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Global ocean heat content redistribution during the 1998-2012 Interdecadal Pacific Oscillation negative phase

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23 **Abstract**

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

43

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

46 **1. Introduction**

47 The continued increase in atmospheric greenhouse gas (GHG) concentrations since the
48 Industrial Revolution induces a positive imbalance in the top-of-atmosphere (TOA) planetary
49 energy budget (e.g. Hansen et al., 2005; Trenberth et al., 2009). The overwhelming majority
50 of this excess energy is absorbed by the oceans, as they have a dominant heat capacity
51 (Bindoff et al., 2013; Trenberth et al., 2014). Some of this heat also warms the atmosphere,
52 leading to a steady rise in global mean surface temperature. In the 15-year period from
53 around 1998 to 2012, however, the global mean surface temperature displayed a relatively
54 weak warming rate compared to the previous decades (Easterling et al., 2009). This reduction
55 in the rate of surface warming has been called the *hiatus* or *slowdown* in the global surface
56 warming (Yan et al., 2016; Medhaug et al., 2017). Several recent studies largely attribute this
57 change in the warming rate to natural climate variability, and more specifically to a
58 prolonged abnormally cold sea surface temperature (SST) in the tropical Pacific Ocean
59 (Meehl et al., 2011; Kosaka and Xie, 2013; Fyfe and Gillett, 2014; Clement and DiNezio,
60 2014; Trenberth, 2015; Dai et al., 2015).

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

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

74 The IPO shifted from a warm to a cold phase around 1998, yielding a cooling trend of
75 central and eastern Pacific SST, associated with an unprecedented intensification of the
76 Pacific Walker circulation (Kociuba and Power, 2015; Choi et al., 2016) and strengthening of
77 Pacific trade winds during the 20-year period beginning around 1992 (Balmaseda et al.,
78 2013; England et al., 2014; de Boissésón et al., 2014). Several studies identified this negative
79 IPO phase during the 1998-2012 period as an important driver of the hiatus, due to the
80 associated internal rearrangement of heat within the climate system. A positive imbalance of
81 $\sim 0.5\text{--}1 \text{ W m}^2$ in the TOA planetary energy budget was indeed observed during this period
82 (Loeb et al., 2012; Allan et al., 2014, Johnson et al., 2016), although volcanic activity and
83 anthropogenic aerosols may have contributed to a small reduction in absorbed solar radiation
84 (Solomon et al., 2011; Santer et al., 2014; Ridley et al., 2014; Smith et al., 2015). This
85 observed imbalance implies that the total heat in the climate system steadily increased during
86 the hiatus, and a smaller fraction of this heat was used to warm the atmosphere. Several
87 studies suggested that the abnormally cold Pacific favored an enhanced oceanic heat uptake
88 during this period (Meehl et al., 2011; Kosaka and Xie, 2013, Douville et al., 2015; England
89 et al., 2014, Watanabe et al., 2014), with increased downward heat flux into the ocean in the
90 tropical Pacific, yielding a larger global OHC increase in the subsurface during the negative
91 IPO phase. Drijfhout et al. (2014) found that the ocean heat uptake was also modified in
92 other basins, such as the Southern and subtropical Indian Oceans or the subpolar North
93 Atlantic. Some of these studies used *pacemaker* experiments specifying observed anomalies
94 of either SST (Kosaka and Xie, 2013; Douville et al., 2015) or wind (England et al., 2014;
95 Watanabe et al., 2014; Douville et al., 2015) over the tropical or central/eastern Pacific. This

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

99 A burning question is thus: where did the extra heat pumped into the Pacific during the
100 hiatus decade go? This question is not easy to tackle from observations alone. Deep ocean
101 observations are indeed too sparse to track the OHC changes, even with improved recent
102 estimates (Cheng et al., 2017). The deployment of Argo floats in the early 2000s greatly
103 improved the sampling (Lyman, 2012; Abraham et al., 2013), but the current configuration of
104 Argo floats only reliably samples the upper 2000m of the ocean. In addition, observed
105 changes are a combination of the effects of global warming and natural decadal variability, so
106 that it is not easy to isolate the heat redistribution related to negative IPO phases from
107 observations only (Liu et al., 2016a). There is hence so far little consensus on the location of
108 OHC changes associated with the IPO. Meehl et al. (2011) point to the deep ocean, below
109 750m in Atlantic and the Southern Ocean, and below 300m in the Pacific and the Indian
110 Ocean. England et al. (2014) suggest an increased OHC in the upper 350m of the tropical
111 Pacific. Some studies suggest that a large fraction of the Pacific heat uptake has been
112 transferred to the Indian Ocean via the Indonesian Throughflow (Lee et al., 2015; Nieves et
113 al., 2015; Liu et al., 2016a). However, other studies combining model and observations have
114 pointed out that the La Niña-related Pacific and Indian OHC increases in upper layers are not
115 enough to account for the early 21st century warming slowdown, and that OHC increases at
116 deeper layers in the Atlantic and Southern Ocean account for most of the sequestration (Chen
117 and Tung, 2014; Drijfhout et al., 2014; Roemmich et al., 2015; Llovel and Terray, 2016; Oka
118 and Watanabe, 2017). Cheng et al. (2017) also verified that such a warming in the Atlantic
119 and Southern Ocean has occurred in observations. Liu et al. (2016a, 2016b) have however
120 argued that the deep Atlantic and Southern Ocean observed warming are due to

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

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

135 The paper is structured as follows: section 2 discusses the modelling setup and
136 methodology; section 3 describes the global signals remotely driven by the strengthened
137 easterlies and the resulting cold anomaly in the tropical Pacific; section 4 evaluates the
138 associated ocean heat content changes, and analyses the related surface heat uptake and
139 oceanic heat transport. We summarize and discuss the implications of our results in Section
140 5.

141

142

143 **2. Methods and Data**

144 *2.1 Coupled model*

145 We use the coupled IPSL-CM5A-LR global climate model (Dufresne et al., 2013).
146 The atmospheric component is LMDZ5A (Hourdin et al., 2013), with a resolution of $3.75^\circ \times$
147 1.875° with 39 vertical levels. The land surface component is ORCHIDEE (Krinner et al.,
148 2005), with the same resolution as the atmosphere. The ocean component is NEMOv3.2
149 (Madec, 2008) coupled to the LIM2 sea ice model (Fichefet and Maqueda, 1997). It uses a
150 resolution of about 2° , with a refinement to 0.5° meridional resolution near the equator, and
151 31 vertical levels.

152 IPSL-CM5A-LR simulates the El Niño–Southern Oscillation (ENSO) amplitude,
153 structure, and spectrum reasonably (Bellenger et al., 2014), although it displays an incorrect
154 seasonal phase locking. It simulates well the main characteristics of the North Pacific
155 variability (Fleming and Anchukaitis, 2016), the links between ENSO and IPO (Nidheesh et
156 al., 2017), the tropical intra-seasonal variability (Maury et al., 2013; Hung et al., 2013) and
157 the North Atlantic variability (Cattiaux et al., 2013; Gastineau et al., 2013). However, like
158 many global climate models, IPSL-CM5A-LR has a double intertropical convergence zone
159 (ITCZ) bias and a too strong equatorial upwelling in the Pacific (Dufresne et al. 2013).
160 Compared to other Coupled Model Intercomparison Project Phase 5 (CMIP5; Taylor et al.,
161 2012) climate models, IPSL-CM5A-LR also has relatively large transient and equilibrium
162 climate sensitivities to CO₂ doubling (Forster et al., 2013). This response is mainly due to a
163 large positive low-cloud feedback (Brient and Bony, 2013; Andrews et al., 2012).

164

165 *2.2 Experimental protocol*

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

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

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

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

194 a vertical shear, which in turn generates turbulence and deepens the mixed layer. Therefore,
 195 we add randomly scrambled high-frequency (3–30 day band-passed) wind stress anomalies
 196 from the ERA-Interim reanalysis (Dee et al., 2011) to the model climatological values.
 197 Figure 1b shows the zonal wind stress model climatology averaged over the Niño 4 region
 198 (160°E–150W°, 6°S–6°N). Though the trade winds are weaker in the model, the seasonal
 199 cycle agrees with ERA-Interim, with maximum values in January, and minima in boreal
 200 spring and fall.

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

205

206 *2.3 Ocean heat content and heat transport*

207 OHC is calculated from:

$$208 \quad OHC = \int \rho_0 C_p T dz \quad (1)$$

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

$$212 \quad \Delta OHC = \bar{F} \Delta t + \left(\int \rho_0 C_p \nabla \cdot (\overline{\mathbf{u} \cdot T}) dz \right) \Delta t + R \quad (2)$$

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

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

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

228

229 *2.4 Observations and reanalysis data*

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

240

241 *2.5 Climate indices*

242 We use the Niño 3.4 index as a standard metric for ENSO: the SST anomaly over
243 120° – 170° W and 6° S– 6° N. The east–west thermocline tilt is given by the average 6° S– 6° N

244 20°C isotherm depth in 120°E–170°W minus that in 150°–80°W. In order to analyse the IPO,
245 we use the SST tripole index (TPI, Henley et al. 2015), defined as the central equatorial
246 Pacific SST (170°E–90°W, 10°S–10°N) minus the average of northwest (140°E–145°W,
247 25°–45°N) and southwest (150°E–160°W, 15°–50°S) Pacific SST, smoothed using a 1-3-4-3-
248 1 filter (Trenberth et al., 2007). The equatorial Pacific zonal SLP gradient, defined as the
249 area-averaged SLP in the eastern (160°–80°W, 6°S–6°N) minus that in the western equatorial
250 Pacific (80°–160°E, 6°S–6°N) is used as a proxy of the Walker circulation (Vecchi et al.
251 2006; Kociuba and Power 2015). To evaluate the atmospheric circulation changes in the mid-
252 latitudes we also use the averaged SLP over the North Pacific (as in the North Pacific Index,
253 Trenberth and Hurrell 1994) and the Amundsen Sea, defined over 160°E–140°W, 30°–65°N
254 and 170°E–70°W, 60°–75°S respectively. Lastly, we use the normalized difference between
255 35°N and 65°N of the zonally averaged sea-level pressure (SLP) in 80°W–30°E as a NAO
256 index (Li et al., 2013).

