



1 **Surging of global surface temperature due to decadal legacy of ocean heat uptake**

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Abstract

Global surface warming since 1850 consisted of a series of slowdowns (hiatus) followed by surges. Knowledge of a mechanism to explain how this occurs would aid development and testing of interannual to decadal climate forecasts. In this paper a global climate model is forced to adopt an ocean state corresponding to a hiatus (with negative Interdecadal Pacific Oscillation, IPO, and other surface features typical of a hiatus) by artificially increasing the background diffusivity for a decade before restoring it to its normal value and allowing the model to evolve freely. This causes the model to develop a decadal surge which overshoots equilibrium (resulting in a positive IPO state) leaving behind a modified, warmer climate for decades. Water mass transformation diagnostics indicate that the heat budget of the tropical Pacific is a balance between large opposite signed terms: surface heating/cooling due to air-sea heat flux is balanced by vertical mixing and ocean heat transport divergence. During the artificial hiatus, excess heat becomes trapped just above the thermocline and there is a weak vertical thermal gradient (due to the high artificial background mixing). When the hiatus is terminated, by returning the background diffusivity to normal, the thermal gradient strengthens to pre-hiatus values so that the mixing (diffusivity x thermal gradient) remains roughly constant. However, since the base layer just above the thermocline remains anomalously warm this implies a warming of the entire water column above the trapped heat which results in a surge followed by a prolonged period of elevated surface temperatures.

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36 **1. Introduction**

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38 The estimated 0.85K global surface warming since 1850 (IPCC, 2013) was realized as
39 a succession of periods of stronger followed by weaker warming (Risbey 2015).
40 Understanding the causes of these hiatus and surge events is of key importance for
41 interannual-to-decadal prediction (Guemas et al. 2013; Meehl et al. 2014; Sévellec and
42 Drijfhout 2018) and for constraining uncertainty in long-term climate projections.

43 A large uncertainty remains concerning the detailed processes responsible for surge
44 and hiatus periods, with debate centered around the relative roles of external forcing
45 (anthropogenic greenhouse gas and aerosol emission variations, solar radiation changes,
46 volcanic eruptions) versus internal variability (Smith et al. 2015) and the dominance or
47 otherwise of the Pacific over the Atlantic (England et al. 2014; Chen and Tung 2014;
48 Drijfhout et al. 2014; Meehl et al. 2011).

49 During the recent global surface warming hiatus of the 2000s Smith et al. (2015)
50 estimated a reduction in net top-of-atmosphere (TOA) radiation of $-0.31 \pm 0.21 \text{ W m}^{-2}$ between
51 1999 and 2005, but could not identify whether this was due to internal processes or external
52 drivers. Even if the source of the TOA reduction were disentangled, Song et al. (2014)
53 demonstrate using the CMIP5 multimodel ensemble that on decadal timescales TOA
54 imbalance is only very weakly related to surface air temperature (SAT), because natural
55 variability is superimposed on global warming (see also Xie et al. 2016). Also, Hedemann et
56 al. (2017) point out that it is the difference between TOA imbalance and ocean heat transfer
57 between the surface ($\sim 100 \text{ m}$) layer and the deep ocean which matters for a hiatus, and this
58 could be too small ($\sim 0.08 \text{ W m}^{-2}$) to resolve, with the current observing system.

59 Leaving aside the ultimate causes of hiatus and surge events (external forcing versus internal
60 variability), the hiatus of the 2000s coincided with a negative phase of the Interdecadal
61 Pacific Oscillation (IPO) consistent with the results of Meehl et al (2013) who found the IPO
62 as one of three main processes responsible for hiatus and surge behaviour in the CCSM4
63 coupled climate model, the others being Antarctic Bottom Water Formation and the Atlantic
64 Meridional Overturning Circulation (AMOC). Observations demonstrated that the hiatus was
65 associated with anomalously cold Pacific sea-surface temperatures (SSTs) and warm
66 thermocline anomalies (Nieves et al. 2015), due to strengthened subtropical circulation cells
67 forced by strengthened trade winds (Farneti et al. 2014; England et al. 2014). That the IPO,
68 especially the equatorial Pacific, played a key role in the hiatus was demonstrated by England
69 et al. (2014), Watanabe et al. (2014), and Kosaka and Xie (2014) using pacemaker
70 experiments where either Tropical Pacific winds or SST were forced to follow observations in
71 coupled model simulations, which then reproduced the hiatus. However increased ocean heat
72 uptake was also seen in the deep North Atlantic and Southern Oceans during this period
73 (Balmaseda et al. 2011; Chen and Tung 2014; Drijfhout et al. 2014) and the role of these
74 regions, especially the North Atlantic where the AMOC is strongly associated with heat
75 content changes (Moat et al., 2018; Williams et al. 2015) and the Surface Atmospheric
76 Temperature (SAT) is highly variable (Sévellec et al. 2016), remains unclear.

77 More recent work has documented the long residence time of heat taken up by the
78 ocean and stored below the mixed layer during the recent hiatus of the 2000s. Maher et al.
79 (2018), using an ocean model forced by negative IPO-like surface atmospheric conditions,
80 found drawdown of heat into the main thermocline (~300-m depth) and surface cooling in the
81 Pacific. The extra subsurface heat was advected by subtropical cells, strengthened as a result
82 of enhanced trade winds, into the Indian Ocean where it remained buried in the thermocline,
83 as seen earlier in analysis of hiatus periods in the CMIP5 multimodel ensemble (Lee et al.

84 2015). A similar result was found in the pacemaker experiments of Gastineau et al. (2019),
85 although the spatial redistribution of heat showed differences, for example much of the
86 additional heat entering the Indian Ocean from the Pacific during the hiatus was returned to
87 the atmosphere via surface fluxes, or advected southwards into the Southern Ocean. The
88 spreading of this additional heat from the Pacific to the Indian Ocean, seen in observations as
89 well as models (Nieves et al. 2015; Liu et al. 2016; Lee et al., 2015) and its partial
90 reemergence after the hiatus suggests a substantial legacy of, and potentially an active role
91 for, the ocean heat uptake during the hiatus period. This was illustrated in the Maher et al.
92 (2018) study by the response to a subsequent reversal of IPO phase, in which the sequestered
93 heat was not re-extracted back to the atmosphere.

94 The consensus then, is that the IPO determines the pattern of heat drawdown above
95 the tropical thermocline and subsequent horizontal redistribution via associated changes in the
96 strength of the shallow subtropical overturning cells.

97 The significance of the increased heat uptake seen in the deep Southern and North Atlantic
98 Oceans (Meehl et al. 2013; Drijfhout et al. 2014; Chen and Tung, 2014) remains ambiguous.
99 Indeed, Drijfhout et al. (2014) suggest that the enhanced Southern Ocean heat uptake may be
100 an ongoing multidecadal process linked to external forcing of the Southern Annular Mode
101 rather than natural variability. The role of the North Atlantic remains a thorny issue with the
102 mechanism relating AMOC changes to ocean heat uptake still debated (Chen and Tung, 2014
103 versus Drijfhout et al. 2014).

104 The central role of ocean heat uptake in variations of global mean surface temperature
105 and the TOA radiation balance motivates us to explore the effects of its variability on decadal
106 timescales. Tanaka et al. (2012) show that changes in ocean mixing can affect Pacific decadal
107 variability, a key player in the hiatus of the early 2000s (Meehl et al. 2011; England et al.
108 2014; and others, see above). Hence here we explore the effects of transient enhancement of

109 ocean mixing, and the subsequent climate adjustment, on ocean heat uptake and global mean
110 surface temperature.

111

112 **2. Method**

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114 **2.1 Climate model and experiment design**

115

116 We use the HadCM3 climate model (Gordon et al. 2000 which maintains a stable
117 climate simulation without flux adjustments despite relatively coarse grid resolution ($1.25^\circ \times$
118 1.25° ocean/ $3.75^\circ \times 2.5^\circ$ atmosphere). The HadCM3 preindustrial control simulation
119 provides a reasonable simulation of the observed energy flows between components of the
120 Earth System (Table 1). We spin experiments off the control by instantaneously doubling the
121 background vertical diffusivity at every gridpoint and integrating the model for one decade
122 before instantaneously returning the diffusivity to its control value and continuing the
123 integration for one or more further decades. To address internal variability we create a 4-
124 member ensemble by starting from different points of a 140-year portion of the 2000-year
125 control simulation. All ensemble members begin on the same day of the year, December 1.

126

127 The *in situ* vertical diffusivity for tracers consists of a time-independent (background)
128 part, which only varies with depth and a time-dependent part which depends on the
129 stratification and velocity shear present and varies spatially and temporally (Gordon et al.
130 2000, Appendix 2). The background part increases with depth from $\sim 10^{-5} \text{ m}^2 \text{ s}^{-1}$ at the surface
131 to $\sim 15 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ in the abyss (Gordon et al. 2000). We instantaneously double this
132 background part at the beginning of the ensemble experiment and set it back to normal at the
133 beginning of the second decade. Ocean models also include substantial numerical mixing up

134 to several times the explicit mixing (Megann 2018), hence the mixing perturbation we apply
135 only modifies the actual *in situ* mixing coefficient by perhaps 10-20%, not far removed from
136 its observational uncertainty.

137

138 Ensemble mean quantities are evaluated for significance against the control (all
139 variables are first filtered using a 5-year running mean unless otherwise stated) using a one
140 tailed *t*-test, taking into account the number, *n*, of independent degrees of freedom in the 140
141 year control simulation (estimated to be $n=26$ due to autocorrelation in the time series).

142

143 **2.2 Watermass transformation framework**

144

145 Since we modify the diffusion coefficient and hence the mixing in our experiments, a
146 natural framework for diagnosis of the results is water mass transformation analysis (Walin
147 1982). The strength of this framework is that it diagnoses mixing from all sources, including
148 numerical mixing, which cannot easily be quantified by alternative methods (Megann 2018;
149 Lee et al. 2002). Additionally, the water mass transformation framework can be applied
150 directly to observations and was developed in that context (Walin 1982).

151 For convenience of exposition, we describe this framework for a zonally integrated
152 domain, however the extension to three dimensions is important for the real ocean, hence we
153 also discuss how to interpret the equations in a 3D context.

154 Referring to Fig 1(a), we begin with the ‘transformation’ i.e. the rate $G(\theta, \phi)$ at which
155 water crosses some isotherm with potential temperature θ integrated zonally over all
156 longitudes, λ , and meridionally north of some latitude ϕ . This is related by mass (or volume
157 in a Boussinesq fluid) conservation to the meridional influx across ϕ ,

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159
$$\psi(\theta, \phi) = \iint_{\lambda, z: \theta'(\lambda, \phi, z) < \theta} v R \cos \phi \, dz d\lambda, \quad (1)$$

160 of water colder than θ (where v is meridional velocity, and R is the radius of the Earth)
 161 and the rate of change of the volume $V(\theta, \phi)$ north of ϕ and colder than θ by

162
$$\frac{\partial V(\theta, \phi)}{\partial t} = \psi(\theta, \phi) - G(\theta, \phi) \quad (2)$$

163 This transformation $G(\theta, \phi)$ represents water crossing the isotherm from cool to
 164 warm, and thus requires heat input; to link $G(\theta, \phi)$ to heat inputs we consider (Fig. 1b) the
 165 (Boussinesq) equation for the heat content of water colder than θ (e.g. Nurser et al., 1999):

166
$$\frac{\partial \mathbb{H}}{\partial t} = -D_{\text{diff}} - D_{\text{surf}} + A - c_p \rho_0 G \theta. \quad (3)$$

167 Here

168
$$\mathbb{H} = \rho_0 c_p \iiint_{\lambda, \phi', z: \theta'(\lambda, \phi', z) < \theta, \phi' < \phi} \theta' R \cos \phi' \, d\lambda \, d\phi \, dz \quad (4)$$

169 is the heat contained within seawater with potential temperature less than or equal to θ
 170 integrated everywhere north of a given latitude ϕ and $\frac{\partial \mathbb{H}}{\partial t}$ is its time derivative. D_{diff} is the
 171 (cold-to-warm) diffusive temperature flux across the θ isotherm north of latitude ϕ due to
 172 mixing, while

173
$$D_{\text{surf}} = - \iint_{\lambda, \phi: \theta_s(\lambda, \phi) \leq \theta} I(\lambda, \theta, z_\theta) R \cos \phi \, d\lambda \, d\phi$$

 174
$$- \iint_{\lambda, \phi: \theta_s(\lambda, \phi) \leq \theta} (I_s - Q_{\text{net}}(\lambda, \phi)) R \cos \phi \, d\lambda \, d\phi \quad (5)$$

175 represents the heat fluxes upwards and out (first term) through the θ isotherm due to solar
 176 irradiance I (sign convention is +ve downwards) where the isotherm is close to the ocean
 177 surface and upwards through the surface boundary where the θ isotherm has outcropped
 178 (second term; the net turbulent air-sea flux $Q_{\text{NET}} = Q_{\text{latent}} + Q_{\text{sensible}} + Q_{\text{LW}}^{\text{net}}$ is made up of

179 (+ve upwards) latent, sensible and net longwave (LW) radiative heat fluxes and the surface
 180 irradiance I_s is +ve downwards, note that, θ_s denotes the SST). The term

$$181 \quad A(\theta, \phi) = \rho_0 c_p \iint_{\lambda, z: \theta'(\lambda, \phi, z) < \theta} \theta' v R \cos \phi \, dz d\lambda \quad (6)$$

182 represents the advective heat transport of water colder than θ northwards across latitude ϕ ,
 183 while $c_p \rho_0 G \theta$ represents the advective heat flux up out across the θ isotherm.

