwave surface heat flux displays positive anomalies during both

phases. This can be related to the observed non-linearity of the shortwave flux associated with ENSO-related SST anomalies (Fig. 13b). During La Niñas, observations indicate a decrease in the incoming shortwave radiation, consistent with a stronger atmospheric inversion at the top of the planetary boundary layer, and increased low-level cloud coverage (Fig. 13b; Stephens, 2005). During El Niños, positive SST anomalies lead to atmospheric convection and increasing high and mid-level cloud coverage, which also reduce the surface shortwave radiation (Fig. 13b). Bellenger et al. (2014) have shown that most CMIP5 models have trouble reproducing this nonlinear behavior. This is also the case in the IPSL-CM5A model we use here, even though using prescribed wind stress improves this aspect of the model (Fig. 13a). The shortwave dependence on SST anomalies is reasonable during La Niñas, but the effect of warm SSTs onto shortwave surface heat flux is largely underestimated in the model. Due to this issue, we chose to only focus on the 1998-2012 negative IPO phase in the current study. Nevertheless, the observed asymmetry of the surface shortwave heat flux to SST suggests that there may also be some asymmetry in heat uptake between IPO positive and negative phases in the real world.

The model used in this study (IPSL-CM5A-LR) also displays a large climate sensitivity, with a rather strong TOA radiative imbalance compared to observations (Loeb et al., 2012; Allan et al., 2014, Johnson et al., 2016) or other models such as CCSM4 (Meehl et al., 2011) or CESM (Liu et al., 2016a). It is likely that the cloud-related processes linked to this large sensitivity are also involved in the global TOA radiative imbalance and global OHC variations associated with the IPO. More work is needed to verify if the OHC changes related to the IPO found in this study are robust. This can be done through other pacemaker ensemble experiments with other models or analyses of the OHC modulation by the IPO phase in the CMIP5 preindustrial control runs.

Outside the Pacific, our simulations can also be used to reveal the influence of the Pacific Ocean variability onto the Indian and Atlantic Oceans. Our results indicate a large increase of the heat transport to the Indian Ocean during the negative IPO phase. This agrees with several recent studies using both modeling and observations (Lee et al. 2015; Nieves et al. 2015; Liu et al. 2016a). Despite this large heat transfer to the Indian Ocean via the Indonesian throughflow, there is hardly any IPO-induced change in the Indian Ocean OHC over the 1998-2012 period in our simulations, due to increased southward heat transport across 31°S and heat released to the atmosphere through surface fluxes. Is this realistic? To investigate this, we compare the sea-level changes attributable to the IPO (W-FULL minus W-CLIM) to observed sea level over 1998-2012 in Fig. 14. The globally-averaged sea-level signal has been subtracted to remove potential effects of changes in the cryosphere and land storage (Greatbatch, 1994). The model underestimates the observed sea level rise in the eastern and southeastern Indian Ocean. This region is directly connected to the western Pacific via the Indonesian throughflow at decadal and multidecadal timescales (e.g. Feng et al. 2004). The sea level anomalies at the equator in the western Pacific (where the throughflow entrance is located) are also strongly underestimated relative to observations, probably due to the cold tongue bias in the Pacific and associated westward shift of equatorial easterly anomalies. There is thus probably an underestimation of the heat transfer from the Pacific to the Indian Ocean. It is more difficult to gauge whether the southward ocean transport across 31°S and the strong surface heat losses to the atmosphere in the model (Fig. 8) are realistic or not.

The Atlantic Ocean is also influenced by the Pacific variability in W-FULL. The 1998–2012 Atlantic zonal-mean changes linked to the cold IPO phase are shown in Fig. 15. The IPO-related signal in our simulation is weak in the Northern Atlantic. In contrast, Chen and Tung (2014) find a large temperature increase over the whole north Atlantic from