257

258 *2.6 Trend analysis and statistical testing*

259 In time series plots, the ensemble mean is displayed, and the ensemble spread is
260 depicted using 95% confidence intervals for the ensemble mean. We will first focus on the
261 21-year period of strengthening Pacific trade winds (Balmaseda et al., 2013; de Boissésou et
262 al., 2014; England et al., 2014) to investigate if W-FULL successfully describes the
263 associated transition from a positive to negative IPO global pattern. Unless stated otherwise,
264 trends are reported for annual means. The level of statistical significance of simulated trends
265 is evaluated using two-tailed Student *t*-tests, using the values of all members. Correlations
266 between observations and simulations are based on linearly detrended values, and use the
267 ensemble means. The significance level of observed trends and correlations between
268 observations and simulations is calculated using a *t*-test, adjusting the standard error and the

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

274

275

276 **3. Climatic impacts of the increasing tropical Pacific wind stress**

277 *3.1 Pacific sea surface temperature*

278 The observed wind stress anomaly in the Niño 4 region (Fig. 1c) displays a
279 strengthening trend of $-1.65 \cdot 10^{-2}$ Pa decade⁻¹ over 1992–2012 consistent with previous
280 studies (Balmaseda et al., 2013; de Boissésou et al., 2014; England et al., 2014). It also
281 displays clear ENSO-related interannual variability, to which the east-west tilt in the
282 thermocline (Fig. 1d, **0.98** correlation with ORAS4 reanalysis) and SST anomalies in the
283 central Pacific (Fig. 1e, **0.91** correlation with OISST) respond. W-FULL is able to reproduce
284 observed ENSO SST and thermocline variations in response to the prescribed observed wind
285 stress anomalies. This is consistent with the well-known dominant role of wind anomalies for
286 the thermocline depth at the equator and for the ocean temperature (Sarachik and Cane,
287 2010), while the different amplitudes might be due to the heat flux feedback, which depends
288 on the coupled model used (Bellenger et al., 2014). At longer timescales, W-FULL
289 reproduces the increasing east-west thermocline tilt and cooling SST trend in the central
290 Pacific associated with strengthening easterlies (see Table 1). Overall, W-FULL reproduces
291 the observed tropical Pacific decadal variability, capturing the transition from the more
292 predominant El-Niños in the 1990s to the more predominant La Niñas in the 2000s.

293 Conversely, imposing the wind stress climatology in W-CLIM (Figs. 1d and 1e, red
294 line) yields no interannual variability in thermocline tilt. As a result there is no interannual
295 variability in SST, except for the cooling following the Pinatubo (1991) and El Chichón
296 (1982) volcanic eruptions. W-CLIM also reveals a warming trend associated with the
297 imposed evolution of GHG (Table 1), that is difficult to see in W-FULL due to the larger
298 interannual and decadal variations.

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

304

305 *3.2 Surface air temperature*

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

316 Figure 2d displays annual global mean 2m air temperature time series, with the 1992–
317 2012 trend values in Table 1. The 1992–2012 ensemble mean trend is significantly smaller in

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

326

327 *3.3 Large-scale atmospheric circulation changes*

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

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

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

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

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

369

370 **4. Evaluation of the ocean heat content changes**

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

378

379 *4.1 Global energy balance and ocean heat content evolution*

380 We first diagnose and quantify the global OHC changes and their causes. W-FULL
381 and W-CLIM both display a stable TOA radiative imbalances from 1995 onward of about 1.2
382 W m^{-2} (not shown), when accounting from energy nonconservation in the atmosphere by
383 removing the value from the TAO imbalance calculated over the whole preindustrial control
384 simulation, as in Hobbs et al. (2016). This value is larger than the present-day observed
385 estimate, which ranges between 0.5 and 1 W m^{-2} (Loeb et al. 2012; Allan et al. 2014; Johnson
386 et al., 2016), and reflects the large climate sensitivity of IPSL-CM5A-LR (Brient and Bony
387 2013; Andrews et al. 2012). The monthly globally-integrated OHC increases throughout
388 1995-2014 in both W-FULL and W-CLIM due to this TOA imbalance (Fig. 4a).

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

392 linear superposition of the global warming simulated in W-CLIM and the IPO-related
393 changes, taking the difference W-FULL minus W-CLIM should also remove the changes
394 induced by the large climate sensitivity of the model used. The average monthly heat flux
395 into the ocean (positive downward) in Fig. 4b displays sustained positive anomalies after
396 1998, and variations from positive to negative anomalies before 1998. This is related to a
397 larger TOA radiative imbalance of 0.11 (0.02) W m^{-2} in W-FULL during 1998-2012 (1979-
398 1997) linked to the more frequent La Niña after 1998, as found in observations (Allan et al.,
399 2014). As a result, the accumulated heat into the global ocean (orange curve on Fig. 4c)
400 displays a steady rise from 1998 onward. The global OHC variations (black curve on Fig. 4c)
401 closely track those of the integrated heat fluxes, which confirms that the heat budget is closed
402 at the global scale. As 1998-2012 experiences most of the IPO-related OHC increase,
403 hereafter we will focus on this period to track the OHC changes associated with the IPO
404 negative phase. We will discuss the OHC changes occurring before 1998 in section 5.

405

406 *4.2 SST and OHC changes during the period 1998–2012*

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

416 The OHC changes between early 1998 and late 2012 (hereafter, ΔOHC) in W-CLIM
417 (Fig. 6a) display a relatively uniform increase, with heat content mostly increasing in the top
418 300m (not shown). OHC actually diminishes between 10°N and 20°N in the Pacific,
419 presumably in response to a northward shift of the ITCZ. It also decreases in the eastern
420 North Atlantic as the Atlantic Meridional Overturning circulation (AMOC) slows down
421 (Drijfhout et al., 2012). ΔOHC in W-CLIM however bears little resemblance to the ORAS4
422 estimate (Fig. 6c). On the other hand, the W-FULL ΔOHC (Fig. 6b) displays a much better
423 agreement with observations (Fig. 6c). In particular, OHC diminishes over the eastern tropical
424 Pacific and increases over the western tropical Pacific, consistent with the anomalously strong
425 easterlies during that period (e.g. England et al. 2014). While this is expected due to the
426 prescribed wind stress anomalies in the tropical Pacific in W-FULL, W-FULL also better
427 reproduces OHC changes in other regions. For instance, both W-FULL and ORAS4 display
428 increasing OHC over the Kuroshio extension, the southwestern Pacific, and (to a lesser
429 extent) the eastern tropical Indian Ocean and Gulf Stream extension regions. The better
430 agreement of W-FULL than W-CLIM with the observed 1998-2012 OHC changes shows that
431 our methodology constrains reasonably the OHC changes associated with the IPO, at least
432 over the Indo-Pacific.

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

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

444

445 *4.3 Regional heat budget for the IPO-induced OHC changes*

446 We now investigate the contributions from surface fluxes, oceanic transport, and the
447 residue, as in Eq. (2), for $\Delta\text{OHC}_{\text{IPO}}$. Figure 8 shows this regionally-integrated budget for the
448 tropical Pacific, North Pacific (north of 31°N), Atlantic (including the Arctic), Southern
449 (south of 31°S), and Indian Ocean basins (basin boundaries are depicted in Figs 7 and 8e).

450 The residual in Eq. (2) (Fig. 8d) represents the contributions of parameterized diffusive lateral
451 mixing and changing sea ice heat content. This residual is in general close to zero or much
452 weaker than the dominant terms in the budget, so that OHC changes can be accounted for by
453 the sum of the surface heat uptake and ocean heat flux transport convergence.

454 We find that the anomalous surface heat uptake during the 1998-2012 IPO negative
455 phase (Fig. 8c) exclusively occurs in the tropical Pacific Ocean ($65 \pm 4 \times 10^{21}$ J), with other
456 basins experiencing either negligible IPO-induced surface fluxes (Atlantic, Southern Ocean)
457 or more moderate heat losses to the Atmosphere (Indian Ocean, North Pacific). Despite this
458 major heat input from the atmosphere, the tropical Pacific OHC hardly changes over the
459 1998-2012 period (Fig. 8a). The tropical Pacific surface heat uptake is indeed entirely
460 transferred to other basins. The ratio of the oceanic transports from the tropical Pacific (Fig.
461 8b and arrows in Fig. 8e) versus the surface heat uptake (Figs. 8c and 8e) shows that 53% of
462 the heat received at the surface of the tropical Pacific is transferred to the North Pacific, 26%
463 to the Indian Ocean, and 21% to the South Pacific. Overall, 65% of the global $\Delta\text{OHC}_{\text{IPO}}$
464 occurs in the Southern Ocean ($26 \pm 11 \times 10^{21}$ J, as illustrated in Fig. 8a), where the surface heat
465 loss is negligible (Fig. 8c). The ratio of oceanic heat transport (arrows in Fig. 8e) versus the

466 southern ocean $\Delta\text{OHC}_{\text{IPO}}$ illustrates that 49% of the OHC increase is due to the heat transport
467 across 31°S from the tropical Pacific, 32% is due to the heat transport from the Indian ocean,
468 and 19% is due to the transport from the Atlantic Oceans. Despite receiving the largest heat
469 transport from the tropical Pacific, the North Pacific experiences a smaller OHC increase than
470 the Southern Ocean (40% of the global OHC increase), due to heat losses to the atmosphere
471 ($13\pm 3 \times 10^{21}$ J, see Fig. 8c and 8e). In the Indian Ocean, about half of the input of heat from
472 the tropical Pacific is evacuated southward to the Southern Ocean via the Indonesian
473 throughflow (Fig. 8e), with a relatively large heat quantity ($17\pm 2 \times 10^{21}$ J) released back to the
474 atmosphere. As a result, the Indian Ocean experiences little $\Delta\text{OHC}_{\text{IPO}}$. In the Atlantic, the
475 IPO induces little changes in either atmospheric heat uptake or transport, with no significant
476 $\Delta\text{OHC}_{\text{IPO}}$ as a result, as will be discussed in section 5.