184 Noting that the transport of water with $\theta < \theta' < \theta + \delta\theta$:

$$185 \quad \delta\psi = \iint_{\lambda, z: \theta < \theta'(\lambda, \phi, z) < \theta + \delta\theta} v R \cos \phi \, dz d\lambda \approx \frac{\partial\psi}{\partial\theta} \cdot \delta\theta,$$

186 we can re-express (integrating by parts for the last result)

$$187 \quad A(\theta, \phi) = \rho_0 c_p \int^{\theta} \frac{\partial\psi}{\partial\theta} \theta d\theta' = \rho_0 c_p \psi \theta - \rho_0 c_p \int^{\theta} \psi d\theta', \quad (7a)$$

188 and similarly obtain:

$$189 \quad \mathbb{H}(\theta, \phi) = \rho_0 c_p \int^{\theta} \frac{\partial V}{\partial\theta} \theta d\theta' = \rho_0 c_p V \theta - \rho_0 c_p \int^{\theta} V d\theta'. \quad (7b)$$

190 Eliminating \mathbb{H} and A from (3) using (7a) and (7b), and removing the $G\theta$, $\psi\theta$ and $\frac{\partial V}{\partial t} \theta$
 191 terms using (2) gives

$$192 \quad -\rho_0 c_p \int^{\theta} \frac{\partial V}{\partial t} d\theta' = -D_{\text{diff}} - D_{\text{surf}} - \rho_0 c_p \int^{\theta} \psi d\theta'.$$

193 Differentiating this by θ and again using (2) to replace $-\frac{\partial V}{\partial t} + \psi$ by G gives the key result

$$194 \quad \rho_0 c_p G = -\frac{\partial D_{\text{diff}}}{\partial\theta} - \frac{\partial D_{\text{surf}}}{\partial\theta}. \quad (8)$$

195 Multiplying (8) by $\delta\theta$ makes clear that (8) represents the heat budget of the volume
 196 between the θ and $\theta + \delta\theta$ isotherms, relating the transformation (volume flux from cold to
 197 warm) across the isothermal surfaces to the heat acquired from the convergence of air-sea
 198 fluxes and mixing. It is customary to introduce F , the transformation driven by air-sea fluxes ϕ

199
$$\rho_0 c_p F = -\frac{\partial D_{\text{surf}}}{\partial \theta}; \quad (9)$$

200 it then follows (Fig. 1c) that

201
$$\rho_0 c_p F \delta \theta = -\frac{\partial D_{\text{surf}}}{\partial \theta} \delta \theta$$

202
$$= \iint_{\lambda, \phi: \phi' > \phi, \theta_s(\lambda, \phi) > \theta + \delta \theta} I(\lambda, \phi', z_{\theta + \delta \theta}) R \cos \phi' d\lambda d\phi'$$

203
$$- \iint_{\lambda, \phi: \phi' > \phi, \theta_s(\lambda, \phi) > \theta} I(\lambda, \phi', z_{\theta}) R \cos \phi' d\lambda d\phi'$$

204
$$+ \iint_{\lambda, \phi': \phi' > \phi, \theta < \theta_s(\lambda, \phi) \leq \theta + \delta \theta} (I_s - Q_{\text{net}}(\lambda, \phi')) R \cos \phi d\lambda d\phi' \quad (10)$$

205 where the three terms on the RHS of (10) represent (i) the solar irradiation down through the
 206 $\theta + \delta \theta$ isotherm (a heat source for the layer $\theta < \theta' < \theta + \delta \theta$) (ii) the solar irradiation down
 207 through the θ isotherm (a heat sink) and (iii) the surface heat flux down through the outcrop
 208 $\theta < \theta_s < \theta + \delta \theta$. In terms of F , the watermass transformation G is then given as

209
$$\rho_0 c_p G = \rho_0 c_p F - \frac{\partial D_{\text{diff}}}{\partial \theta}. \quad (11)$$

210 While it is in principle possible to evaluate the diffusive fluxes, they are difficult to evaluate
 211 in observations and many models, so they are often in practice estimated as a residual
 212 between G , (which can be found by (1) from changes in volume $\frac{\partial V}{\partial t}$ and transport ψ) and F
 213 which is known from the surface fluxes.

214 We can use Eq (11) to eliminate G in Eq (2) arriving at

215
$$\frac{\partial V(\theta, \phi)}{\partial t} = \psi(\theta, \phi) - F(\theta, \phi) + \frac{1}{c_p \rho_0} \frac{\partial D_{\text{diff}}}{\partial \theta} \quad (12)$$

216 Now $\frac{\partial V}{\partial \theta} \delta \theta$ represents the volume of water with temperature between θ and $\theta + \delta \theta$,

217 so differentiating Eq (12) by θ gives:

218
$$\frac{\partial^2 V(\theta, \phi)}{\partial t \partial \theta} = \frac{\partial \psi(\theta, \phi)}{\partial \theta} - \frac{\partial F(\theta, \phi)}{\partial \theta} + \frac{1}{c_p \rho_0} \frac{\partial^2 D_{\text{diff}}}{\partial \theta^2} \quad (13)$$

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which describes the increase in volume of water of temperature θ in terms of its southern influx $\frac{\partial \psi(\theta, \phi)}{\partial \theta} \delta \theta$, and the convergence of flow across the θ and $\theta + \delta \theta$ isotherms.

To bring out the physical meaning in three dimensions of the terms in Eqs (12) and (13), Fig. 1(d) shows simulated temperature in one particular December in a region of the Pacific dominated by the shallow northern subtropical overturning cell (19.5°N-30°N, 140°E-120°W and 0- 300m depth). The Equatorial warm pool is visible to the south west ($\theta > 27^\circ\text{C}$). From here, isotherms slope up, east- and northwards. Water between 22°C and 23°C (purple shading outcrops at the surface in a crescent oriented northwest to south east. In 3D, the watermass resembles a bowl shape truncated by the southern boundary of the region where the watermass persists as a thin near-horizontal feature at about 80m depth at ,160°E extending eastwards to about 160°W where it rises vertically to intersect the surface.

Applying Eq (12), we integrate the surface heat flux across all surface waters to the north with $23^\circ\text{C} < \theta \leq 24^\circ\text{C}$ (purple shaded surface region extending to the North Pole). This is the larger yellow arrow ($F_{23^\circ\text{C}, 19.5^\circ\text{N}}$), a loss of volume, which would be compensated in steady state by flow of warmer water across the 23°C isotherm (solid black arrows). With the heat flux integrated back to the North Pole, we cannot say anything about the latitudinal distribution of this diapycnal flow. Now, restricting ourselves to waters to the north with $22^\circ\text{C} < \theta \leq 23^\circ\text{C}$ we obtain a smaller value (smaller yellow arrow = $F_{22^\circ\text{C}, 19.5^\circ\text{N}}$) which must be balanced by a diapycnal flow of water across the 22°C isotherm (dashed black arrows). Hence the *difference*, i.e. applying Eq (13), between these two transformation rates (known as the formation rate) results in a tendency for the volume of the purple region to increase.

242 The formation rate may be partially balanced by ocean circulation/overturning (green
 243 arrows). The upper (lower) green arrow represents the volume flux out of the domain for
 244 water with $\theta \leq 23^\circ\text{C}$ ($\theta \leq 22^\circ\text{C}$). Hence, the volume flux out of the purple region, equals the
 245 difference between the green arrows. Both arrows are southwards because in this part of the
 246 ocean the circulation consists of a strong northwards flow in a thin layer at the surface (higher
 247 θ) and a weaker southwards flow in a thicker, deeper layer (lower θ). As we integrate up from
 248 the lowest temperatures, the streamfunction itself is negative, but its vertical gradient is
 249 negative at depth/low θ and positive at the surface/high θ .

250 Similar reasoning leads us to conclude that changes to the volume (mixing) in the
 251 purple region are given by the difference in $\frac{\partial V}{\partial t} \left(\frac{\partial D_{diff}}{\partial \theta} \right)$ with respect to temperature.

252 To localise these quantities spatially, consider two nearby latitudes. Fig. 1(e) is
 253 identical to 1(d) except an additional slab of ocean, to 18°N is added to the south, and the
 254 yellow arrows represent the surface heat flux integrated from the North Pole down to the new
 255 southern boundary ($F_{23^\circ\text{C}, 18^\circ\text{N}}$, $F_{22^\circ\text{C}, 18^\circ\text{N}}$). Evidently, the diapynal fluxes in the region between
 256 18°N and 19.5°N (hollow arrows) are given by the difference between values of F at the two
 257 different latitudes. Similar considerations apply to the overturning (green arrows) and to the
 258 time derivative and mixing terms.

259

260 Note the resemblance of Eq (12) to the equation for the evolution of ocean
 261 temperature (neglecting horizontal diffusion):

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$$263 \quad \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(K_v \frac{\partial T}{\partial z} \right) + \frac{I}{\rho_0 c_P} - \nabla \cdot (\mathbf{u}T) . \quad (14)$$

264

265 Here K_v is the diffusion coefficient, \mathbf{u} the three-dimensional ocean velocity and I is the
266 solar irradiance encountered earlier. The correspondence between the terms in Eqs (12) and
267 (14) becomes more obvious when we consider the surface boundary condition

268

$$269 \quad K_v \left. \frac{\partial \theta}{\partial z} \right|_{z=0} = \frac{Q_{NET} - I_s}{\rho_0 c_P}. \quad (15)$$

270

271 where Q_{NET} , the net surface heat flux and I_s , the surface irradiance were introduced
272 earlier. Recall that K_v consists of a background value, which increases with depth and a time-
273 varying part, which depends on stratification and velocity shear.

274

275 In the surface mixed layer K_v becomes large whilst $\partial \theta / \partial z$ is small and the surface
276 flux, $Q_{NET} - I_s$, is effectively spread over a vertical distance of the order of the mixed layer
277 depth. We shall use this property to develop a conceptual model in Section 3.6.

278

279 **3. Results**

280

281 **3.1 Ocean temperature response**

282

283 The global temperature response in the perturbed ensemble experiment is shown in
284 Fig. 2, a time-depth plot of horizontal averaged ensemble mean ocean temperature anomaly
285 relative to the mean of the 140-year control. In the first decade after the diffusivity is
286 increased, near-surface ocean temperatures (0-100m) decrease by $\sim 0.15\text{K}$, reaching a
287 minimum at year 8 (2 years before the diffusivity is returned to its original value – this is seen
288 in the unfiltered data (not shown) and is not an artefact of filtering). Simultaneously, the
289 subsurface ocean (below 100m) warms, with maximum warming at $\sim 200\text{m}$ around year 9.