observations, but they do not separate the IPO signal from that associated with anthropogenic climate change. During the negative IPO, the Atlantic OHC increase is largely concentrated in the first 1000m at 40°N, while OHC decreases between 50°N and 60°N, and at 15°N. These changes are consistent with the SST anomalies (Fig. 5a) showing the typical SST tripole driven by the positive NAO-like SLP anomalies occurring over the North Atlantic sector (Figs. 3b and 11a). W-FULL also simulates a 0.6 Sv weaker Atlantic meridional overturning circulation relative to W-CLIM (Fig. 15, contours), which is consistent with the fast Ekman-driven barotropic oceanic response to a positive NAO (Eden and Willebrand, 2001). The positive NAO-like anomalies in W-FULL are consistent with those arising from La Niña (or negative IPO) teleconnections (Gastineau et al., 2013). In observations, the North Atlantic SLP anomalies are not significant during 1998-2012 (see NAO trends in Table 1) and the atmospheric internal variability dominates the decadal variability. Our results therefore agree with a dominant role of the global heat redistribution induced by the negative IPO, with little influence of the Atlantic Ocean, as found by Liu et al. (2016a, 2016b). Nevertheless, the Atlantic variability might have influenced the Pacific Ocean through its impacts onto the Walker circulation, as found by Li et al. (2016) or Ruprich-Robert et al. (2017), but this role of the Atlantic Ocean cannot be addressed with our simulations.

Lastly, our working hypothesis is that external forcings have no impact on the Pacific trade wind intensification during 1998-2012, so that we attribute the differences between the results of W-FULL and W-CLIM entirely to IPO variability. However, some studies have suggested that external forcing may have lead to a significant part of the observed trade wind strengthening (Luo et al. 2012; Han et al. 2014; Kohyama et al., 2016; Dong and McPhaden 2017), in particular due to accelerated Indian Ocean warming. This is in contradiction to other studies that indicate that anthropogenic climate change should drive a weakening of the Walker circulation (e.g. Held and Soden, 2005; Kociuba and Power, 2015). Other studies

point to the Atlantic (Chikamoto et al. 2012; McGregor et al. 2014; Li et al. 2015; Kucharski et al. 2016) or to Pacific aerosols (Allen et al., 2015; Takahashi and Watanabe 2016; Smith et al., 2016) as the ultimate cause of the trade wind strengthening during recent decades. More research is hence probably needed to reach a consensus on the main cause of the trade wind strengthening during recent decades.

Despite these caveats, our experiments provide a robust assessment of the links between the cold IPO phase (be it purely occurring as internal variability or externally forced), the global surface temperature warming slowdown, and the large OHC increase in the Southern Ocean during the first decade of the 21st century. Our results point towards the role of tropical Pacific decadal variability, acting through an atmospheric stationary wave response, to change surface winds and Ekman transport and store heat in the Southern Ocean and North Pacific. We suggest that further modeling simulations would help clarify the link between the IPO and heat content increase in the Southern Ocean, to assess the model dependence of the processes, and to investigate if such links remain at eddy-permitting ocean resolution (Treguier et al., 2010).

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Table 1: 1992–2012 trends of different climate indices associated with the IPO transition from the positive to the negative phase. The 95% confidence intervals are also given. Note that the confidence intervals are calculated differently for the models and observations. See Section 2.5 and 2.6 for details.