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

486 Figure 10 shows the spatial distributions of the heat budget terms. $\Delta\text{OHC}_{\text{IPO}}$ (Fig.
487 10a) is repeated from Fig. 7a, while Fig. 10b and 10c are the contributions from surface fluxes
488 and ocean heat transport. Table 2 quantifies these contributions to the global and regional
489 $\Delta\text{OHC}_{\text{IPO}}$ using pattern regression: 68% of the global OHC changes attributable to the IPO
490 are driven by the oceanic transport. The tropical Pacific is the only region with a dominant

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

501 In the North Pacific, Indian, and Atlantic Oceans, ocean dynamics are the dominant
502 contributor to $\Delta \text{OHC}_{\text{-IPO}}$, with surface fluxes to the atmosphere acting as a negative feedback
503 (Fig. 10, Table 2). In particular, ocean heat transport changes explain the two bands of
504 $\Delta \text{OHC}_{\text{-IPO}}$ increase around 40° latitude in both hemispheres. This anomalous oceanic heat
505 transport convergence is mainly due to the eulerian transport, with only a small contribution
506 from the eddy bolus transport (not shown). In the following section, we will focus on the
507 mechanisms that explain the main IPO-induced ocean heat flux convergence seen in Figs. 8
508 and 10c.

509

510 *4.4 Mechanisms of IPO-induced ocean heat transport*

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

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

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

529 The influence of these processes is clearly seen in the anomalous W-FULL zonal-
530 mean temperature change (Fig. 12b) that is dominated by the Pacific Ocean anomalies. The
531 positive temperature anomalies at 40°S and 40°N penetrate downward into the ocean where
532 the mean slope of the isopycnals is large (not shown), which enhances the heat penetration
533 into the deeper layers through the mean flow. We also note an intensification of the shallow
534 subtropical cells between 20°S and 20°N in the upper 400m of the ocean (see contours in Fig.
535 12b) as described in England et al. (2014), but these only lead to shallow temperature
536 anomalies. Such anomalies are associated with the large OHC increase found at 50-300m in
537 the tropical Pacific (see Fig. 9). A positive anomaly of the global meridional streamfunction is
538 also found between 65°S and 50°S, as the Deacon cell intensifies and shifts southwards.

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

541 changes specifically to either climate change or natural variability. Using our experimental
542 setup, we can quantify those two influences. Figure 12a shows the zonal-mean wind stress
543 anomalies associated with the negative IPO phase (black line) and the anomalies simulated in
544 W-CLIM (black dashed line). In W-CLIM, increasing greenhouse gases and decreasing
545 stratospheric ozone yield positive wind stress anomalies around 50°S (Shindell and Schmidt,
546 2004). The IPO-related (i.e. W-FULL minus W-CLIM) anomalies are similar, but they are,
547 however, larger and 10° further south. Similarly, we evaluate that the Southern Ocean heat
548 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
549 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-
550 CLIM. For comparison, the average rate for the Southern Ocean heat storage is 33.1 ± 2.5
551 $(39.2 \pm 2.9) \text{ } 10^{21} \text{ J decade}^{-1}$ over 1979-2014 in W-CLIM (W-FULL, respectively). Therefore,
552 external forcings have caused an accelerated Southern Ocean OHC increase during the
553 negative IPO period (increased rate of $23.1 \pm 4.3 \%$), but we find that these changes are
554 smaller than those induced by tropical SSTs through atmospheric teleconnections (increase of
555 $48.1 \pm 8.6 \%$).

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

562

563 **5. Summary and discussion**

564

565 *5.1. Summary*

566 In this study, we examine the OHC changes over the 1998-2012 period, during which
567 global mean surface temperature displayed a reduced warming rate, which has been
568 attributed to oceanic heat redistribution and enhanced heat uptake in the Pacific (Meehl et al.,
569 2011; Kosaka and Xie, 2013; Fyfe and Gillett, 2014; Clement and DiNezio, 2014; Trenberth,
570 2015; Dai et al., 2015). To test this hypothesis, we prescribe the 1979–2014 tropical Pacific
571 (20°N-20°S) surface wind stress in two ensembles with identical GHG, aerosol, and ozone
572 forcing: one with climatological wind stress (W-CLIM), and the other with wind stress
573 anomalies from ERA-Interim (W-FULL). As already explored in previous studies (England
574 et al. 2014; Watanabe et al. 2014; Douville et al. 2015; Delworth et al. 2015; Thoma et al.
575 2015; Saenko et al. 2016), we find that this strategy is successful in constraining ENSO
576 variations in the tropical Pacific and their decadal modulation, with an El Niño residual
577 signature during 1979-1997 and a La Niña signature during 1998-2012. Atmospheric
578 teleconnections also produce global air temperature and SLP patterns characteristic of a
579 transition from a positive to a negative IPO phase. As a result, W-FULL also simulates the
580 observed reduced rate in global surface warming during 1992–2012 (Gleisner et al. 2015).

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

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

601

602 *5.2. Discussion*

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

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

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

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

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

661 The Atlantic Ocean is also influenced by the Pacific variability in W-FULL. The
662 1998–2012 Atlantic zonal-mean changes linked to the cold IPO phase are shown in Fig. 15.
663 The IPO-related signal in our simulation is weak in the Northern Atlantic. In contrast, Chen
664 and Tung (2014) find a large temperature increase over the whole north Atlantic from

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

682 Lastly, our working hypothesis is that external forcings have no impact on the Pacific
683 trade wind intensification during 1998-2012, so that we attribute the differences between the
684 results of W-FULL and W-CLIM entirely to IPO variability. However, some studies have
685 suggested that external forcing may have lead to a significant part of the observed trade wind
686 strengthening (Luo et al. 2012; Han et al. 2014; Kohyama et al., 2016; Dong and McPhaden
687 2017), in particular due to accelerated Indian Ocean warming. This is in contradiction to
688 other studies that indicate that anthropogenic climate change should drive a weakening of the
689 Walker circulation (e.g. Held and Soden, 2005; Kociuba and Power, 2015). Other studies

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

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

705

706

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721

722

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1073 **Table 1:**1992–2012 trends of different climate indices associated with the IPO transition from
1074 the positive to the negative phase. The 95% confidence intervals are also given. Note that the
1075 confidence intervals are calculated differently for the models and observations. See Section
1076 2.5 and 2.6 for details.

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

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1082 **Table 2.** Pattern regression (unitless) of the ocean surface heat uptake (first term in the r.h.s
1083 of Eq. (2)) and ocean heat flux convergence (second term in the r.h.s of Eq. (2)) during the
1084 1998-2012 period onto $\Delta\text{OHC}_{\text{-IPO}}$ (l.h.s. of Eq. (2)). The two pattern regression coefficients
1085 sum to 1.0 by construction, and quantify the relative contributions of surface-flux and
1086 circulation-induced OHC changes to the total.

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	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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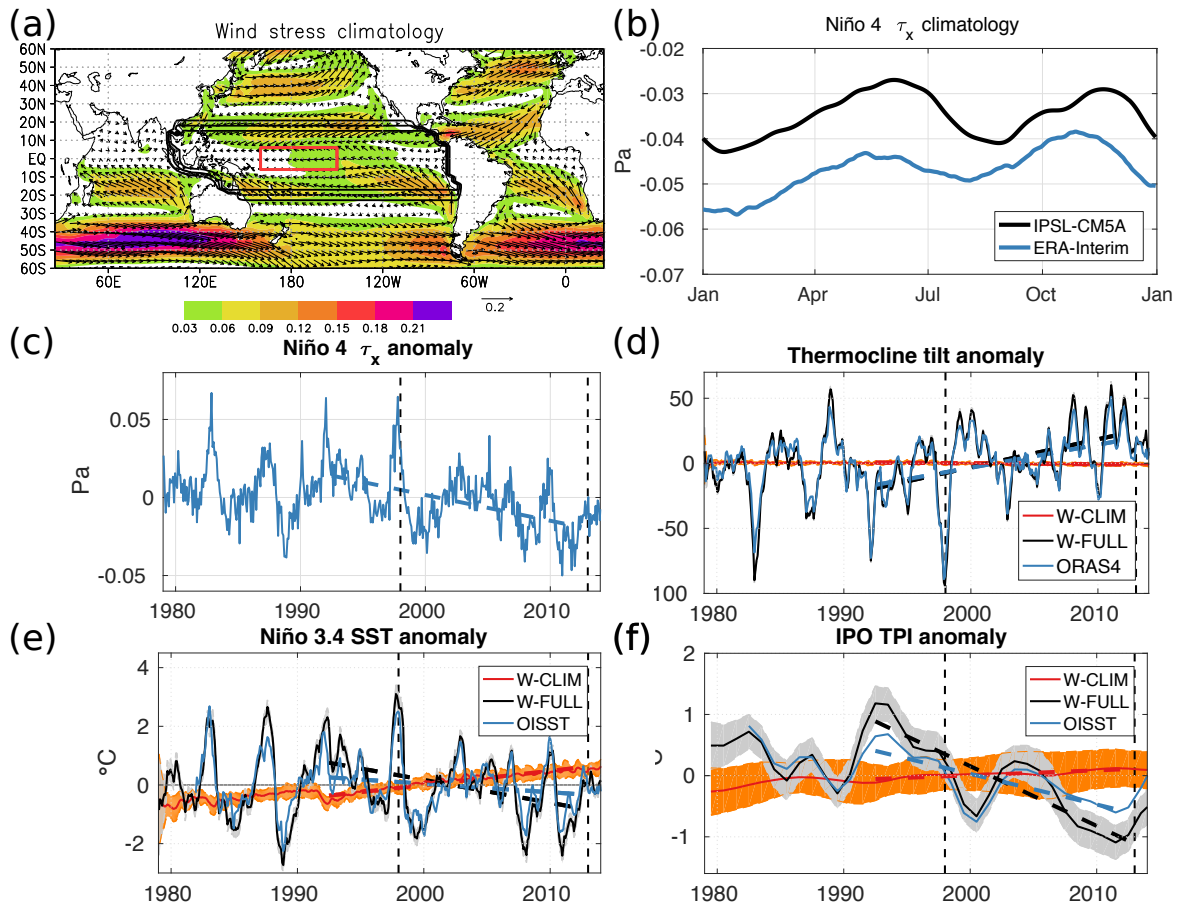
1091 **Table 3.** W-FULL minus W-CLIM ensemble mean differences of SST (in °C), surface heat
1092 fluxes (positive downward, in $W m^{-2}$) and total cloud cover (TCC) over the entire tropical
1093 Pacific (see Fig. 8e for basin boundaries), for the 1998–2012 (negative IPO) and 1979–1997
1094 (positive IPO) periods. The 1979–2014 mean value for W-FULL is also given as a reference.
1095 Total HF designates the total heat flux. SW_{net} CS is the clear-sky surface shortwave flux.