290 Thus, the change in mixing results in transfer of heat from the surface layer to the
291 thermocline, reminiscent of the observed hiatus of the 2000s. When the diffusivity is returned
292 to its original value at the beginning of year 11, the surface ocean warms, with peak surface
293 temperature occurring at year 15. In the subsurface ocean, there is slight cooling in the 200-
294 300 m depth interval, but below this the temperature remains roughly constant for the
295 remaining 10 years of the experiment. Hereinafter, we refer to years 1-10 of the ensemble
296 experiment as the hiatus period and years 11-20 as the surge period, although strictly this
297 should be defined in terms of anomalous global mean temperature tendency. The heat taken
298 up during the simulated hiatus period remains in the ocean for at least a decade after the
299 diffusivity has been restored to its original value. Although anomalously cool during the
300 hiatus, global mean SST becomes anomalously warm over the surge period.

301

302 Ensemble mean zonal average ocean temperature anomalies during years 6-10 (Fig.
303 3(a)) indicate cooling in the tropics-subtropics (30°S - 30°N) from the surface to about 100m
304 and warming in the permanent thermocline throughout the tropics and midlatitudes, but
305 particularly marked in the tropics (20°S - 20°N , 100-400m). This pattern of surface cooling and
306 deep warming is strongly reminiscent of the observed hiatus of the 2000s (e.g. Drijfhout et al.
307 2014).

308

309 During years 11-15 there is warming throughout the tropics and midlatitudes down to
310 1000m depth (Fig. 3(b)). However, weak cooling of the high latitudes present during years 1-
311 10 remains, particularly in the northern hemisphere to depths of 800m, but is counterbalanced
312 by a slight warming from 800-2500m (not shown). Once again, the heat taken up by the ocean
313 into the main thermocline during the hiatus period persists into the surge period.

314

315 Fig. 4(a) shows the ensemble mean SST anomaly averaged over years 6-10. Surface
316 cooling is particularly marked in the tropics. The western boundary current regions by
317 contrast, exhibit warming, albeit over a smaller area. A negative IPO-like pattern is apparent
318 with a wedge shaped region of cooling in the tropical Pacific, and warming in the western
319 north Pacific and the southeastern Pacific.

320

321 Following restoration of the diffusivity to its original value, the surface warms, with
322 Pacific SST showing a strongly positive IPO pattern averaged over years 11-15 (Fig. 4(b)).
323 Marked warming remains over the western boundary currents and their extensions in the
324 northern hemisphere, and the Indian ocean shows substantial surface warming. The Southern
325 Ocean, south western Pacific and south Atlantic show a weak cooling. Areas of statistically
326 significant warming during the surge are less extensive than areas of cooling during the
327 hiatus. However a clear global-scale pattern is evident.

328

329 The SST anomaly patterns can be compared to the 0-1000m ocean heat content
330 anomaly. During years 6-10 there are substantial rises in ocean heat content in the tropics and
331 subtropics (40°S - 40°N) which remain largely in place in years 11-15, that is, the uptake of
332 heat during the hiatus is not reversed during the surge (Fig. 4(c), (d)). Unlike the hiatus, the
333 surge SST anomaly pattern in years 11-15 echoes that of the ocean heat content anomaly over
334 much of the ocean: in the tropical and south Pacific, the Indian and the North Atlantic Oceans,
335 but not in the South Atlantic, the North Pacific, or the south west Pacific.

336

337 Turning to the ocean surface heat uptake anomaly (annual mean anomaly of net
338 surface heat flux into the ocean, Fig. 4(e), (f)), there are large increases in years 6-10
339 throughout the tropics and subtropics of all ocean basins, and in some subpolar regions. A

340 zonally banded structure is seen, seemingly related to oceanic fronts and mode water
341 formation regions with strong anomalies over the ITCZ and the western boundary currents
342 (particularly the Gulf Stream) and adjacent to Antarctica. When the diffusivity is returned to
343 its standard value in years 11-15, the ocean heat uptake anomaly reduces substantially across
344 the globe, but remains positive in many regions including the tropical Pacific and over the
345 Gulf Stream. The North Atlantic subpolar gyre by contrast shows a strong reduction in ocean
346 heat uptake during the surge period.

347

348 **3.2 Energy budget**

349

350 Fig. 5 indicates the impact of the perturbation on Earth's energy flows (Wild et al.
351 2013). There is no effect on incident solar radiation at the TOA hence this is not plotted. Fig.
352 5(a) shows the impact on the remaining shortwave (SW) flows. The reflected SW radiation at
353 the TOA (black) is initially reduced by 0.25 W m^{-2} (a 5-year filter is applied to the data), and
354 then gradually returns to its control value over the next 20 years. The atmospheric absorption
355 (red) is also initially reduced, but returns to the control value earlier, after about 12 years, and
356 overshoots slightly in the surge period. The reduction in both reflection and absorption by the
357 atmosphere means that the incident surface SW radiation (green, cyan) is increased by about
358 0.6 W m^{-2} , the change in surface reflection (blue) being negligible. The incident surface SW
359 radiation returns to its unperturbed value after about 12 years (2 years after the diffusivity
360 returned to normal).

361

362 The impact on non-solar energy flows is depicted in Fig. 5(b). Outgoing LW radiation
363 at the TOA (black) displays a symmetrical pattern with a reduction in the first 10 years
364 (consistent with reduced SST) and an increase in years 11-20 consistent with temperatures

365 warmer than the control. The biggest changes occur in the surface upwelling (red) and
366 downwelling (green) LW, which decrease by about 1 W m^{-2} relative to the control during the
367 hiatus, but increase rapidly by 2 W m^{-2} once the diffusivity returns to its control value. Thus
368 unlike the shortwave, the LW terms strongly overshoot their original values in the surge
369 regime. Of the turbulent heat flux terms, sensible heat (blue) shows negligible variation,
370 whilst latent heat (cyan) bears a strong resemblance to atmospheric absorption (weaker latent
371 heat loss and reduced SW absorption during the hiatus period may be linked via reduced
372 atmospheric humidity). Both the latter terms remain slightly higher than their control values
373 in the last 10 years of the perturbed simulations.

374

375 Global mean ocean heat uptake, net TOA heat flux and global mean heat content
376 tendency show very similar variation as required by energy conservation, since most of the
377 heat capacity resides in the ocean, hence only the ocean heat uptake is illustrated (Fig. 6). The
378 control displays large decadal fluctuations superimposed on a global net heat loss (TOA
379 imbalance) of about -0.2 W m^{-2} (vertical black line), indicative of long-term drift. This is
380 associated with the model Antarctic Bottom Water (see Fig. 10(a)) in the deep Southern
381 Ocean and is unlikely to affect the signals in the top 1000m seen in Figs. 2–4.

382

383 The diffusivity perturbation causes substantially increased ocean heat uptake of about
384 $+0.6 \text{ W m}^{-2}$ relative to the control. This elevated heat uptake declines steeply between years 8
385 and 12, and becomes indistinguishable from the control. The ocean heat uptake thus displays
386 an asymmetry: heat gained during the hiatus is not fully discharged from the ocean in the
387 subsequent surge (within the timescale of our experiments).

388

389 **3.3 Ocean and atmosphere surface temperature response**

390

391 The ensemble mean global average SST is plotted in Fig. 7 (red) together with the
392 global average SST from the control (vertical black line) and one ensemble member (green)
393 which was integrated for an extended period after the diffusivity was restored to its original
394 value. The mean SST over the 140-year portion of the control was 17.95 K. In response to the
395 reduced diffusivity, the ensemble mean global average SST reduces by about 0.15 K to
396 17.80°C. When the diffusivity is returned to its original value at the beginning of year 11,
397 global average SST increases steeply by ~0.25 K in 2-3 years, although strictly speaking, the
398 hiatus ended and the surge began when global temperature stopped declining and began to
399 rise around year 8 (this is seen in the unfiltered data (green) and is not an artefact of time
400 filtering). Subsequently the ensemble mean SST stabilizes at about 18.05 °C. The green curve
401 indicates the evolution of a single ensemble member rather than the ensemble mean, which
402 was integrated for a much longer period than the others. The surge in temperatures in years 8-
403 12 leaves the global mean SST higher than the control, possibly for several decades or more.
404 During the surge and hiatus periods the ensemble mean SST is significantly (95% confidence)
405 below/above the control, but the ensemble spread remains similar to the temporal variability
406 of the control. In contrast in the transition period (years 10-12), the SST anomaly is not
407 significantly different from the control, but the ensemble spread is smaller than the control
408 variability. This suggests a potentially high predictability of this transition phase.

409

410 Fig. 8 provides a scatter plot of annual mean SAT versus SST for the control (red
411 circles) and the individual ensemble members for the first ten years of each simulation (black)
412 and the last ten years (blue). SAT correlates very well with SST so a hiatus in SST is
413 equivalent to one in SAT on an annual mean and global average basis. The same linear
414 relationship between SST and SAT seen in the control holds for both surge and hiatus

415 periods. Spatial and seasonal differences in this relationship are of interest, but are not
416 pursued further in this paper.

417

418 **3.4 Analysis in temperature space**

419

420 A consistent finding is that following the relaxation of conditions that force a hiatus,
421 the surface warms rapidly and remains warmer than the previous long-term average state. We
422 now investigate this behavior using water mass transformation analysis.

423

424 *(i) Watermass transformation*

425

426 We first apply the water mass transformation framework, Eqs. (12), (13), to the
427 control simulation. The net trend $\partial V/\partial t$, Fig. 9(a), is very small in most temperature classes
428 and latitudes, but $\partial^2 V/\partial t \partial \theta$ is negative between 0-2°C (80°S-20°N) representing a loss of
429 volume. This is counterbalanced by positive $\partial^2 V/\partial t \partial \theta$ at the same latitude range, between -
430 2-0°C representing a gain in volume. Thus there is cooling in the deep Southern Ocean
431 (~0°C) consistent with the overall negative TOA balance.

432

433 From Eq (12) surface fluxes, ocean circulation and mixing compete to create a net
434 trend in the volume of seawater in temperature classes. In Fig. 9(b)-(d) each panel shows one
435 term from Eq (12). Values plotted for each latitude, ϕ , and temperature, θ , represent all the
436 fluid with temperature $\leq \theta$ and north of ϕ . Therefore, as with streamfunction plots in depth-
437 latitude space, a convergence/divergence of isolines represents accumulation/loss of volume
438 at a particular θ , ϕ . Along isolines, input and output cancel – water transformed from the class
439 (and/or latitude) below is counterbalanced by the same amount of water being transferred to

440 the class (and/or latitude) above. Hence we can draw arrows connecting sources of water with
441 particular properties to sinks, or closed loops representing continuous transformation
442 pathways. The reader should remember that the 2D temperature-latitude plots in Fig. 9
443 represent complex water pathways in 3D physical space (Fig. 1)

444

445 Fig. 9(b) shows the overturning streamfunction, ψ , in temperature space (Eq. (1)). It is
446 calculated by integrating up lateral transports of fluid of various temperatures, so in principle
447 the diagnosed apparent “vertical” diathermal motion may be associated with volume trends
448 $\partial V/\partial t$ rather than actual warming or cooling. However, since the trends are weak, (Fig. 9(a))
449 the diagnosed diathermal motion should be accurate. Negative (positive) values indicate
450 clockwise (anticlockwise) transformation cells. Shallow subtropical cells (STCs), 25-30Sv in
451 strength, are apparent, centered at about 22°C, 10°S in the southern hemisphere and, slightly
452 cooler, 20°C, 10°N in the north. These deliver cold water upwelled near the equator (sharp
453 gradient between 15-25°C at 0° latitude implies transfer from cold to warm temperatures as
454 explained in Section 2.2) to the downwelling/subduction regions in the subtropics (Fig. 1
455 gives a 3D view of this process). The model North Atlantic Deep Water (NADW) cell (i.e. the
456 AMOC) carries ~25Sv of surface waters northwards into the North Atlantic and returns cold
457 (3-7°C), deep waters southwards. These upwell at mid-high southern latitudes (~40°S) and
458 rejoin the surface circulation. The Antarctic Bottom Water (AABW) cell downwells ~25Sv of
459 cold surface waters in the Southern Ocean and carries them northward as near-bottom
460 currents (at ~0°C). These rise and warm slightly at mid-high latitudes in the northern
461 hemisphere and join the NADW, making their way back southwards.