	Observation (ERA-Interim/OISST/ORAS4)	W-FULL	W-CLIM
Niño 3.4 SST	$-0.29 \pm 0.47 \text{ }^{\circ}\text{C dec}^{-1}$	$-0.78 \pm 0.02 \text{ }^{\circ}\text{C dec}^{-1}$	$0.48 \pm 0.02 \text{ }^{\circ}\text{C dec}^{-1}$
W–E thermocline tilt	$17.7 \pm 12.0 \text{ m dec}^{-1}$	$21.5 \pm 0.1 \text{ m dec}^{-1}$	$-0.7 \pm 0.1 \text{ m dec}^{-1}$
IPO TPI SST	$-0.50 \pm 0.4 \text{ }^{\circ}\text{C dec}^{-1}$	$-1.0 \pm 0.04 \text{ }^{\circ}\text{C dec}^{-1}$	$0.08 \pm 0.04 \text{ }^{\circ}\text{C dec}^{-1}$
Global mean T2m	$0.17 \pm 0.08 \text{ }^{\circ}\text{C dec}^{-1}$	$0.26 \pm 0.01 \text{ }^{\circ}\text{C dec}^{-1}$	$0.47 \pm 0.02 \text{ }^{\circ}\text{C dec}^{-1}$
Pacific Δ SLP	$0.61 \pm 0.59 \text{ hPa dec}^{-1}$	$0.82 \pm 0.03 \text{ hPa dec}^{-1}$	$-0.06 \pm 0.03 \text{ hPa dec}^{-1}$
North Pacific SLP	$0.78 \pm 0.34 \text{ hPa dec}^{-1}$	$1.45 \pm 0.16 \text{ hPa dec}^{-1}$	$-0.10 \pm 0.20 \text{ hPa dec}^{-1}$
Amundsen Sea SLP	$-1.08 \pm 1.11 \text{ hPa dec}^{-1}$	$-1.34 \pm 0.53 \text{ hPa dec}^{-1}$	$0.16 \pm 0.37 \text{ hPa dec}^{-1}$
NAO	$-1.10 \pm 1.20 \text{ dec}^{-1}$	$0.98 \pm 0.14 \text{ dec}^{-1}$	$0.49 \pm 0.26 \text{ dec}^{-1}$

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Table 2. Pattern regression (unitless) of the ocean surface heat uptake (first term in the r.h.s of Eq. (2)) and ocean heat flux convergence (second term in the r.h.s of Eq. (2)) during the 1998-2012 period onto $\Delta\text{OHC}_{\text{IPO}}$ (l.h.s. of Eq. (2)). The two pattern regression coefficients sum to 1.0 by construction, and quantify the relative contributions of surface-flux and circulation-induced OHC changes to the total.

	Global	PacT	PacN	SO	Ind	Atl
Regression heat uptake onto OHC	0.32	0.72	-0.74	0.16	-0.64	-0.29
Regression ocean heat flux convergence onto OHC	0.68	0.28	1.74	0.84	1.64	1.29

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Table 3. W-FULL minus W-CLIM ensemble mean differences of SST (in °C), surface heat fluxes (positive downward, in W m^{-2}) and total cloud cover (TCC) over the entire tropical Pacific (see Fig. 8e for basin boundaries), for the 1998–2012 (negative IPO) and 1979–1997 (positive IPO) periods. The 1979–2014 mean value for W-FULL is also given as a reference. Total HF designates the total heat flux. SW_{net} CS is the clear-sky surface shortwave flux.

	W-FULL minus W-CLIM		W-FULL
	1998-2012 (IPO-)	1979-1997 (IPO+)	1979-2014
SST (in °C)	-0.12	0.10	25.9
LW_{net} Surf (W m^{-2})	-0.12	0.37	-68.1
SW_{net} Surf (W m^{-2})	0.61	0.26	230.2
Sensible (W m^{-2})	-0.10	0.05	-17.2
Latent (W m^{-2})	0.75	-0.65	-130.0
Total HF (W m^{-2})	1.14	0.03	14.9
SW_{net} CS Surf (W m^{-2})	0.17	-0.12	283.5
TCC (%)	-0.38	-0.02	53.4

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1103 **Figure 1:** (a) Mean 1979–2014 IPSL-CM5A-LR surface wind stress climatology (vectors and
1104 shading), in Pa. Colors indicate the wind stress magnitude. The black contours (contour
1105 intervals 0.3, 0.6, and 0.9) delineate the mask used to prescribe the wind stress in the tropical
1106 Pacific. The red box indicates the Niño 4 region. (b) Wind stress climatology over Niño 4, in
1107 Pa, in IPSL-CM5A-LR historical (black) and ERA-Interim (blue). (c) Monthly Niño 4 wind
1108 stress anomalies from ERA-Interim applied in the W-FULL experiment, in Pa. (d) Monthly
1109 Pacific west–east thermocline (20°C isotherm depth) tilt anomalies. (e) Monthly Niño 3.4
1110 SST anomalies. (f) IPO low-pass filtered SST tripole index (TPI) anomaly. Thick solid lines
1111 in (d–f) indicate ensemble means (black for W-FULL, red for W-CLIM) and observations
1112 (blue). Shading in (d–f) illustrates the ensemble spread, with the 95% confidence intervals.
1113 The dashed lines in (c–f) show the 1992–2012 annual mean trends, colored respectively as the
1114 data and simulations. The vertical dashed lines mark 1 January 1998 and 1 January 2013 (i.e.
1115 the negative IPO period).