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	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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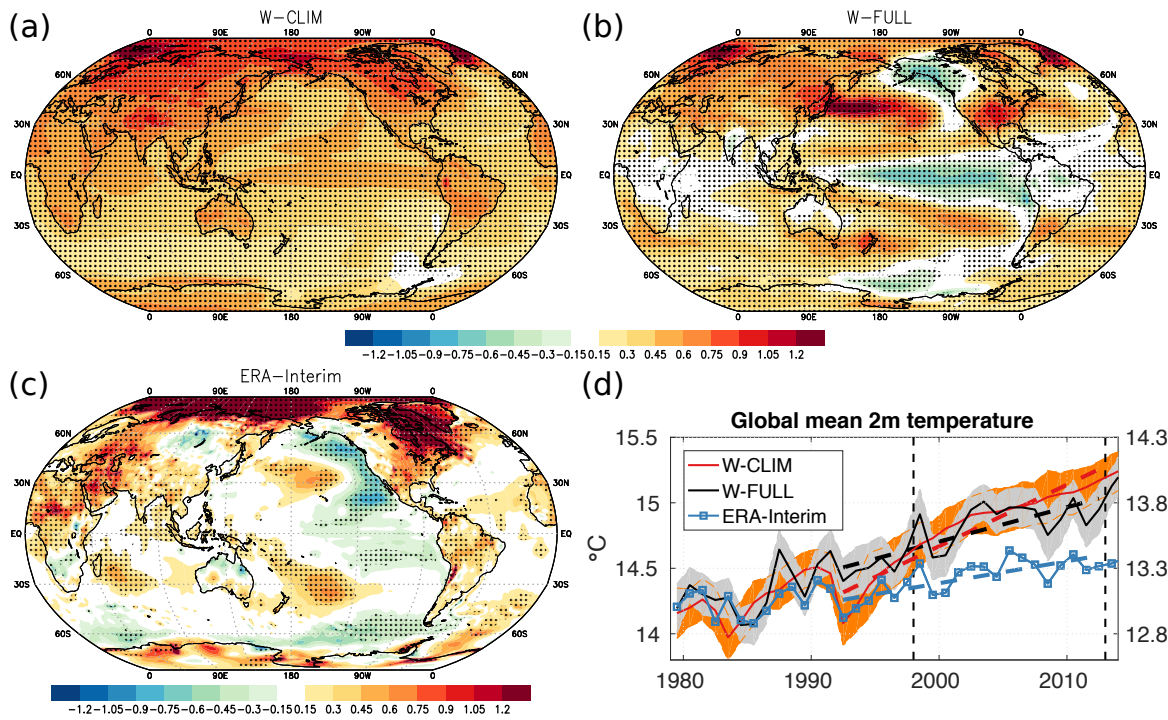
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1103 **Figure 1:** (a) Mean 1979–2014 IPSL-CM5A-LR surface wind stress climatology (vectors and
1104 shading), in Pa. Colors indicate the wind stress magnitude. The black contours (contour
1105 intervals 0.3, 0.6, and 0.9) delineate the mask used to 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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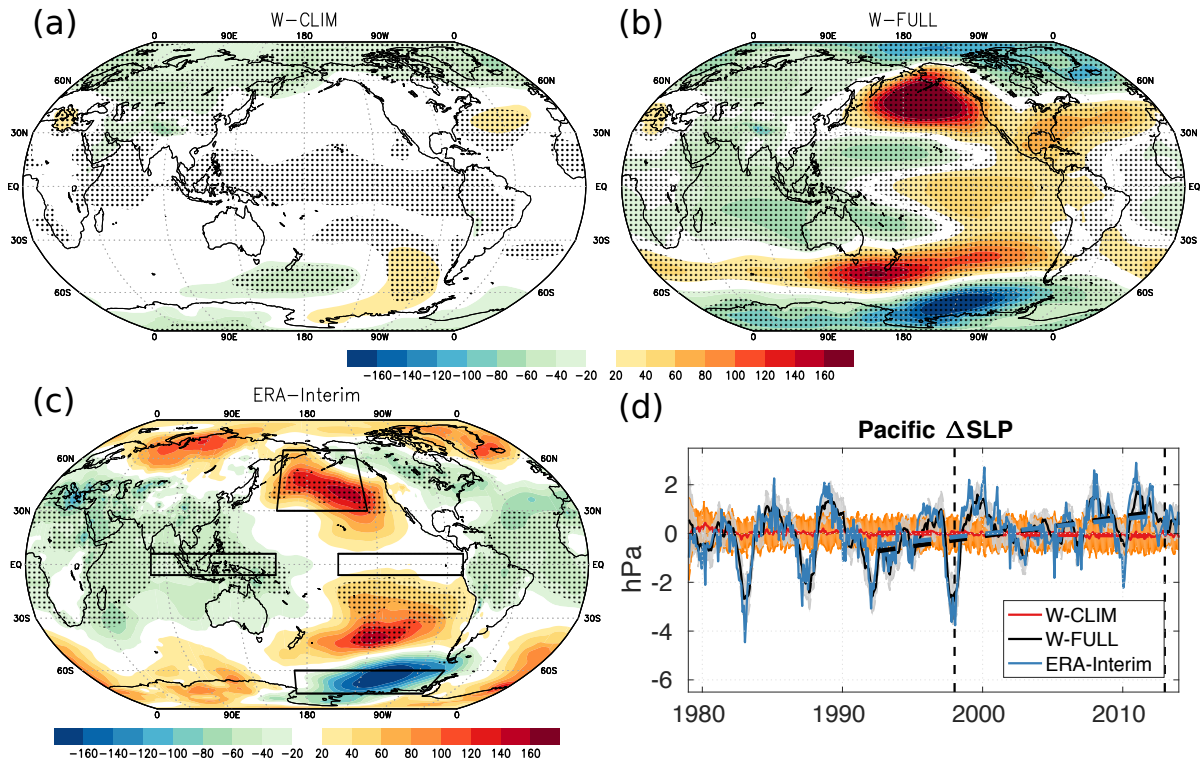


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1120 **Figure 2.** 1992-2012 2m temperature trend, in °C per decade for (a) W-CLIM, (b) W-FULL,
1121 and (c) ERA-Interim. Stippling denotes regions where the trends are significant at the 95%
1122 confidence level. (d) Globally-averaged annual mean 2m temperature, in °C, for W-CLIM
1123 (red), W-FULL (black), and ERA-Interim (blue); note that the simulations are offset by 1.2°C
1124 to correct for the mean temperature bias (right axis). Thick solid lines indicate the ensemble
1125 means; shading illustrates the ensemble spread, with the 95% confidence intervals. Dashed
1126 lines show the 1992–2012 linear trends. The vertical dashed lines mark 1 January 1998 and 1
1127 January 2013.