462

463 Air-sea fluxes directly transform water shifting it from one temperature class to
464 another. The associated streamfunction, F , (Eq. (10)) is analogous to ψ (Fig. 9(c)); but differs

465 in that it is defined from the (known) diathermal flux; it gives diagnosed (unrealistic)
466 horizontal transports consistent with weak volume trends and no mixing. Air-sea fluxes only
467 act on a temperature class if the bounding surface of that class reaches to the surface at the
468 latitude in question (or close enough that penetrative solar radiation passes through its
469 boundary) as illustrated (section 2.2) by the purple region of Fig 1(d). Thus there is a blank
470 region in Fig. 9(c) between 40°S and 30°N below 10°C where air-sea fluxes cause no
471 temperature change. Air-sea fluxes can make warm waters even warmer (tropical heating) and
472 cool waters even cooler (subtropical/high latitude cooling). Water in the 23°C range is
473 transformed by air-sea fluxes into warmer waters (up to 30°C) mainly in the tropics. Waters
474 in this class are also cooled by air-sea fluxes between the subtropics and mid latitudes to
475 temperatures around 15°C. Further poleward, waters around 5°C are transformed into cold
476 deep water below 0°C, largely in the Southern Ocean and the North Atlantic.

477

478 Mixing, obtained as a residual, $(\partial D_{diff}/\partial\theta)$, in Eq. (12) opposes air-sea fluxes,
479 removing extremes of temperature and adding volume to intermediate temperature classes
480 (Fig. 9(d)). However, mixing also acts within the ocean interior, where air-sea fluxes cannot
481 penetrate, for example mixing is active between the Equator and 40°S transforming water
482 from the 15°C to the 20°C range, and balances the NADW overturning cell at around 50°N
483 where waters between 5-10°C are transformed into the range 0-5°C.

484

485 *(ii) Transformation and formation rates*

486

487 We now focus on the tropics and subtropics where a large part of the surge signal
488 originates, and examine the horizontal derivative of the transformation rate. Following Eq.
489 (12), in Fig. 10(a) we display the latitudinal divergence of the volume flux across isotherms

490 (the transformation rate) at the Equator (averaged over 3°S–3°N) as a function of temperature
491 due to air–sea fluxes, mixing, overturning and volume change. This is essentially a section
492 through Figs 9(a)–(d) at the Equator, differentiated with respect to latitude. We plot curves for
493 the control simulation to elucidate the fundamental balances in the model and subsequently
494 investigate how these balances change during the hiatus and surge in the perturbed ensemble.

495

496 At the Equator, air–sea fluxes (green) cause transformation in temperature classes 17–
497 32°C (there is no direct effect on isotherms below 17°C as these do not outcrop at the
498 Equator). The lines have a negative gradient (formation rate) between 26–31°C, hence air–sea
499 fluxes tend to generate water masses in this temperature range (F has a negative sign in eq.
500 (12)). Between 17–25°C, the lines have a positive gradient and air–sea fluxes remove water in
501 these classes. The overturning (red) acts to remove (add) water between 23–30°C (10–23°C).
502 Mixing (blue) differs from the other terms as the associated formation rate has three regimes:
503 between 28–32°C mixing removes water; between 21–28°C it adds water; and between 10–
504 21°C it removes water. This is consistent with mixing tending to eliminate extremes. The
505 result is that at the Equator there are four thermal regimes. In each, two out of the three terms
506 (in formation rate) are in approximate balance and the third is relatively unimportant. Thus
507 between 28–32°C the predominant balance is mixing versus air–sea fluxes (Equatorial regime
508 iv); between 25–28°C mixing versus overturning (regime iii); between 21–25°C mixing versus
509 air–sea fluxes (regime ii); and between 10–21°C mixing versus overturning (regime i). Below
510 15°C there is little coherent signal. In all four regimes, mixing is one of the dominant terms.
511 Volume change is a very small term over most temperature classes, averaged over the 140-
512 year control simulation.

513

514 Fig. 11(b) shows transformation rates for the off-Equatorial regions (average over 18–
515 12°S and 12–18°N – the strongest subsurface anomalies in Fig. 4 are between 20°S-20°N).
516 Transformation rates are much smaller compared to the Equator: 1–2Sv/°latitude compared to
517 8–10Sv/°latitude and the roles of the individual processes differ. Air–sea fluxes add volume
518 between 28-32°C, and between 20-25°C, and remove volume between 25-28°C. The
519 overturning behaves in an opposite manner than at the Equator, adding volume between 25-
520 28°C and removing it between 20–25°C. Mixing removes volume between 28-32°C (off-
521 Equatorial regime iv) and adds it between 20-25°C (regime iii). Between 10°C-20°C,
522 overturning and mixing balance, mixing adding volume and overturning removing it (regime
523 ii). The situation is reversed between 5°C-10°C with overturning adding volume and mixing
524 removing it (regime i). Details vary between the northern and southern off-Equatorial regions
525 (not shown), but the regimes are similar. Again, net volume change is a very small term.
526 Although the thermal regimes are different from those at the Equator, the boundaries between
527 the regimes occur at about the same temperatures.

528
529 In both Equatorial and off-Equatorial regions, the corresponding formation rate curves
530 for the perturbed ensemble (not shown) are similar in shape to the control and the boundaries
531 between the thermal regimes occur at the same temperatures.

532
533 We further characterize the thermal regimes by calculating the formation rate terms
534 (Section 2.2 and Fig. 1) for the control simulation. Fig. 11(a) shows the net volume flux
535 convergence for each regime i–iv, associated with overturning (red), air–sea fluxes (green)
536 and mixing (blue) in the Equatorial region. The net volume inflation in each of these regimes
537 is very small, and there is a near-exact balance between the three terms. In regime iv, air–sea
538 fluxes create about 6Sv/°latitude of water (36Sv in total for the 3°S–3°N region) whilst

539 mixing and overturning remove $4\text{Sv}/^\circ\text{latitude}$ and $2\text{Sv}/^\circ\text{latitude}$ respectively. Mixing delivers
540 $3\text{Sv}/^\circ\text{latitude}$ of water into regime iii, largely balanced by overturning. Regime ii is
541 effectively a reversed version of regime iv (loss of $\sim 6\text{Sv}/^\circ\text{latitude}$ by air–sea fluxes, gain of
542 $4\text{Sv}/^\circ\text{latitude}$ and $2\text{Sv}/^\circ\text{latitude}$ by mixing and overturning respectively). Similarly regime i is
543 a reversed version of regime iii. This suggests that regimes i and iii and ii and iv are causally
544 connected. The situation is subtly different in the off-Equatorial regions (Fig 11b) where
545 regimes iii and iv both have mixing balancing overturning and air–sea fluxes, but with
546 opposite signs. Similarly, regimes i and ii both have mixing largely balancing overturning, but
547 with opposite signs.

548

549 In both Equatorial and off-Equatorial regions, there are regimes where air–sea fluxes
550 add volume and others where they remove it, suggesting that the regimes are not all stacked
551 vertically, but more likely horizontally (e.g. warming in the western subtropics and cooling in
552 the eastern). This would explain the causal connection between off-Equatorial regimes iii and
553 iv for example if the spatial regions corresponding to these regimes are connected by
554 atmospheric or oceanic teleconnections. Fig. 1(d), depicting a strong east-west temperature
555 gradient in the near-surface of the northern tropical Pacific, supports this conjecture. Albeit
556 only for a single month from one ensemble member, it suggests that above the 20°C isotherm,
557 off-Equatorial regimes ii-iv are stacked east-west and linked by the northern STC.

558

559 *(iii) Changes of volume in temperature classes*

560

561 Fig. 12 shows the logarithm of the ratio (ensemble mean \div control mean) of the
562 globally integrated volume in temperature space, calculated in 1K intervals and based on
563 monthly means, with a 12–month boxcar filter applied to smooth the data (before taking the

564 logarithm). Negative (positive) values indicate reduced (increased) volume in temperature
565 classes in the ensemble mean versus the control. In years 1-10 of the perturbed simulations,
566 there is a depletion of water in the range 27–30°C. Conversely there is a small increase in
567 volume at cooler temperatures, confined mainly to 25–27°C for the first five years, but
568 subsequently extending to cooler temperatures (down to 15°C). From year 11 onwards there
569 is an abrupt increase in the volume of the warmest waters, 27–30°C and a reduction of
570 volume in the 25–27°C range. The volume of water in the 18–25°C range decreases more
571 gradually, and in the 15–16°C range there is no reduction at all; indeed there is a long term
572 increase in the volume of these waters with little sign of impending decrease a decade after
573 the surge began. There are also notable increases of volume at around 10°C, still present at
574 year 20. The increase in volume in the –1–0°C range is a reflection of the drift in the control
575 associated with the long term negative TOA imbalance of -0.2 W m^{-2} .

576

577 Extending the plots to cover latitude, the logarithm of the absolute volume per unit
578 latitude for temperature classes in the control simulation (Fig. 13(a)) shows the largest
579 volume of seawater in the coldest temperature classes, and a comparatively low volume in the
580 warmest classes (black lines indicate the zonal and time average SST ± 1 standard deviation at
581 each latitude).

582

583 Fig. 13(b) shows the logarithm of the ratio (ensemble mean \div control mean) of
584 globally integrated volume for the first (blue) and second (red) decades of the perturbed
585 simulations. Negative (positive) values indicate a loss (gain) of volume compared to the
586 control. In the first decade there is a strong decrease in volume above 25°C and a
587 compensating, but more modest increase in the 15–25°C layer. In years 11–20 the warmer
588 layers regain some of the volume that was lost in the hiatus decade at the expense of loss in

589 volume of waters between 20-25°C; however the volume gained between 15- 20°C is retained
590 in the subsequent surge and new equilibrium period. The 15°C and 10°C volume classes in
591 particular remain much thicker than in the control in years 11–20.

592

593 Next we examine the change in volume per unit latitude in temperature classes as a
594 function of latitude for the hiatus (years 1–10) and the surge (years 11–20). In years 1–10
595 (Fig. 13(c)) there is considerable loss of volume of the warmest tropical waters (> 25°C)
596 between 20°S and 20°N. Below these waters we note an increased volume of waters between
597 15-25°C, extending poleward to 30°S/30°N, slightly greater in latitudinal extent than the
598 surface cooling/loss of volume. There is also increased volume of the warmest waters in the
599 Southern Ocean (80–40°S) and the northern midlatitudes (20–50°N).

600

601 In year 11–20 (Fig. 13(d)), we see a widespread increase in volume of the warmest
602 waters, including a large increase in the tropics where the volume was previously depleted.
603 The Southern Ocean remains as warm during the surge period as it was during the hiatus, but
604 the northern midlatitudes see a very strong increase in the volume of warmer waters. At
605 temperatures between 20–25°C, the extra volume deposited during the hiatus is lost, but
606 below 20°C the increased volume of water gained at the expense of the warmest waters
607 remains throughout the subsequent surge and the new equilibrium period.

608

609 In Figs 13(c) and (d) we have superimposed the circulation cells from Fig 9(b). These
610 demonstrate that the warming of the waters warmer than about 15°C during the hiatus in years
611 1-10 coincides with the positions of the subtropical cells. The warming and cooling pattern
612 north of 40°N seems associated with the AMOC.

613

614 Finally we investigate pentadal changes to the formation rate in the thermal regimes in
615 the perturbed ensemble (Fig. 14) in order to explain the changes in volume seen in Figs. 12
616 and 13. At the Equator, across all regimes, during the hiatus the changes to the mixing (blue)
617 create anomalies in the overturning and surface heat fluxes (Fig. 14(a), (b)). The overturning
618 anomalies die down in the early part of the surge (Fig. 14(c)) and completely disappear in the
619 second pentad (Fig. 14(d)), leaving a modified balance in the long term where the mixing
620 remains changed, but is balanced by a modified surface heat flux. There is a suggestion that
621 regimes ii and iv, and i and iii are opposite to each other in the final pentad.