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Figure 2. 1992-2012 2m temperature trend, in °C per decade for (a) W-CLIM, (b) W-FULL, and (c) ERA-Interim. Stippling denotes regions where the trends are significant at the 95% confidence level. (d) Globally-averaged annual mean 2m temperature, in °C, for W-CLIM (red), W-FULL (black), and ERA-Interim (blue); note that the simulations are offset by 1.2°C to correct for the mean temperature bias (right axis). Thick solid lines indicate the ensemble means; shading illustrates the ensemble spread, with the 95% confidence intervals. Dashed lines show the 1992–2012 linear trends. The vertical dashed lines mark 1 January 1998 and 1 January 2013.

Figure 3. 1992–2012 annual mean SLP trends, in Pa per decade for (a) W-CLIM, (b) W-FULL, and (c) ERA-Interim. Stippling denotes regions that are significant at the 95% confidence level. Boxes in (c) delineate the regions used to compute the average SLP in the western and eastern equatorial Pacific (used in Δ SLP), North Pacific, and Amundsen Sea in Table 1. (d) Monthly Pacific Δ SLP anomalies, in hPa. Thick solid lines in (d) indicate the ensemble means; shading illustrates the ensemble spread, with the 95% confidence intervals; dashed lines show the 1992–2012 trends. The vertical dashed lines mark 1 January 1998 and 1 January 2013.

Figure 4. (a) Global W-FULL (black) and W-CLIM (red) ensemble mean OHC, in 10^{21} J. (b) Globally-averaged W-FULL minus W-CLIM ensemble mean deseasonalized monthly surface heat flux into the ocean, in W m^{-2} (positive downward). The thin black line represents monthly values, while the thick blue line is smoothed using a 12-month running mean. (c) Global W-FULL minus W-CLIM ensemble mean OHC (black), and the time-integrated net surface heat flux since 1 January 1979 (orange), in 10^{21} J. In (a) and (c) the time series are smoothed using a 12-month running mean. The shading in (b) and (c) indicates the 95% confidence intervals. The vertical dashed lines mark 1 January 1998 and 1 January 2013.

Figure 5. 1998–2012 average SST anomalies (in °C), from: **(a)** W-FULL minus W-CLIM; with shading indicating regions statistically significant at the 95% confidence level; **(b)** HadISST observations after removal of the global warming signal using an optimal linear inverse modeling filter.

Figure 6. OHC change over the 1998-2012 period, ΔOHC , shown here in 10^9 J m^{-2} for the entire ocean column, in (a) W-CRIM, (b) W-FULL and (c) ORAS4.

Figure 7. Vertical distribution of the W-FULL minus W-CLIM OHC change over the 1998-2012 period, $\Delta\text{OHC}_{\text{IPO}}$, in 10^9 J m^{-2} : **(a)** entire ocean column, **(b)** 0-300m, **(c)** 300-1200m, and **(d)** 1200m-6000m. The colors are plotted where significant above the 95% confidence level. The boundaries of the different basins used in Fig. 8 are indicated by black lines.

Figure 8. Regionally-integrated heat budget for the W-FULL minus W-CLIM OHC change over the 1998-2012 period ($\Delta\text{OHC}_{\text{IPO}}$): **(a)** $\Delta\text{OHC}_{\text{IPO}}$ partitioned into the contributions from **(b)** ocean heat transport, **(c)** surface heat fluxes, and **(d)** the residual (see section 2.3 and Eq. (2) for more details). Glo, PacT, PacN, Atl, SO, and Ind respectively denote the global, tropical Pacific, North Pacific, Atlantic, Southern, and Indian Ocean domains, shown with distinct colors on panel (e). The percentages on panel (a) indicate regional contributions to the global OHC change. The colors on panels a, c, and d match those of the corresponding ocean basins on (e). The colors on panel b match those of arrows representing the transport between basins on panel (e). Error bars on panels a–d indicate the 95% confidence intervals. **(e)** Schematic summarizing the heat redistribution during 1998-2012. The grey circles represent OHC changes, white circles surface heat uptake, and arrows exchanges of heat between the basins (value indicated next to each symbol, in 10^{21} J). The area of the circles and arrows is proportional to the heat quantity, with scales given below panel (e).