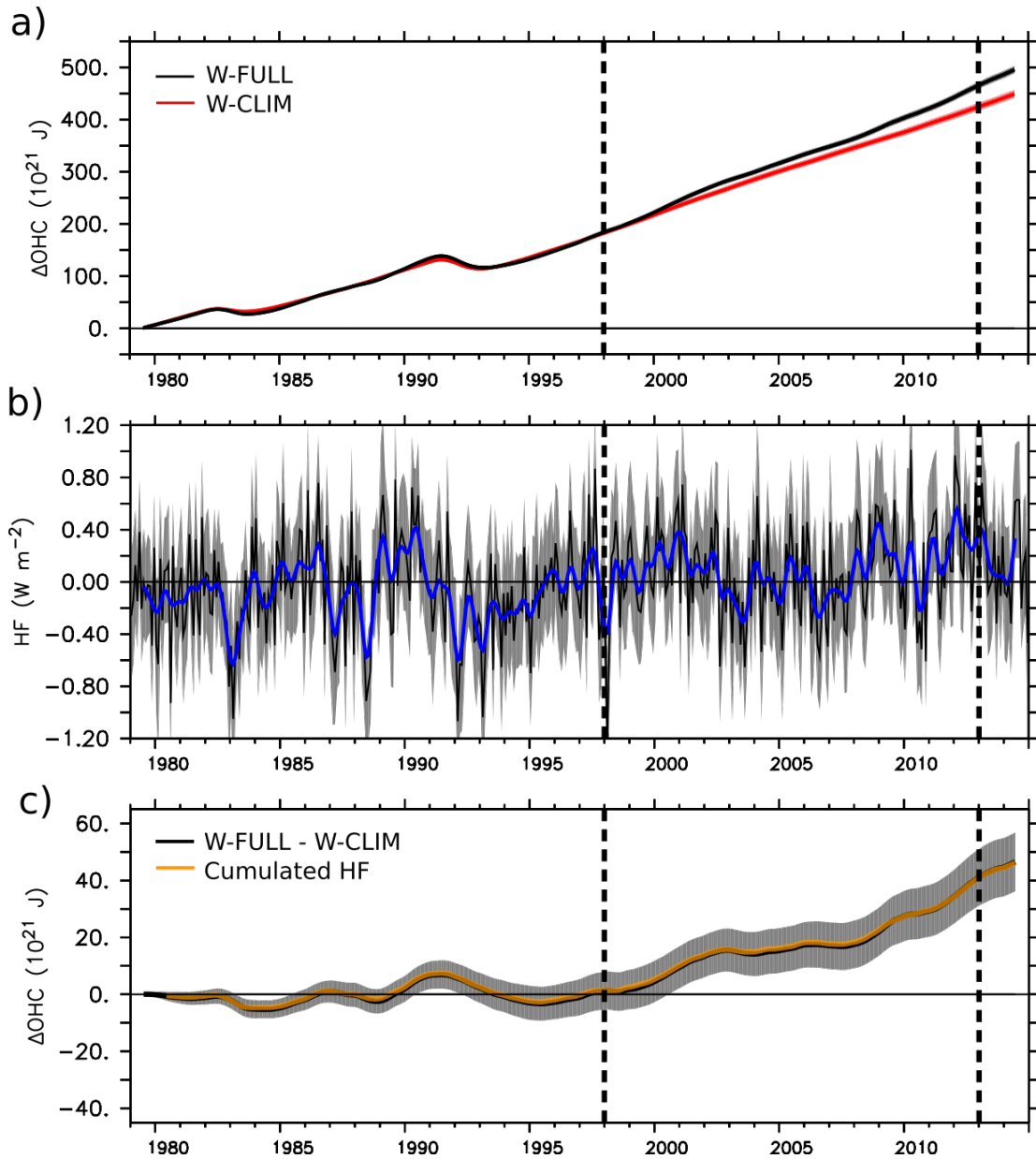
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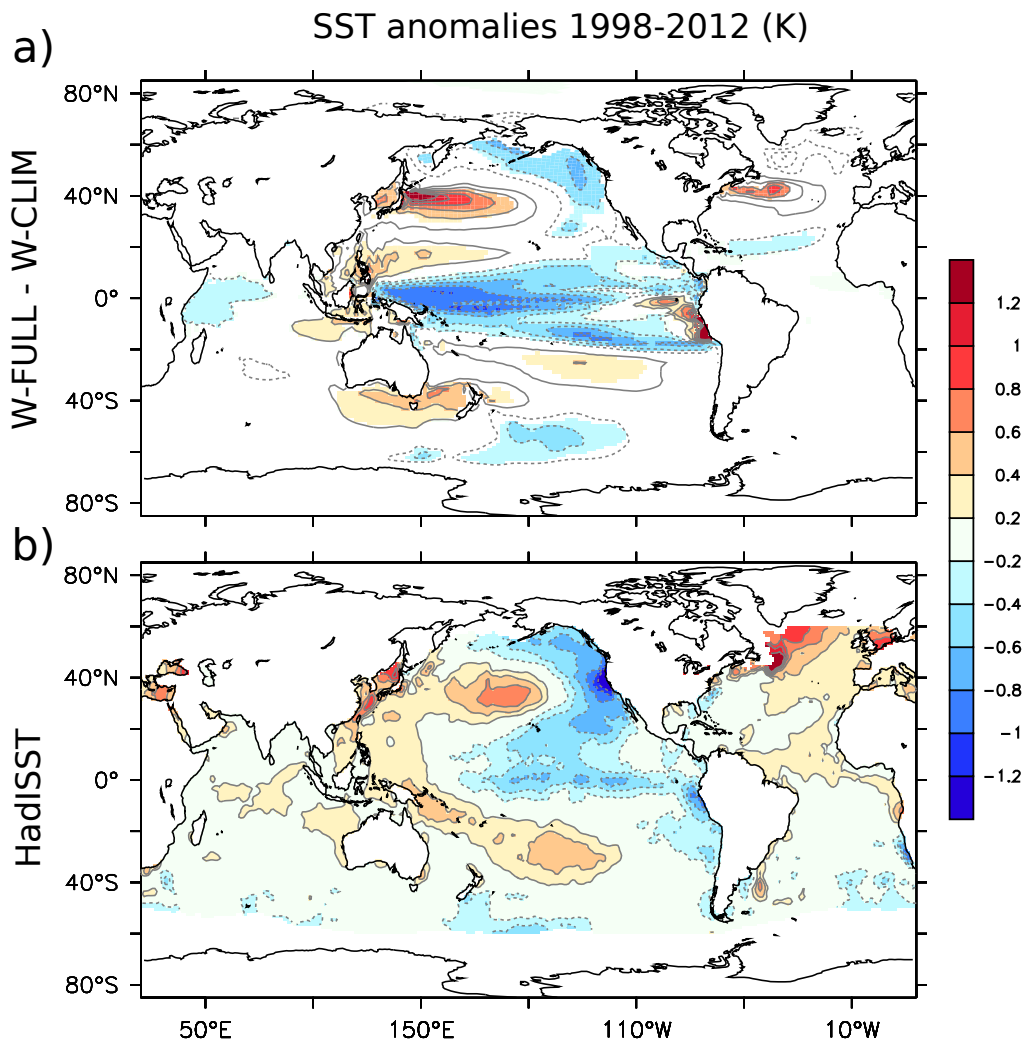
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1134 **Figure 3.** 1992–2012 annual mean SLP trends, in Pa per decade for (a) W-CLIM, (b) W-
1135 FULL, and (c) ERA-Interim. Stippling denotes regions that are significant at the 95%
1136 confidence level. Boxes in (c) delineate the regions used to compute the average SLP in the
1137 western and eastern equatorial Pacific (used in Δ SLP), North Pacific, and Amundsen Sea in
1138 Table 1. (d) Monthly Pacific Δ SLP anomalies, in hPa. Thick solid lines in (d) indicate the
1139 ensemble means; shading illustrates the ensemble spread, with the 95% confidence intervals;
1140 dashed lines show the 1992–2012 trends. The vertical dashed lines mark 1 January 1998 and 1
1141 January 2013.



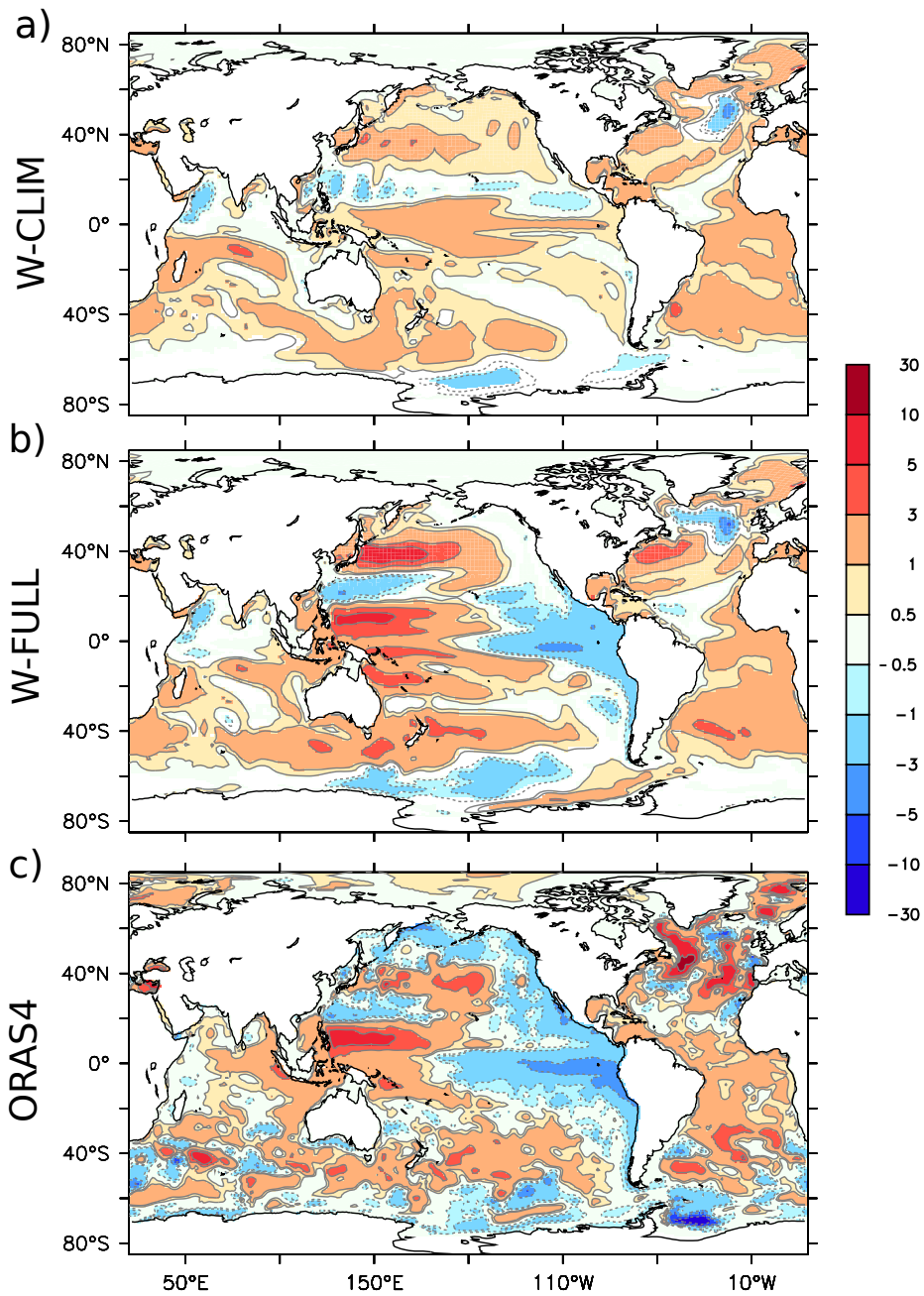
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Figure 4. (a) Global W-FULL (black) and W-CLIM (red) ensemble mean OHC, in 10^{21} J. (b) Globally-averaged W-FULL minus W-CLIM ensemble mean deseasonalized monthly surface heat flux into the ocean, in W m^{-2} (positive downward). The thin black line 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.