622
623 A similar pattern occurs in the off-Equatorial regions where the changes in mixing
624 during the hiatus induce changes in the ocean circulation (Fig. 14(e), (f)), which however die
625 down and disappear during the surge (Fig. 14(g), (h)). The final state once again pitches long
626 term changes in mixing against changes in air–sea fluxes and regimes iii and iv mirror each
627 other, reinforcing the earlier suggestion that there is a causal connection between them (Fig.
628 11).

629 We defer detailed analysis of the mechanism governing this transient behaviour to a
630 future study, however we link the behaviour in temperature space in Fig. 14 to depth space
631 and previous studies (e.g. England et al. 2014) by plotting the pentadal evolution of the STCs
632 (Fig. 15(a)-(d)). The thin contours show the meridional overturning streamfunction from the
633 control simulation in the tropical (30°S-30°N) Indo-Pacific. There is a ~30Sv clockwise cell
634 in the northern tropics/subtropics and a somewhat stronger (~40 Sv) anticlockwise cell in the
635 Southern Hemisphere. Both have a meridional extent of ~25° of latitude, with the southern
636 cell extending slightly deeper (~300 m compared to ~200 m for the northern cell).

637 In the first pentad (hiatus, Fig. 15(a)) we see a modest strengthening of the southern
638 cell, but in the second there is a much bigger strengthening (order 5 Sv) of the northern cell

639 (Fig. 15(b)). In the third pentad (surge, Fig. 16(c)) the northern cell returns to normal strength,
640 whilst the southern cell weakens by ~ 5 Sv. In the fourth (Fig. 15(d)) and consistently with
641 Figure 14, we find both STCs have returned to normal. Note that when only one of the
642 overturning cells increases, we are likely to see opposing changes in the Equatorial versus the
643 off-Equatorial regions (e.g. in regimes iii and iv the red bars are of opposite sign between the
644 Equatorial and off-Equatorial regions), since increasing divergence at the Equator is
645 associated with increasing convergence off-Equator and *vice-versa*.

646 Consistent with previous studies (Meehl et al. 2013), increased heat drawdown in the
647 ocean is associated with strengthening STCs, and the reverse is true for surge conditions.
648 Thus, whilst the perturbed background diffusivity artificially induces a hiatus in our ensemble
649 experiment, heat uptake in the 300 m layer increases due to a realistic physical mechanism.
650 The stronger/weaker STCs in Figs. 15(b), (c) are in addition associated with
651 strengthened/weakened easterly trade winds (Figs. 15(f), (g)). In contrast there is no marked
652 signal in the wind stress when the STC's are near normal (Figs. 15(e), (h)). Therefore ocean
653 heat uptake and climate variability feed back on each other and artificial changes to the
654 background diffusivity induce a physically realistic circulation response in line with previous
655 studies.

656

657 **3.5 Conceptual model of transition**

658

659 The previous section demonstrated that, in the model, mixing is always of first order
660 importance in the thermal balance in the tropics and subtropics (Figs. 10 and 11). Combining
661 this result with the fact that the temperature tendency due to mixing is proportional to the 2nd
662 derivative of the temperature (Eq. (14)), we hypothesize that any modification to one of the

663 forcing terms will modify the temperature leading to changes in stratification, and bringing
664 the mixing term back into balance with its antagonist.

665

666 Consider an idealized upper ocean consisting of a 200m layer in which surface heating
667 (I) is balanced by diffusion $(\partial(K_v \partial\theta/\partial z)/\partial z)$ similar to regime iv, Fig. 10(a), with the
668 further simplification that ocean heat transport divergence, $\nabla \cdot (\mathbf{uT})$, is identically zero. We
669 apply a uniform heating of 20 W m^{-2} distributed evenly throughout the upper 80m
670 representing solar input. Initially, we specify zero heat flux through the surface boundary,
671 relying on the upper ocean heating to drive the model to equilibrium. We use the diffusive
672 heat flux at the lower boundary to drive a deeper layer of thickness 100m balanced by a fixed
673 ocean heat transport divergence of 20 W m^{-2} . This deeper layer is thus similar to regime iii in
674 Fig. 10(a.)

675

676 The upper 200m is discretized with a forward timestepping scheme and standard
677 second order accurate finite differences in the vertical (timestep $\Delta t=864\text{s}$, vertical grid
678 spacing $\Delta z=1\text{m}$). The lower 100m is treated numerically as a single layer. We assume a
679 constant background diffusion coefficient $K_v=5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$, set the initial temperature to 20°C
680 in all the layers, then integrate the model forward for 120 years. At the beginning of year 121
681 the background diffusivity is increased by 20% to $6 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ and in addition, heat flux
682 feedback to the atmosphere is switched on (Eq. 15), damping the SST back to the long term
683 mean (based on years 40-80) with a feedback parameter of $1 \text{ W m}^{-2} \text{ K}^{-1}$. At the beginning of
684 year 131, the background diffusivity is returned to its standard value, but the surface heat flux
685 feedback is retained and the model is integrated for a further 30 years.

686

687 The system evolves to equilibrium in about 20 years such that diffusion transports heat
688 downward from the surface (where it is input by heating) to the deepest layer where it is
689 removed by heat transport divergence and continues without significant change until year
690 120. When the diffusivity is increased at the beginning of year 121, there is an immediate
691 elevation of the isotherms above 140m and a depression of isotherms below, resulting in a
692 cold (warm) anomaly above (below) 140m depth (Fig. 16(a)), i.e. a hiatus.

693

694 When the diffusivity is returned to normal at the beginning of year 131, the system
695 experiences a temporary perturbation such that the export of heat by mixing from the upper
696 ocean through the lower boundary is reduced. There is thus a temporary imbalance in favour
697 of the surface heat flux, which modifies the temperature gradient to restore a balanced regime
698 (note that although we perturbed the system by changing the diffusion coefficient, we could
699 have modified one of the other processes, advection or air–sea fluxes to create the hiatus).

700

701 This behavior is very reminiscent of Fig. 2, albeit with larger anomalies as it is
702 intended to represent behavior near the Equator rather than an average over the globe. In Fig.
703 16(b) temperature anomaly profiles with depth are plotted before and after the hiatus–surge
704 transition. The hiatus state (red) shows a $\sim 0.7\text{K}$ reduction in temperature at the surface and a
705 corresponding increase at 200m. The cyan line shows how the temperature profile changes at
706 year 145 when the diffusivity has been reduced to its original value. A surface temperature
707 increase of a 0.2K over control values, with little depth variation is seen. This final
708 equilibrium state thus has a similar temperature gradient to the pre-hiatus state in the top 200
709 m, but is systematically warmer at all depths.

710

711 We now compare the zonally averaged response in the full climate model with this
712 conceptual model. Fig. 16(c) shows the ensemble mean zonal average (confined to the Pacific
713 basin) of the temperature in the top 200m of the Equatorial Pacific (3°S–3°N) in the full
714 climate model. We have noted previously that the major temperature anomaly during the
715 simulated hiatus sits below 200m and remains there during the surge. During the hiatus (red
716 line, representing an average of years 6–10 of the ensemble experiment), temperature above
717 120m reduces by up to 0.5K, whilst below this there is a temperature increase of a similar
718 magnitude. In the first year of the surge (cyan line, average over year 11), temperature at
719 200m stays at the same elevated level attained during the hiatus. Above this depth there is an
720 increase in temperature of 0.4K, approximately independent of depth. Hence in the surge
721 simulated by the climate model, as in the conceptual model, the temperature gradient returns
722 to pre-hiatus values, but the temperatures themselves are elevated due the increased
723 temperature of the deep layer.

724

725 The northern off-Equatorial region, 12–18°N, (16(d)), displays similar behavior to that
726 at the Equator. Here, the cyan lines are plotted for year 3 of the surge (not year 1) because it
727 takes longer for the system to attain equilibrium away from the Equator (varying Rossby
728 wave speed is the likely explanation). The southern off-Equatorial region displays the same
729 behavior, except the warming during the hiatus occurs deeper down at 400m (not shown).

730

731 Interestingly, in the conceptual model (Fig. 16(b)) we only had to increase the
732 diffusion coefficient by 20% to achieve a realistic response, whereas in HadCM3 we applied a
733 100% increase. Beyond the simplification inherent to the conceptual model, it is possible that
734 this mismatch comes from other sources of mixing in HadCM3, such as numerical mixing and
735 shear driven instability (e.g. Megann (2018)).

736

737 **4 Conclusions and Discussion**

738

739 We performed sensitivity experiments using the HadCM3 climate model to
740 understand decadal changes in the rate of global warming. Specifically, we increased the
741 ocean heat uptake by increasing the vertical diffusivity in the model for 10 years, before
742 returning vertical diffusivity to normal values. The main conclusions are as follows:

743

744 • Experiments with increased ocean diffusivity show Pacific-dominated
745 hiatus-like behavior: the Pacific undergoes a transition to a negative Inter-decadal
746 Pacific Oscillation like state. Outgoing longwave radiation at the Top of the
747 Atmosphere, oceanic latent heat loss and surface upwelling and downwelling
748 longwave radiation are all strongly reduced due to reduced surface temperatures,
749 however the ocean absorbs more shortwave radiation due to reduced atmospheric
750 absorption.

751 • When ocean vertical mixing is increased global surface temperature
752 decreases, i.e. the climate is forced into a hiatus-like state. When ocean vertical
753 mixing returns to normal values the surface temperature increases, and overshoots the
754 climatological values, i.e. a surge occurs.

755 • The climatological thermal balance in the warmest (near surface)
756 waters of tropical Pacific Ocean is maintained by mixing balancing surface heat flux.
757 In contrast in the cooler layer below mixing is balanced by ocean heat transport
758 divergence. Related, but slightly different balances prevail in the off-Equatorial
759 regions.

760 • The overshoot into surge conditions is due to long term depression of
761 deep ($\sim 15^{\circ}\text{C}$) isotherms following the hiatus (put simply, the ocean has no mechanism
762 to quickly remove the heat deposited at depth). In order to maintain thermal balance,
763 the upper layer temperature, the most responsive variable, must rise to higher than
764 normal values, via increased surface heat input in order to restore the thermal gradient,
765 which supports mixing (see schematic, Fig. 17).

766 • Other things being equal, the changes in stratification caused by a
767 hiatus increases the probability that a subsequent surge will occur

768

769 Due to our use of fundamental thermal balances in this study, we are confident that
770 our conclusions are correct in their essentials. However there are some caveats to consider:

771

772 • We have used one particular model of relatively coarse resolution
773 (although it maintains a very stable climate similar to the present day).

774 • The thermal balances may differ quantitatively from model to model.

775 • Our model does not resolve ocean mesoscale eddies which play an
776 important role in the thermal balance of the ocean (Griffies et al. 2015). In mitigation
777 our model employs the Gent McWilliams eddy parameterization scheme.

778 Our results are consistent with previous observational and modelling studies, in
779 particular, Maher et al. (2018). The authors modified Pacific trade winds to induce a hiatus in
780 a similar way to our modification of the mixing, and show similar results in terms of the
781 changes in subtropical cells, heat content, the transfer of excess heat to the Indian Ocean and
782 the legacy of anomalously high heat content even under a reversal of the perturbation. This
783 agreement as well as our Figs 14 and 15, suggest a tight coupling between the three
784 competing processes of surface fluxes, wind-driven ocean circulation and mixing, identified

785 in the present study. Similarly, observations (Nieves et al. 2015) confirm the transfer of heat
786 during the recent hiatus period to the Indian Ocean.

787

788 Finally we consider the wider implications of our findings. As is widely appreciated,
789 when the deep ocean is heated on decadal timescales, the additional heat cannot easily and
790 quickly be returned to the atmosphere. We have shown that this additional heat has a legacy
791 impact on the global mean surface temperature on decadal and longer timescales. The
792 additional heat uptake during hiatus periods (which is well attested) conditions the earth
793 system to warm even more strongly when the hiatus ends. Whilst we have engineered a hiatus
794 by modifying the vertical mixing, in principle we would expect to see the same behavior if the
795 heat enters the deep ocean by other means (e.g. via modification of the shallow overturning
796 cells from windstress perturbations, or changes to the surface heating and subsequent
797 subduction, mechanisms which have been hypothesized to have driven the early 2000s
798 hiatus). Idealized/conceptual one-dimensional models of Earth's energy balance do not
799 currently include the effects described in this paper.