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Figure 9. Regionally-integrated heat budget for the W-FULL minus W-CLIM OHC change over the 1998-2012 period ($\Delta\text{OHC}_{\text{IPO}}$) partitioned by depth: **(a)** 0-50m, **(b)** 50-300m, **(c)** 300-1200m, **(d)** 1200-6000m. Glo, PacT, PacN, Atl, SO, and Ind respectively denote the global, tropical Pacific, North Pacific, Atlantic, Southern, and Indian Ocean domains, shown with distinct colors in Fig. 8e. Error bars indicate the 95% confidence intervals.

Figure 10. (a) W-FULL minus W-CLIM OHC change over the 1998-2012 period, $\Delta\text{OHC}_{\text{IPO}}$, in 10^9 J m^{-2} , and contributions from **(b)** surface heat fluxes and **(c)** oceanic transport. The colors are plotted only where significant above the 95% confidence level.

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Fig. 11. (a) W-FULL minus W-CLIM 1998–2012 average SLP (hPa, shading); Ekman pumping (blue contours for positive and red contours for negative values; contour interval $1.5 \cdot 10^{-7} \text{ m s}^{-1}$); and wind stress (Pa, vectors). Wind stress vectors are grey inside the tropical Pacific (i.e. the region where the observed wind stress anomaly is prescribed in W-FULL), and black outside it. (b) W-FULL minus W-CLIM 1998–2012 change in ocean heat content, $\Delta\text{OHC}_{\text{IPO}} (10^9 \text{ J m}^{-2})$, shading) and Ekman transport (arrows). (c) Same as (b), with the 1998–2012 average Ekman pumping (blue contours for positive and red contours for negative values; contour interval 10^{-7} m s^{-1}).

Figure 12. Zonal-mean W-FULL minus W-CLIM: **(a)** 1998-2012 zonal wind stress (black, in 10^{-3} Pa). The black dashed line indicates the 1998-2012 anomalous zonal wind stress (relative to 1979-2014) in W-CLIM (10^{-3} Pa). Shading indicates the 95% confidence intervals. **(b)** Global zonal-mean potential temperature change over 1998-2012 ($^{\circ}\text{C}$, shading) and anomalous zonal-mean residual streamfunction (Sv, contours). Note the different vertical scales for the top 1200 m and below. In (b), colors are plotted only where significant above the 95% confidence level.

Figure 13. Scatter plot of average Niño 3 region (150°W-90°W, 5°S-5°N) surface net shortwave radiation (positive downward, in W m⁻²) versus SST (in °C) monthly anomalies in (a) the W-FULL simulation, and (b) observations (TropFlux and OISST). The lines indicate the regression slopes of surface shortwave radiation onto SST, computed separately for positive and negative SST anomalies. The values of the slopes (in W m⁻² K⁻¹) are also indicated.

Figure 14. Sea surface height change from 1998 to 2012 (in cm) for (a) W-FULL minus W-CLIM; (b) AVISO observations. In (a) and (b), we remove the global mean sea surface height increase: this removes the cryosphere-related and land storage, but also the globally-averaged thermosteric effect that is not reproduced by the model (Greatbatch, 1994).

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Figure 15. Atlantic zonal-mean potential temperature change from 1998 to 2012 (in °C, shading), and 1998–2012 average anomalous residual zonal-mean streamfunction (Sv, contours) for W-FULL minus W-CLIM. The colors are plotted only where significant above the 95% confidence level. Note the different vertical scales for the top 1200 m and below.