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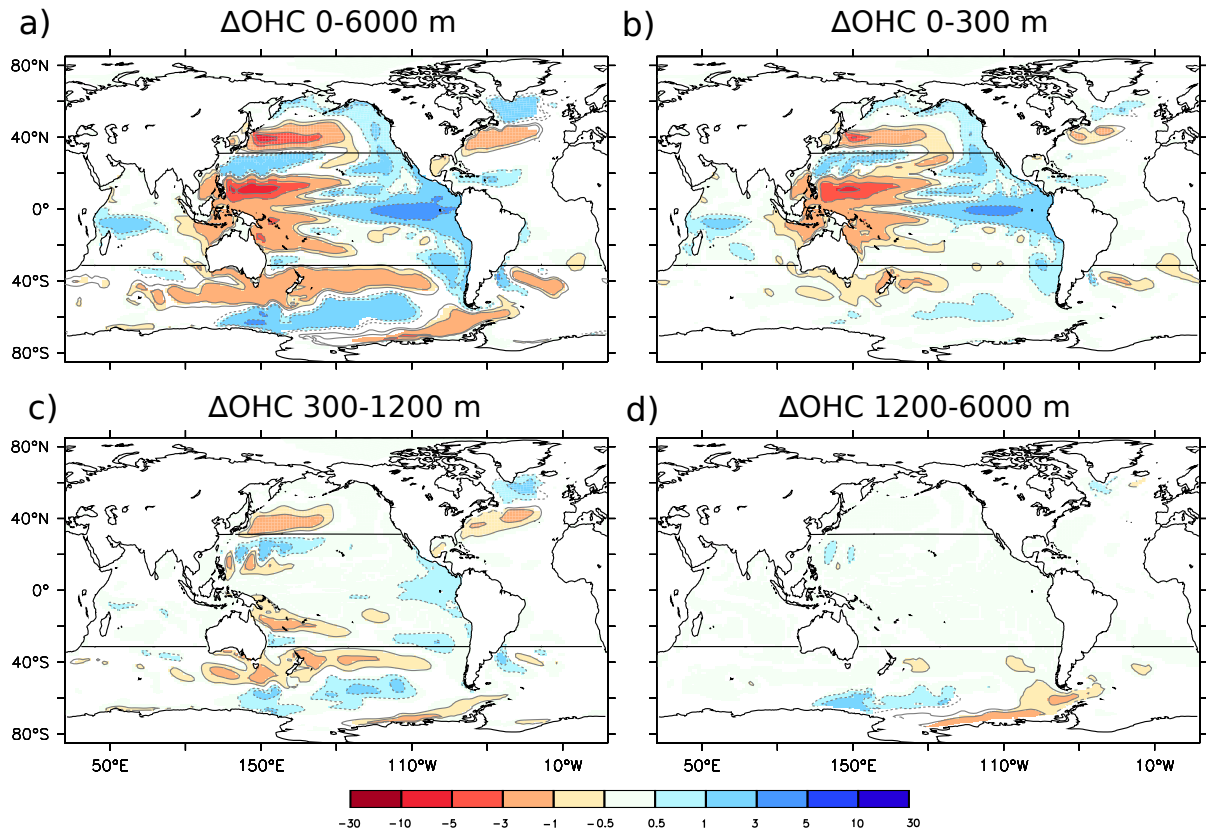
Figure 5. 1998–2012 average SST anomalies (in °C), from: **(a)** W-FULL minus W-CLIM; with shading indicating regions statistically significant at the 95% confidence level; **(b)** HadISST observations after removal of the global warming signal using an optimal linear inverse modeling filter.



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1162 **Figure 6.** OHC change over the 1998-2012 period, ΔOHC , shown here in 10^9 J m^{-2} for the
1163 entire ocean column, in (a) W-CLIM, (b) W-FULL and (c) ORAS4.

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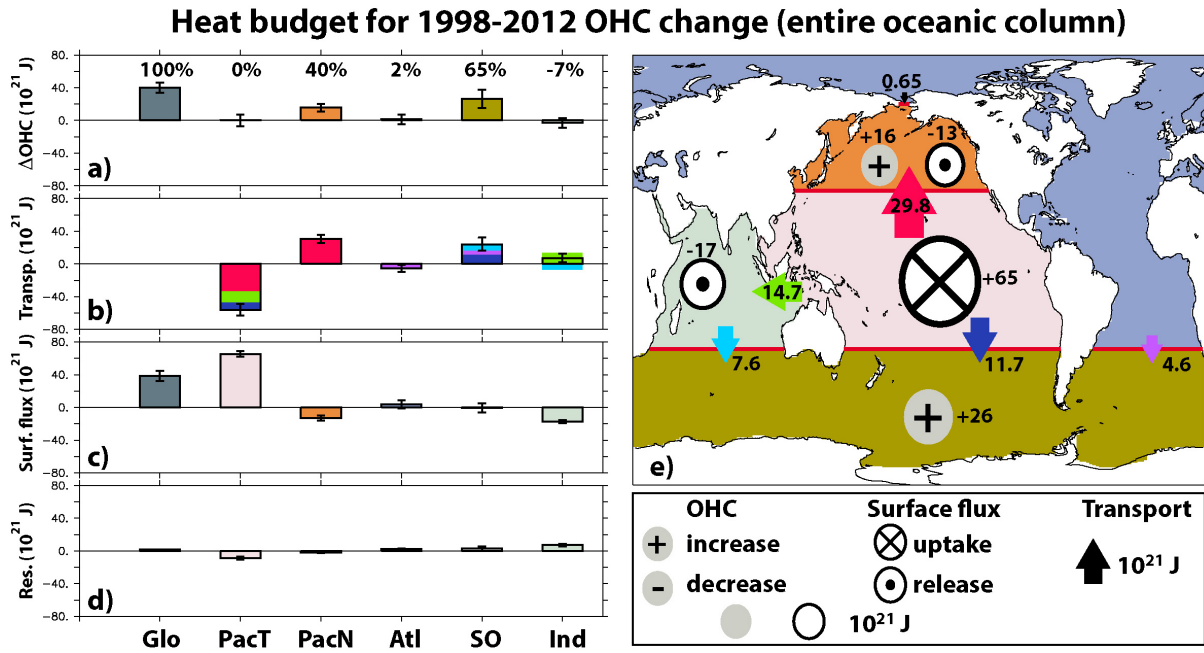


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1167 **Figure 7.** Vertical distribution of the W-FULL minus W-CLIM OHC change over the 1998-
1168 2012 period, $\Delta\text{OHC}_{\text{IPO}}$, in 10^9 J m^{-2} : **(a)** entire ocean column, **(b)** 0-300m,
1169 **(c)** 300-1200m, and **(d)** 1200m-6000m. The colors are plotted where significant above the 95% confidence
1170 level. The boundaries of the different basins used in Fig. 8 are indicated by black lines.

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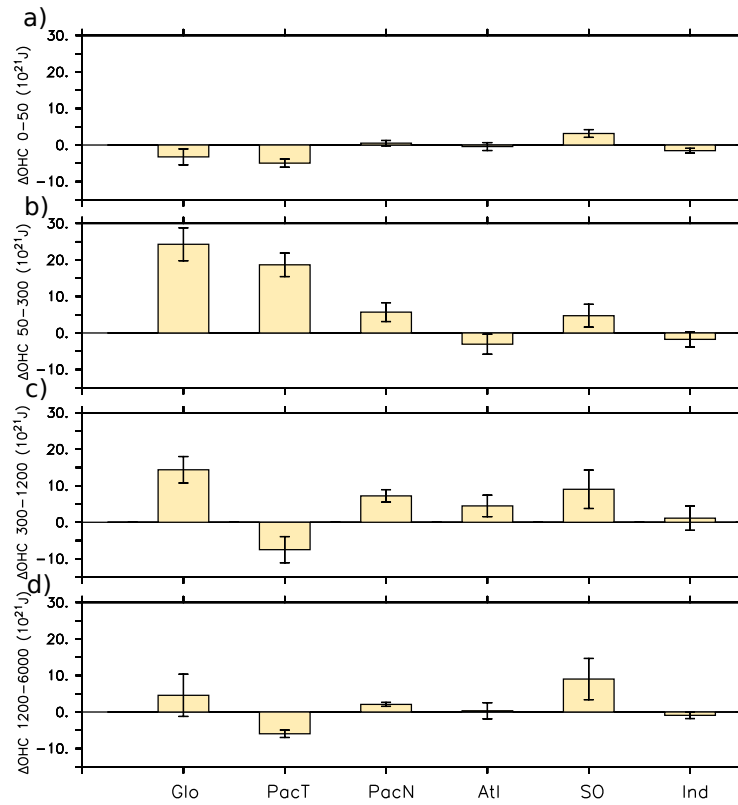
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1176 **Figure 8.** Regionally-integrated heat budget for the W-FULL minus W-CLIM OHC change
 1177 over the 1998-2012 period ($\Delta\text{OHC}_{\text{IPO}}$): **(a)** $\Delta\text{OHC}_{\text{IPO}}$ partitioned into the contributions from
 1178 **(b)** ocean heat transport, **(c)** surface heat fluxes, and **(d)** the residual (see section 2.3 and Eq.
 1179 (2) for more details). Glo, PacT, PacN, Atl, SO, and Ind respectively denote the global,
 1180 tropical Pacific, North Pacific, Atlantic, Southern, and Indian Ocean domains, shown with
 1181 distinct colors on panel (e). The percentages on panel (a) indicate regional contributions to the
 1182 global OHC change. The colors on panels a, c, and d match those of the corresponding ocean
 1183 basins on (e). The colors on panel b match those of arrows representing the transport between
 1184 basins on panel (e). Error bars on panels a–d indicate the 95% confidence intervals. **(e)**
 1185 Schematic summarizing the heat redistribution during 1998-2012. The grey circles represent
 1186 OHC changes, white circles surface heat uptake, and arrows exchanges of heat between the
 1187 basins (value indicated next to each symbol, in 10^{21} J). The area of the circles and arrows is
 1188 proportional to the heat quantity, with scales given below panel (e).