800

801 Our results also have important implications for decadal and longer predictions using
802 model based forecast systems. In particular we suggest that accurate initialization of the
803 subsurface ocean may be a key element of such systems as inaccuracies could result in
804 erroneous predictions of hiatus or surge conditions (consistently with the findings of Sévellec
805 and Federov, 2017, and Germe et al., 2018). Similarly, inaccurate simulation of forced hiatus
806 periods could lead to systematic biases or additional uncertainty in centennial climate
807 projections.

808

809 Further work should establish the robustness of our result in a wider variety of models
810 (including at higher ocean resolution). The watermass transformation method we used could
811 also in principle be applied to observational data, although in practice great care would be
812 needed to take into account of observational uncertainties and internal variability.

813

814 **Acknowledgements**

815

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818 Surge events (SMURPHS) project, grants NE/N005686/1 (BS), NE/N006054/1 (JR),
819 NE/N005767/1 (FS). AGN was supported by the NERC Transient tracer-based Investigation
820 of Circulation and Thermal Ocean Change (TICTOC) project, grant (NE/P019293/1). We
821 thank Jonathan Gregory for useful discussions.

822

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	HadCM3 control simulation	Wild et al. (2013) observation based estimates
TOA incident solar	341.4	340 (340-341)
TOA outgoing solar	101.1	100 (96-100)
Atmospheric solar absorption	76.1	79 (74-91)
Surface incident solar	188.5	185 (179-189)
Surface reflected solar	24.3	24 (22-26)
Surface solar absorption	164.2	161 (154-166)
Outgoing longwave	240.5	239 (236-242)
Surface upwelling longwave	394.6	398 (394-400)
Surface downwelling longwave	332.0	342 (338-348)
Surface sensible heat	17.0	20 (15-25)
Surface latent heat	84.1	84 (70-85)

917

918 Table 1. Summary of annual mean energy flows ($W m^{-2}$) in the Hadcm3 control
919 simulation and comparison with observational estimates) Note that, as defined by Wild et al.

920 (2013), the TOA outgoing solar comprises all reflected solar radiation including that reflected
921 from the surface.

922

923 **Figure Captions**

924 Fig. 1. Derivation of the watermass transformation equations (a) Volume conservation
925 for waters with potential temperature less than or equal to θ (b) Heat conservation for the
926 same region (c) watermass formation rate between isopycnals θ and $\theta + \delta\theta$ due to surface
927 flux forcing. Illustration of 3D watermass transformation diagnostics (d) ocean temperature in
928 the western tropical Pacific (contours). Surface waters below 22°C are shaded brown. Surface
929 and subsurface waters with temperatures between 22°C and 23°C are shaded purple. Coloured
930 arrows indicate watermass transformation by air-sea heat exchange (yellow) and ocean
931 circulation/overturning (green) across the 22°C and 23°C isotherms. Solid (dashed) black
932 arrows indicated diapycnal mixing across the 23°C (22°C) isotherm (e) as (d) but for a
933 domain extended slightly to the south. The corresponding diapycnal mixing in the southward
934 extension to the domain is indicated by hollow arrows.

935

936 Fig. 2. Time series of ensemble mean global mean area-weighted temperature
937 anomaly (K) with respect to the 140-year control. Annual means are plotted and a 4-year
938 boxcar filter is applied to smooth the data. Colour shading indicates values significantly
939 different from the control simulation at the 95% confidence level. The dotted vertical line
940 separates the hiatus period (years 1-10, doubled vertical diffusivity) from the surge period
941 (years 11-20, standard vertical diffusivity).

942

943 Fig. 3. Ensemble mean zonal mean temperature anomaly (K) with respect to the 140–
944 year control mean during (a) hiatus period (years 6–10) (b) surge period (years 11-15). Thick

945 black contour denotes values significantly different from the control simulation at the 95%
946 confidence level. Thin black contours denote the control mean temperature ($^{\circ}\text{C}$).

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948 Fig. 4. Ensemble mean SST minus 140-year mean control SST (K) during (a) hiatus
949 period (years 6–10) (b) surge period (years 11–15) (c) and (d) as (a) and (b) for 0–1000m
950 heat content (TJ) (e) and (f) as (a) and (b) for net surface heat flux (W m^{-2}). Heat flux is
951 positive downward. Thin black contour line denotes values that are significantly different
952 from the control simulation at the 95% confidence level.

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954 Fig. 5. Ensemble annual mean anomalous global energy flows (W m^{-2}) (a) Short wave
955 terms. Black – TOA upwards shortwave; red – atmospheric shortwave absorption; green
956 incident surface shortwave; blue – reflected shortwave; cyan – oceanic shortwave absorption
957 (b) Long wave and turbulent heat flux terms. Black – Outgoing Longwave Radiation at TOA;
958 red – surface upwelling longwave radiation; green – incident downwelling longwave radiation;
959 blue – sensible heat flux; cyan – latent heat flux. Coloured horizontal lines denote zero
960 anomaly. In order to allow the individual anomalies to be seen clearly, each quantity has had
961 an offset subtracted – black – no offset, red -1 W m^{-2} , Green -2.5 W m^{-2} ; blue -3 W m^{-2} ;
962 cyan -4 W m^{-2} . Annual means are plotted and a 4-year boxcar filter is applied to smooth the
963 data. Thick portions of the curves indicate values significantly different from the control
964 simulation at the 95% confidence level. The dotted vertical line separates the hiatus period
965 (years 1–10, doubled vertical diffusivity) from the surge period (years 11–20 standard vertical
966 diffusivity).

967

968 Fig. 6. Ensemble mean global annual mean net ocean heat uptake (net ocean surface
969 downward heat flux) (W m^{-2}) is indicated by the thick red line. Thin red lines indicate ± 1

970 ensemble standard deviation. Annual means are plotted and a 4-year boxcar filter is applied to
971 smooth the data. The vertical black line is centered on the 140-year mean value from the
972 control simulation and indicates ± 1 standard deviations (or ± 2 standard errors). The dotted
973 vertical line separates the hiatus period (years 1–10, doubled vertical diffusivity) from the
974 surge period (years 11–15, standard vertical diffusivity).

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976 Fig. 7. Ensemble mean global mean SST ($^{\circ}\text{C}$) is indicated by the thick red line. Thin
977 red lines indicate ± 1 ensemble standard deviation. The green line shows global mean SST for
978 an individual ensemble member run out for 50 years. Monthly means are plotted and a 36-
979 month boxcar filter is applied to smooth the data. The vertical black line is centered on the
980 140-year mean value from the control simulation and indicates ± 1 standard deviations (or ± 2
981 standard errors). The dotted vertical line separates the hiatus period (years 1–10, doubled
982 vertical diffusivity) from the surge period (years 11–20, standard vertical diffusivity).

983

984 Fig. 8. Scatter plot of global annual mean SAT versus SST (both in $^{\circ}\text{C}$) for control
985 (red circles), hiatus period (years 1–10 of the ensemble experiment, black symbols) and surge
986 period (years 11–20 of the ensemble experiment, blue symbols). Individual ensemble
987 members are differentiated by different symbols: circles, stars, squares and triangles.

988

989 Fig. 9. Transformation streamfunctions (S_v) for the 140-year control simulation in
990 temperature space (a) time derivative (b) ocean circulation (c) surface heat flux (d) mixing.
991 Blue (red) shading and arrows indicate clockwise (anticlockwise) transformation. Black lines
992 show the average SST at each latitude and ± 1 standard deviation.

993

994 Fig. 10. (a) Latitudinal divergence of the volume flux ($\text{Sv}/^\circ\text{latitude}$) across isopycnals
995 (transformation rate) at the Equator (3°S – 3°N) as a function of temperature associated with
996 surface heat flux (green), overturning (red), mixing (blue). Green circles demarcate the
997 boundaries between the prevailing thermal regimes. (b) as (a) for the “subtropical” regions
998 (average of 18°S – 12°S and 12°N – 18°N).

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1000 Fig. 11. (a) Formation rate ($\text{Sv}/^\circ\text{latitude}$) at the Equator (3°S – 3°N) in the control
1001 simulation associated with overturning (red), surface heat flux (green), mixing (blue) for the
1002 four thermal regimes (i–iv) defined in Fig. 8. (b) as (a) for the “subtropical” regions (average
1003 of 18°S – 12°S and 12°N – 18°N).

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1005 Fig. 12. Logarithm to base ten of the ratio of ensemble mean to control seawater
1006 volume in temperature classes (1K bins centered on integral values). Evaluated over the
1007 global ocean. The dotted vertical line separates the hiatus period (years 1–10, doubled vertical
1008 diffusivity) from the surge period (years 11–15, standard vertical diffusivity).

1009

1010 Fig. 13. (a) Logarithm to base ten of the control simulation seawater volume in
1011 temperature classes (1K bins centered on integral values) evaluated at each latitude. (b)
1012 Logarithm to base ten of the ratio of ensemble mean to control global seawater volume in
1013 temperature classes (1K bins centered on integral values). Blue – average over hiatus period;
1014 red – average over surge period. Vertical black line indicates a zero value. (c) Logarithm to
1015 base ten of the ratio of ensemble mean to control seawater volume in temperature classes (1K
1016 bins centered on integral values). Evaluated at each latitude and averaged over the hiatus
1017 period, years 1–10 (d) as (c) for the surge period, years 11–20. Black lines show the average
1018 SST at each latitude and ± 1 standard deviation.

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Fig. 14. Ensemble mean anomalous formation rate ($Sv/^\circ\text{latitude}$) at the Equator (3°S – 3°N) associated with overturning (red), surface heat flux (green), mixing (blue) and volume change (cyan) for the four thermodynamic regimes (i–iv) defined in Fig. 8. (a) years 1–5 (b) year 6–10 (c) years 11–15 (d) years 16–20. (e)–(h) as (a)–(d) for the “subtropical” regions (average of 18°S – 12°S and 12°N – 18°N).

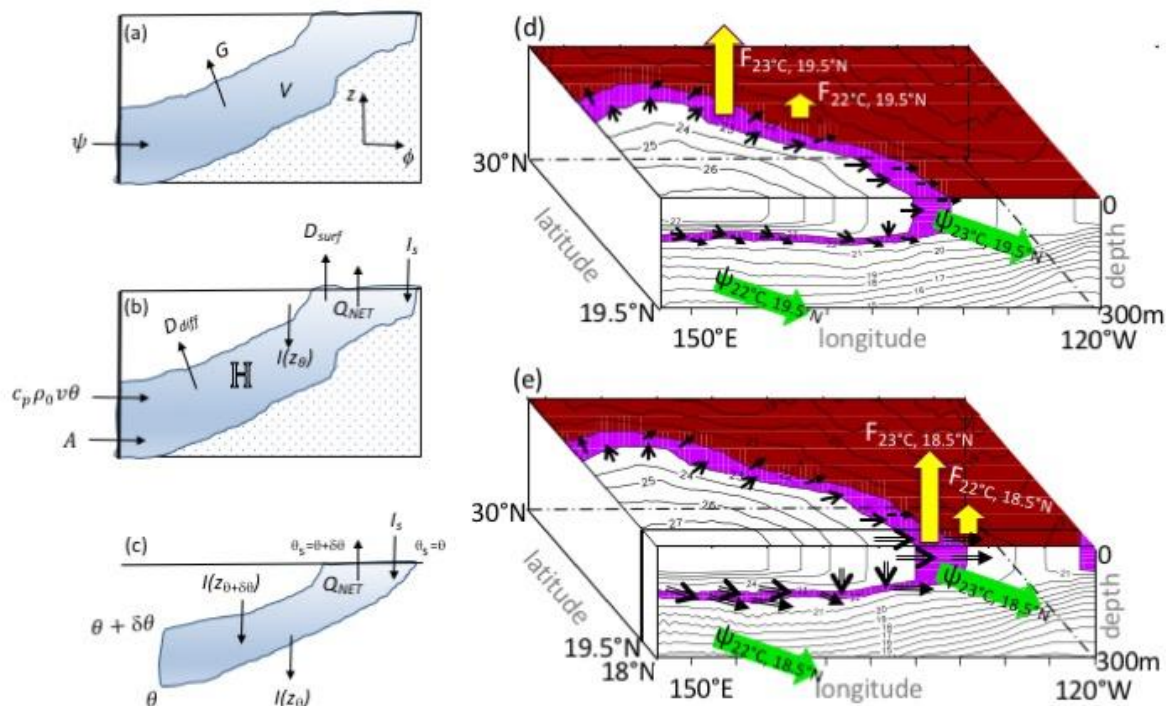
Fig 15 (a) Ensemble mean anomalous overturning streamfunction (Sv) in the subtropics (colour shading) for years 1–5 (b) year 6–10 (c) years 11–15 (d) years 16–20 (e) ensemble mean anomalous zonal wind (m s^{-1}) for years 1–5 (f) year 6–10 (g) years 11–15 (h) years 16–20. Thick black contour and dotted regions denote anomalies significantly different from the control simulation at the 95% confidence level. Thin black contours in (a)–(d) denote the control mean streamfunction (Sv).