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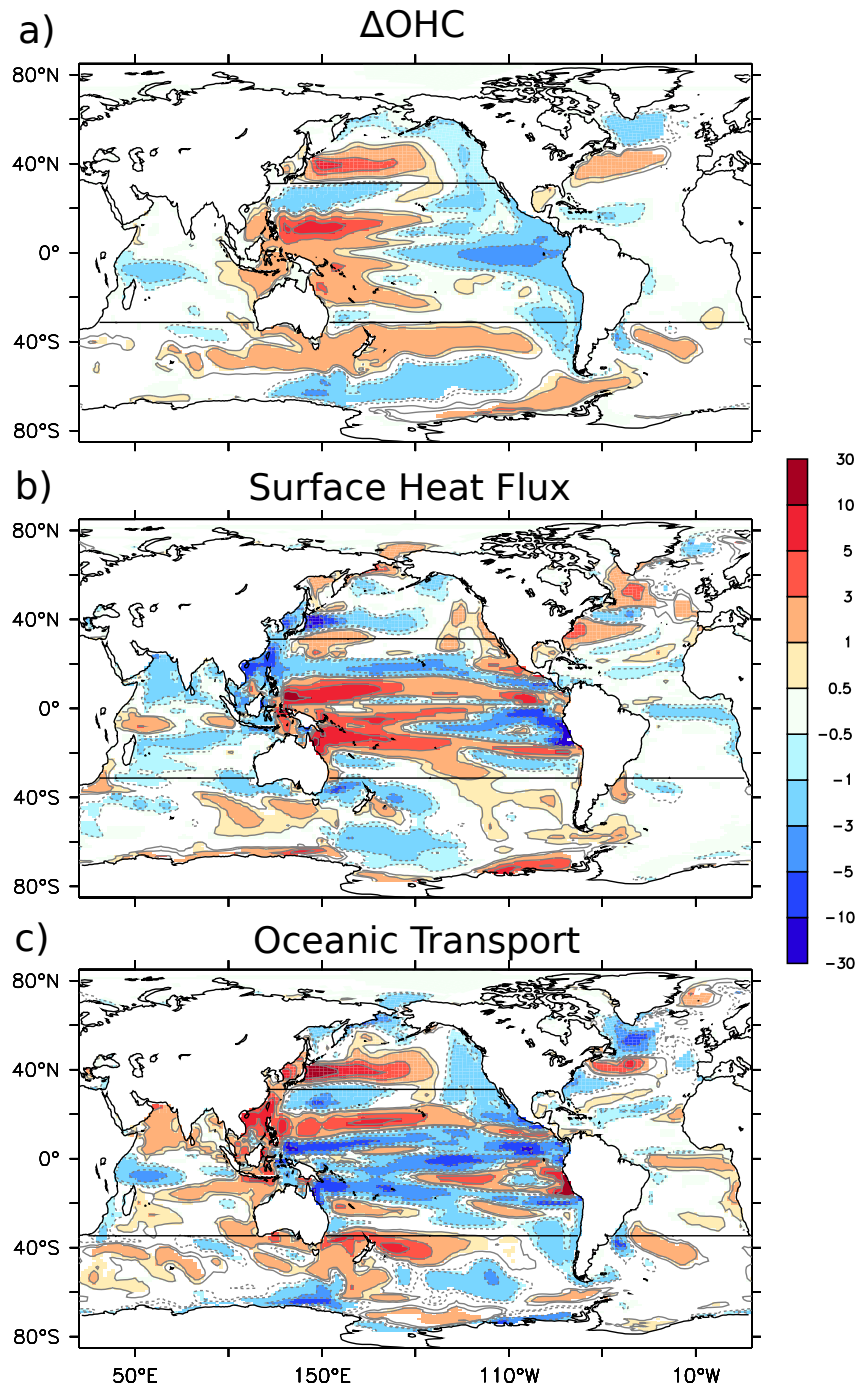


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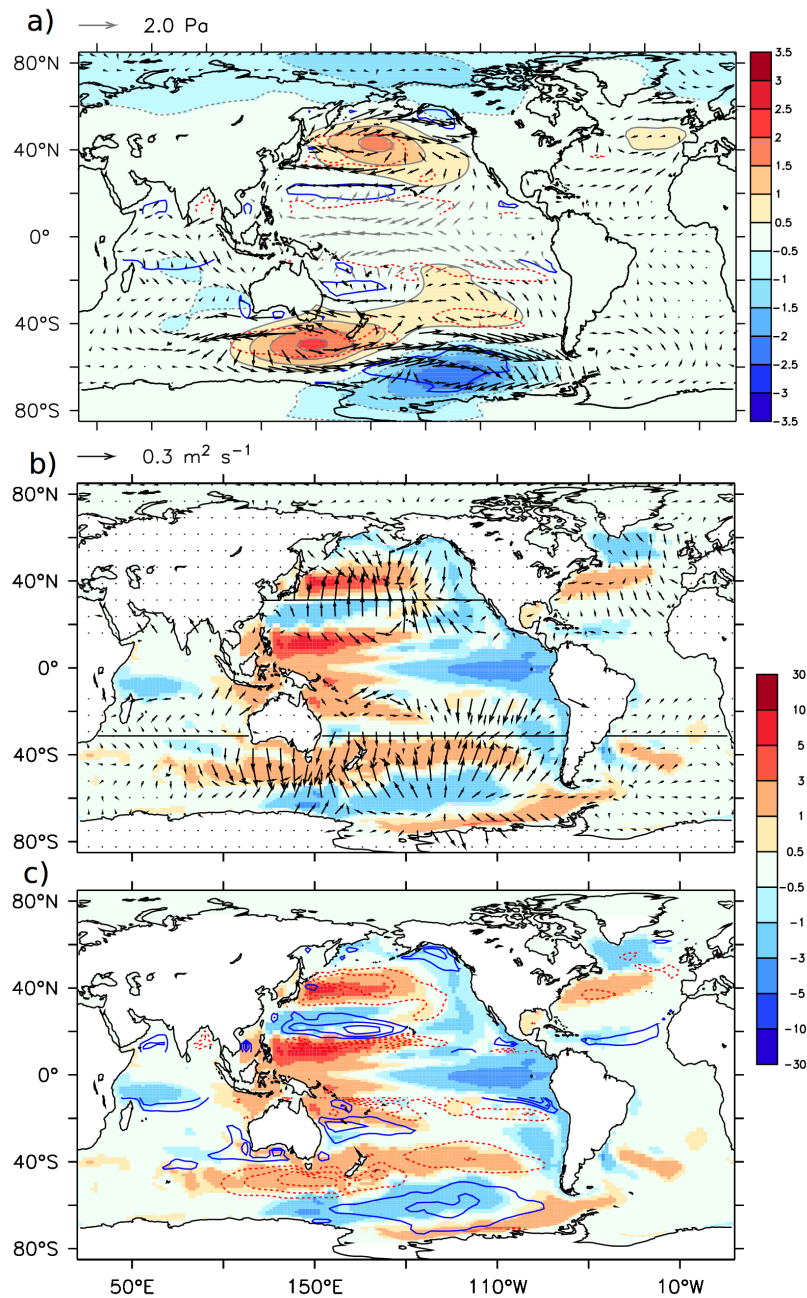
1194 **Figure 9.** Regionally-integrated heat budget for the W-FULL minus W-CLIM OHC change
1195 over the 1998-2012 period ($\Delta\text{OHC}_{\text{IPO}}$) partitioned by depth: **(a)** 0-50m, **(b)** 50-300m, **(c)**
1196 300-1200m, **(d)** 1200-6000m. Glo, PacT, PacN, Atl, SO, and Ind respectively denote the
1197 global, tropical Pacific, North Pacific, Atlantic, Southern, and Indian Ocean domains, shown
1198 with distinct colors in Fig. 8e. Error bars indicate the 95% confidence intervals.