Fig. 16. (a) Temperature anomaly (K) predicted by the conceptual model. The dotted vertical line separates the hiatus period (years 121–130, higher vertical diffusivity) from the surge period (years 131–160, standard vertical diffusivity). (b) anomaly-depth profile predicted by the conceptual model at hiatus year 125 (red); surge year 145 (cyan). (c) Ensemble mean zonally averaged Equatorial (3°S – 3°N) temperature anomaly profile from the coupled climate model experiment as a function of depth in the Pacific Ocean averaged over the hiatus period (red); first year of the surge (cyan); (d) as (c) except for the off-equatorial regions (average of 18°S – 12°S and 12°N – 18°N). Cyan line shows average over year 3 of the surge.

1043 Fig. 17. Schematic of the hiatus to surge mechanism described in this paper (a)

1044 conditions during hiatus (b) conditions during surge (c) conditions after surge.

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Fig. 1. Derivation of the watermass transformation equations (a) Volume conservation

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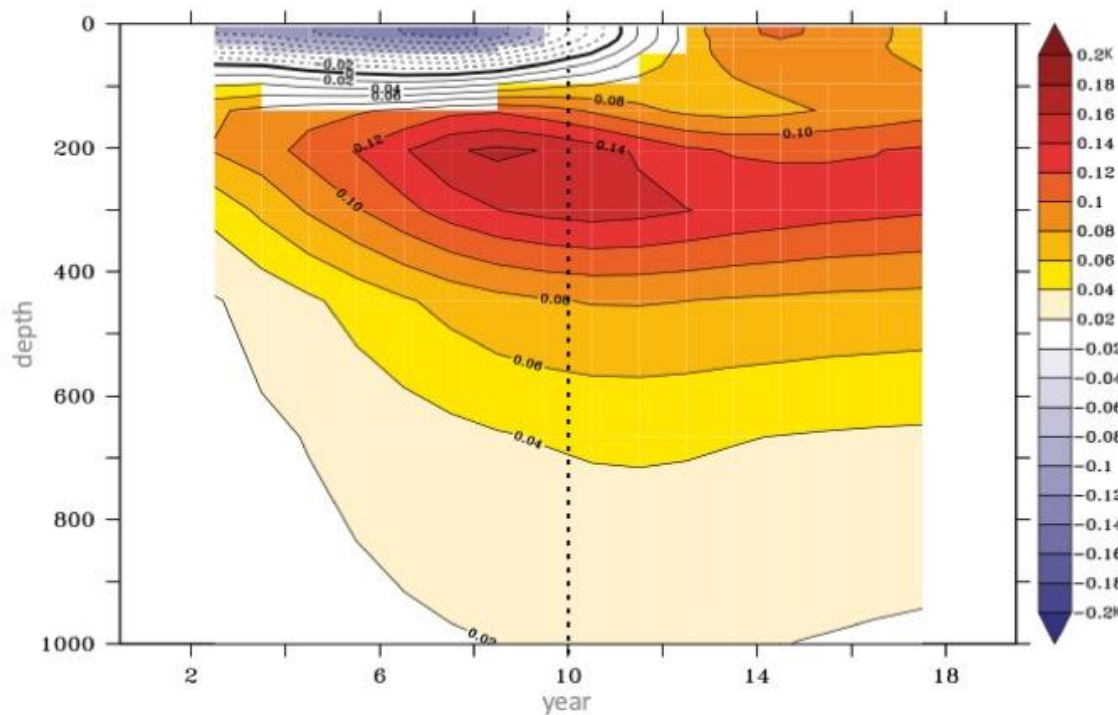
domain extended slightly to the south. The corresponding diapycnal mixing in the southward

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extension to the domain is indicated by hollow arrows.

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1060 Fig. 2. Time series of ensemble mean global mean area-weighted temperature
 1061 anomaly (K) with respect to the 140-year control. Annual means are plotted and a 4-year
 1062 boxcar filter is applied to smooth the data. Colour shading indicates values significantly
 1063 different from the control simulation at the 95% confidence level. The dotted vertical line
 1064 separates the hiatus period (years 1-10, doubled vertical diffusivity) from the surge period
 1065 (years 11-20, standard vertical diffusivity).
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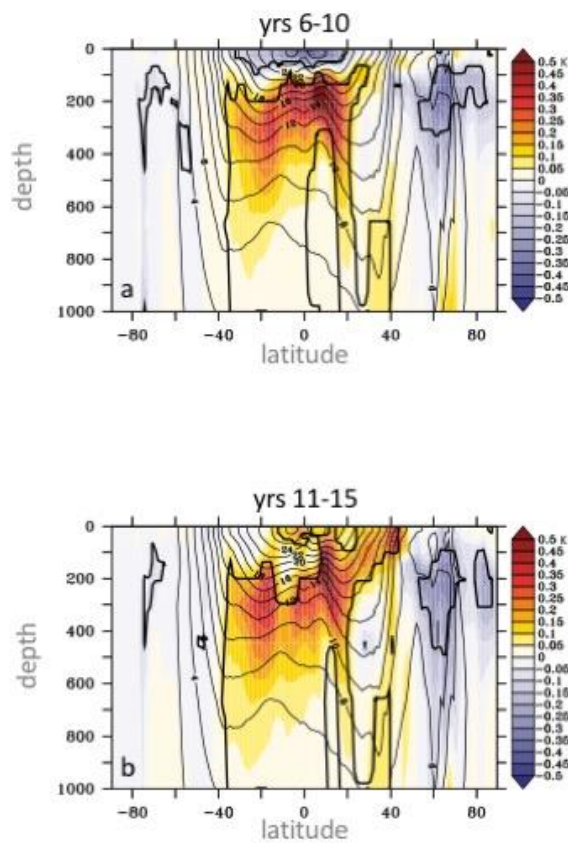


Fig. 3. Ensemble mean zonal mean temperature anomaly (K) with respect to the 140–
year control mean during (a) hiatus period (years 6–10) (b) surge period (years 11–15). Thick
black contour denotes values significantly different from the control simulation at the 95%
confidence level. Thin black contours denote the control mean temperature ($^{\circ}\text{C}$).

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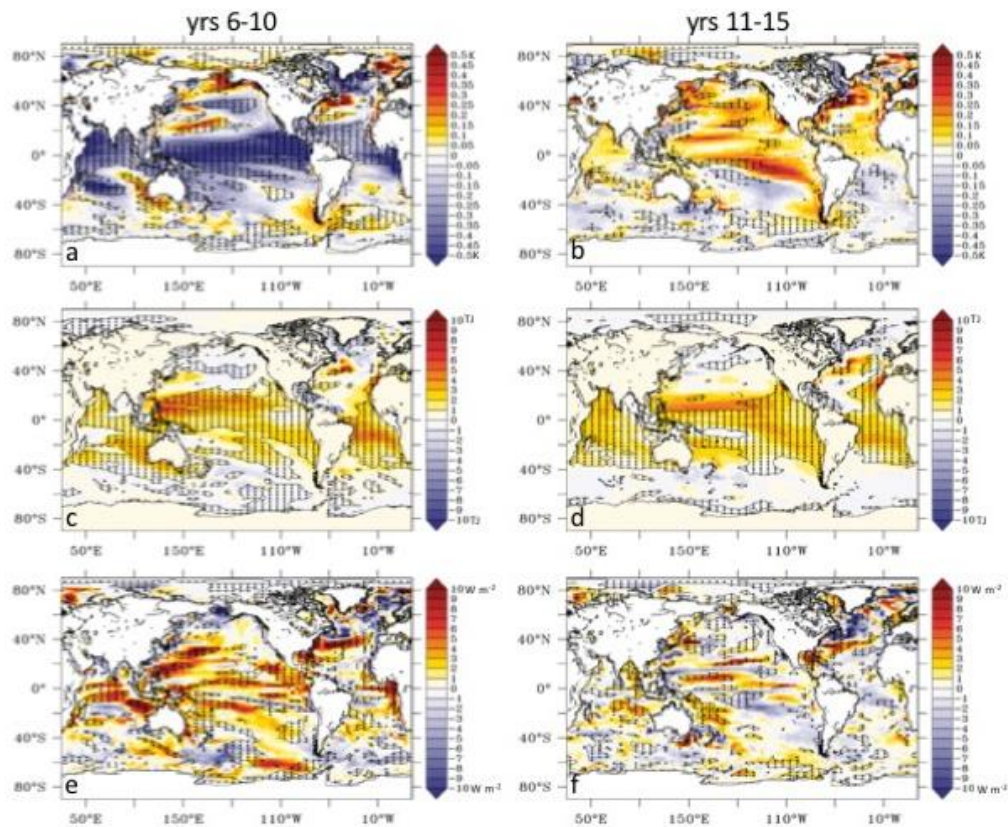
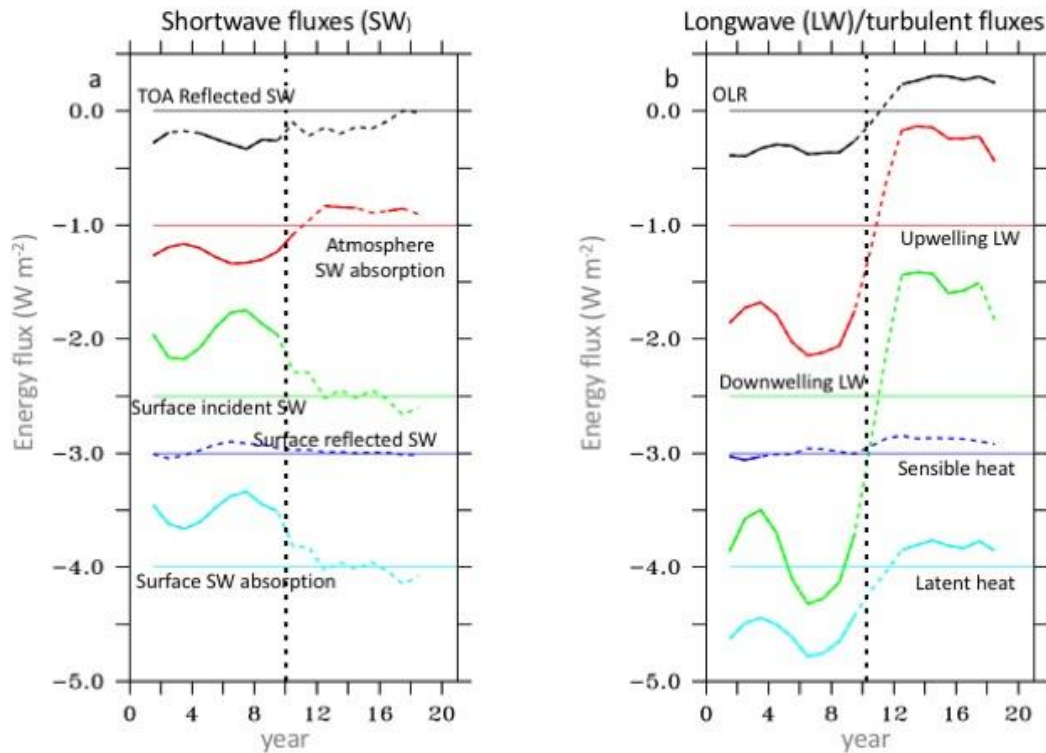


Fig. 4. Ensemble mean SST minus 140-year mean control SST (K) during (a) hiatus period (years 6–10) (b) surge period (years 11–15) (c) and (d) as (a) and (b) for 0–1000m heat content (TJ) (e) and (f) as (a) and (b) for net surface heat flux ($W m^{-2}$). Heat flux is positive downward. Thin black contour line denotes values that are significantly different from the control simulation at the 95% confidence level.