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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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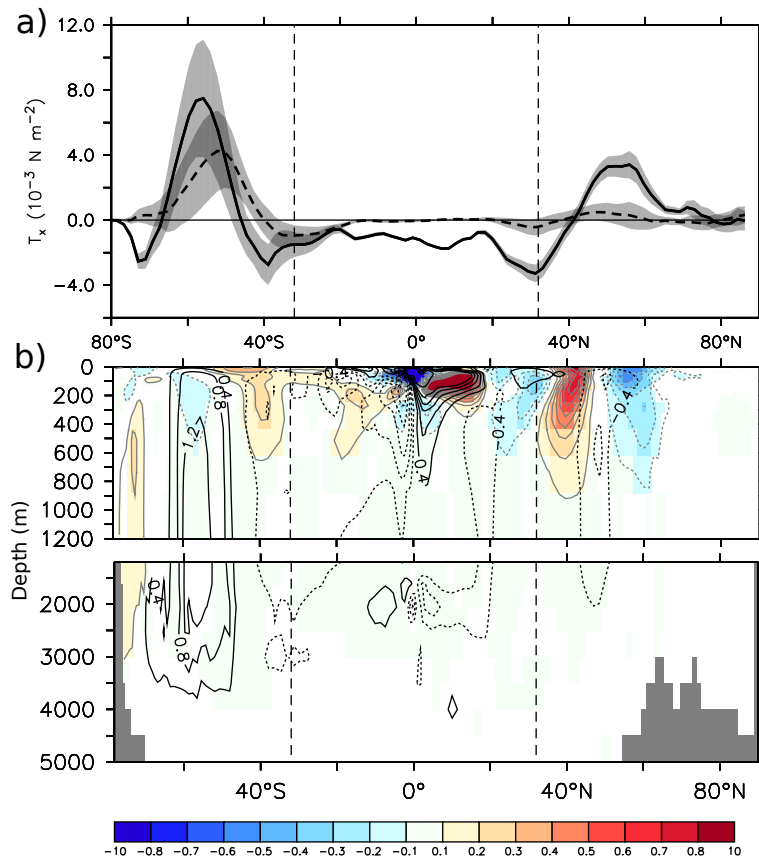
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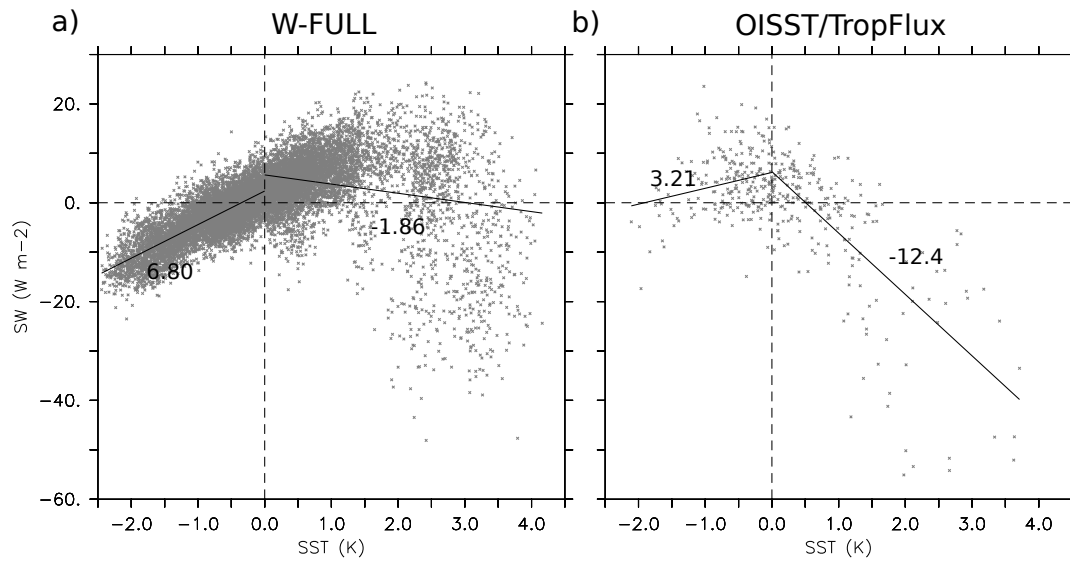
Fig. 11. (a) W-FULL minus W-CLIM 1998–2012 average SLP (hPa, shading); Ekman pumping (blue contours for positive and red contours for negative values; contour interval $1.5 \cdot 10^{-7} \text{ m s}^{-1}$); and wind stress (Pa, vectors). Wind stress vectors are grey inside the tropical Pacific (i.e. 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 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}).



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1221 **Figure 12.** Zonal-mean W-FULL minus W-CLIM: **(a)** 1998-2012 zonal wind stress (black, in
 1222 10^{-3} Pa). The black dashed line indicates the 1998-2012 anomalous zonal wind stress (relative
 1223 to 1979-2014) in W-CLIM (10^{-3} Pa). Shading indicates the 95% confidence intervals. **(b)**
 1224 Global zonal-mean potential temperature change over 1998-2012 ($^{\circ}\text{C}$, shading) and
 1225 anomalous zonal-mean residual streamfunction (Sv, contours). Note the different vertical
 1226 scales for the top 1200 m and below. In (b), colors are plotted only where significant above
 1227 the 95% confidence level.

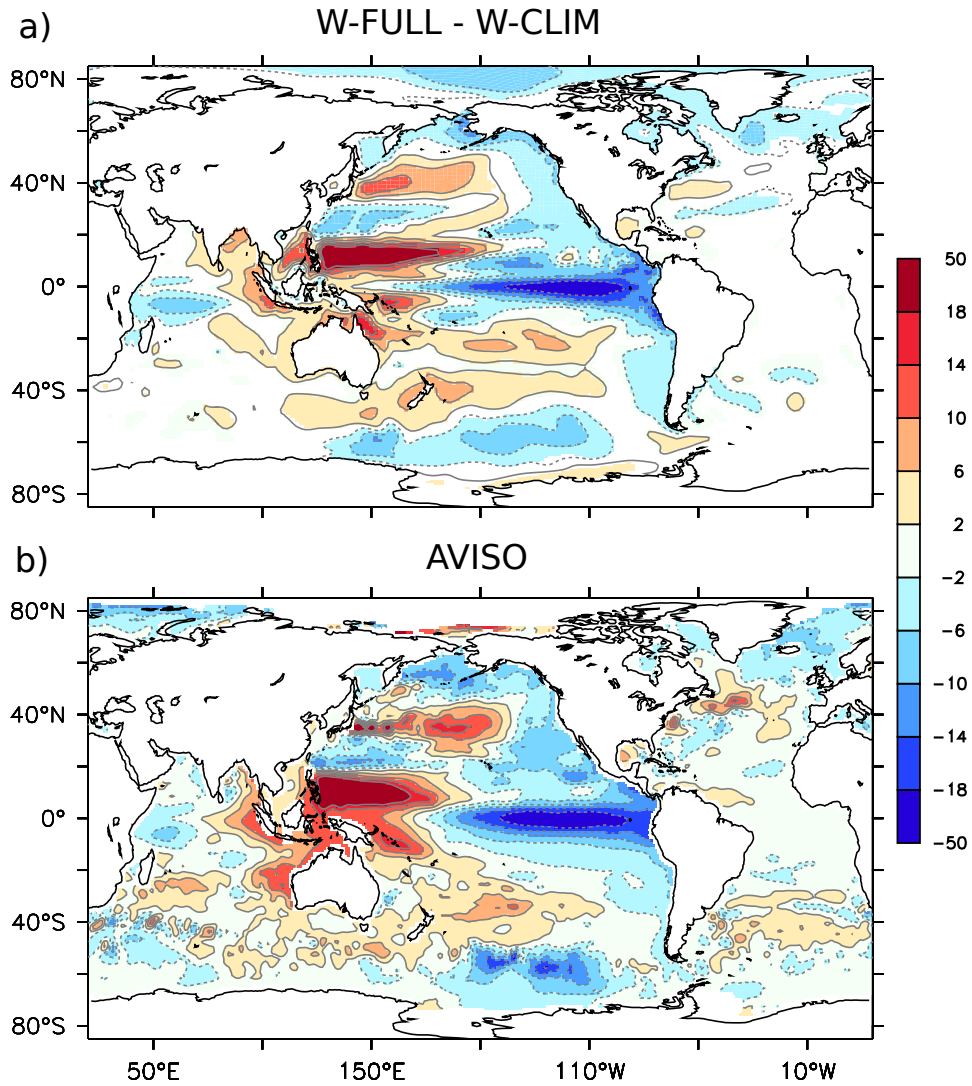
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1232 **Figure 13.** Scatter plot of average Niño 3 region (150°W - 90°W , 5°S - 5°N) surface net
 1233 shortwave radiation (positive downward, in W m^{-2}) versus SST (in $^{\circ}\text{C}$) monthly anomalies in
 1234 **(a)** the W-FULL simulation, and **(b)** observations (TropFlux and OISST). The lines indicate
 1235 the regression slopes of surface shortwave radiation onto SST, computed separately for
 1236 positive and negative SST anomalies. The values of the slopes (in $\text{W m}^{-2} \text{K}^{-1}$) are also
 1237 indicated.

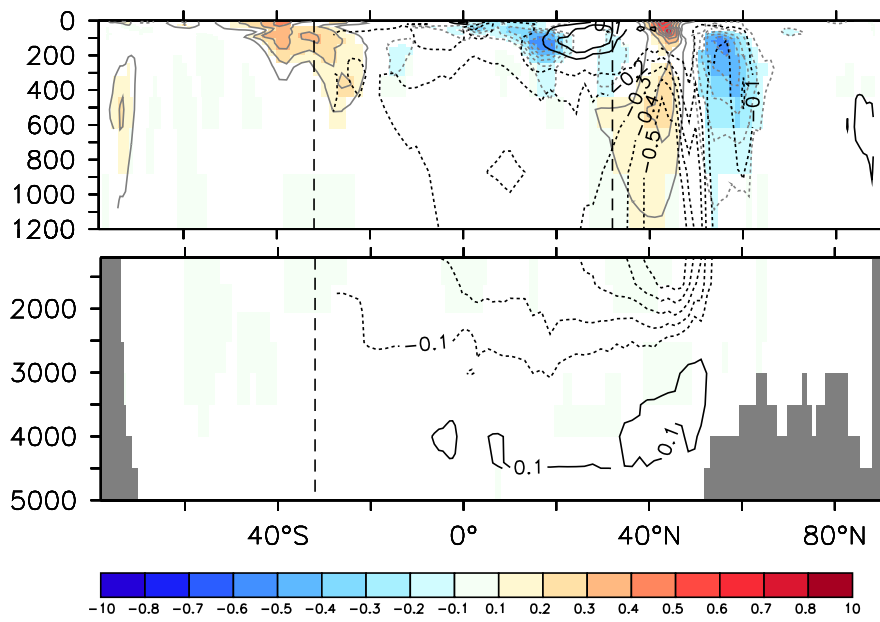
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Figure 14. Sea surface height change from 1998 to 2012 (in cm) for (a) W-FULL minus W-CLIM; (b) AVISO observations. In (a) and (b), we remove the global mean sea surface height increase: this removes the cryosphere-related and land storage, but also 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.