1111 Fig. 5. Ensemble annual mean anomalous global energy flows (W m^{-2}) (a) Short wave
 1112 terms. Black – TOA upwards shortwave; red – atmospheric shortwave absorption; green
 1113 incident surface shortwave; blue – reflected shortwave; cyan – oceanic shortwave absorption
 1114 (b) Long wave and turbulent heat flux terms. Black – Outgoing Longwave Radiation at TOA;
 1115 red – surface upwelling longwave radiation; green – incident downwelling longwave radiation;
 1116 blue – sensible heat flux; cyan – latent heat flux. Coloured horizontal lines denote zero
 1117 anomaly. In order to allow the individual anomalies to be seen clearly, each quantity has had
 1118 an offset subtracted – black – no offset, red -1 W m^{-2} , Green -2.5 W m^{-2} ; blue -3 W m^{-2} ;
 1119 cyan -4 W m^{-2} . Annual means are plotted and a 4-year boxcar filter is applied to smooth the
 1120 data. Thick portions of the curves indicate values significantly different from the control
 1121 simulation at the 95% confidence level. The dotted vertical line separates the hiatus period
 1122 (years 1–10, doubled vertical diffusivity) from the surge period (years 11–20 standard vertical
 1123 diffusivity).
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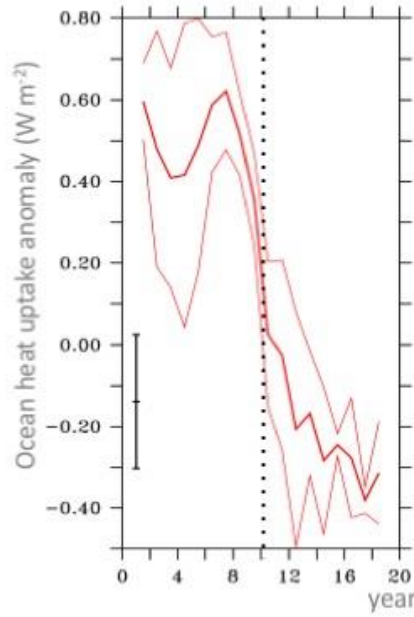
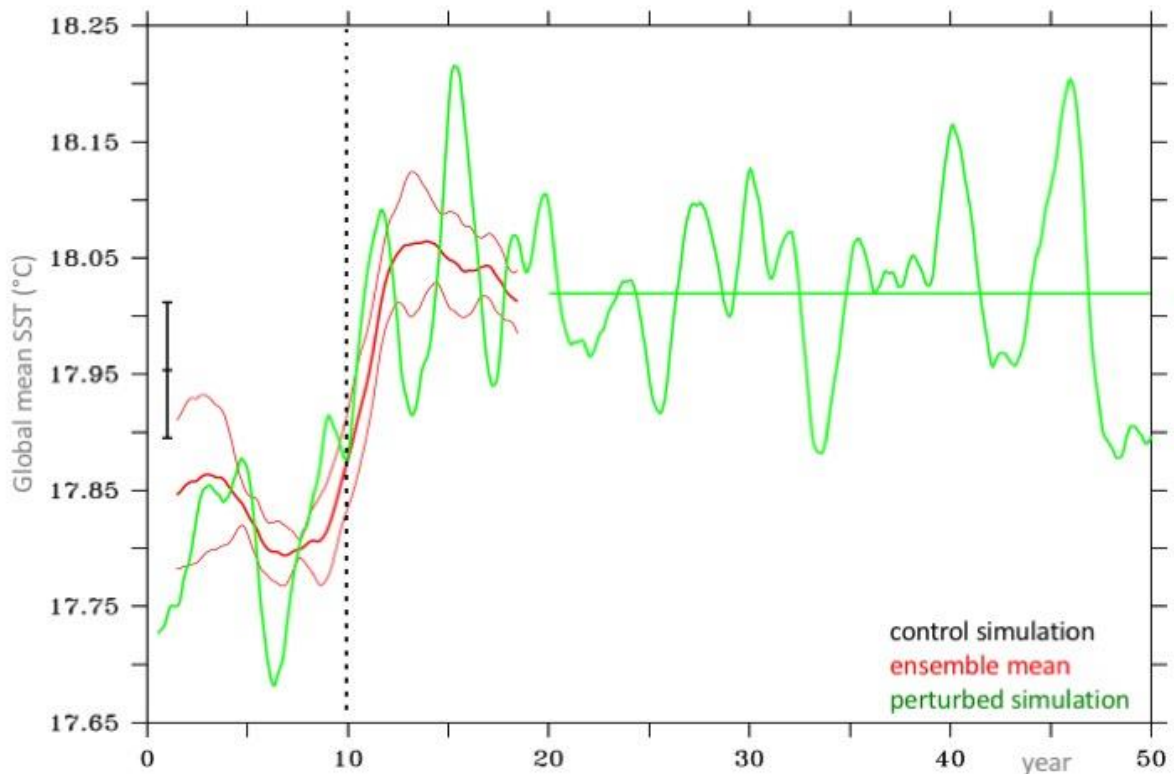
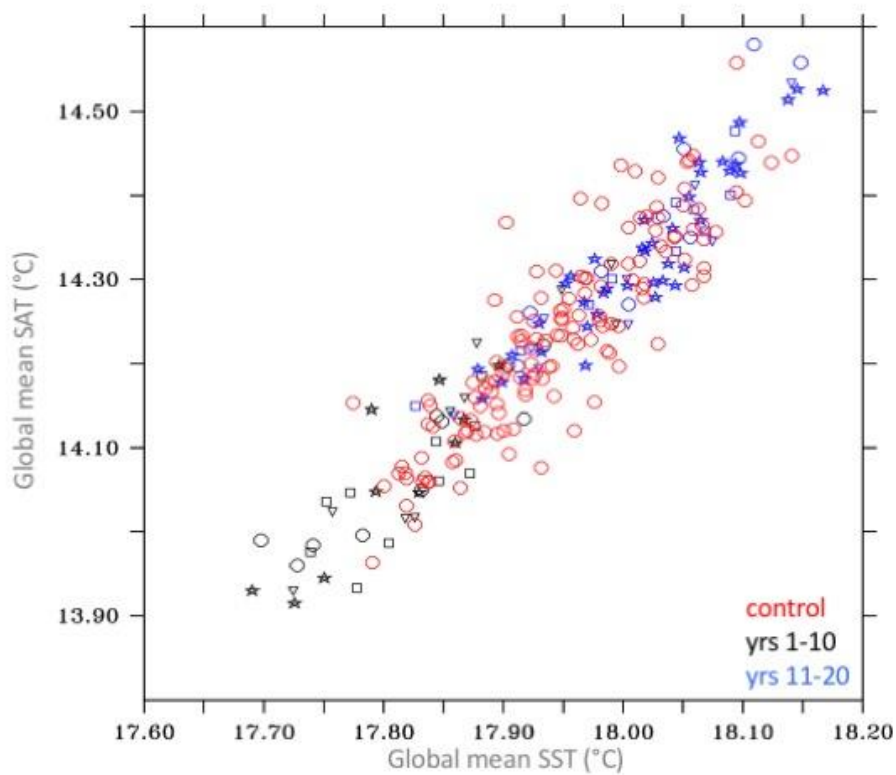


Fig. 6. Ensemble mean global annual mean net ocean heat uptake (net ocean surface downward heat flux) (W m^{-2}) is indicated by the thick red line. Thin red lines indicate ± 1 ensemble standard deviation. Annual means are plotted and a 4-year boxcar filter is applied to smooth the data. The vertical black line is centered on the 140-year mean value from the control simulation and indicates ± 1 standard deviations (or ± 2 standard errors). The dotted vertical line separates the hiatus period (years 1–10, doubled vertical diffusivity) from the surge period (years 11–15, standard vertical diffusivity).



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Fig. 7. Ensemble mean global mean SST ($^{\circ}\text{C}$) is indicated by the thick red line. Thin red lines indicate ± 1 ensemble standard deviation. The green line shows global mean SST for an individual ensemble member run out for 50 years. Monthly means are plotted and a 36-month boxcar filter is applied to smooth the data. The vertical black line is centered on the 140-year mean value from the control simulation and indicates ± 1 standard deviations (or ± 2 standard errors). The dotted vertical line separates the hiatus period (years 1–10, doubled vertical diffusivity) from the surge period (years 11–20, standard vertical diffusivity).

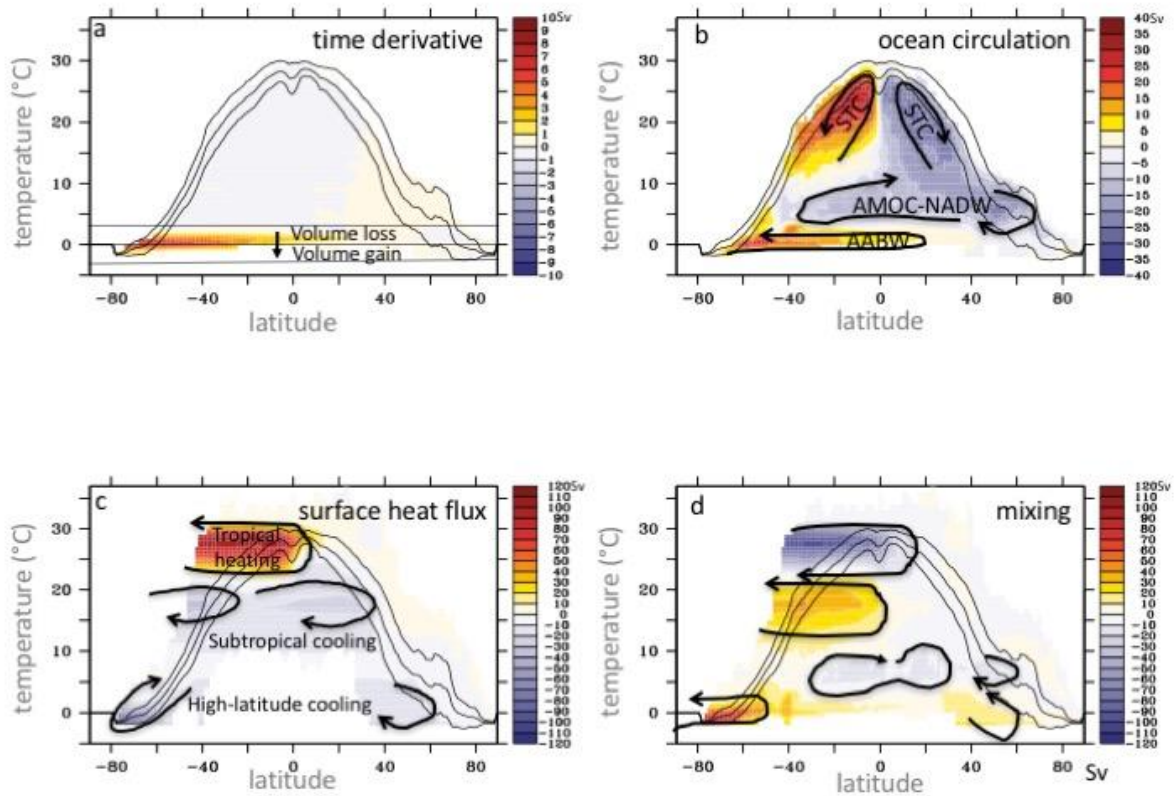


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Fig. 8. Scatter plot of global annual mean SAT versus SST (both in °C) for control

1167 (red circles), hiatus period (years 1–10 of the ensemble experiment, black symbols) and surge
 1168 period (years 11–20 of the ensemble experiment, blue symbols). Individual ensemble
 1169 members are differentiated by different symbols: circles, stars, squares and triangles.

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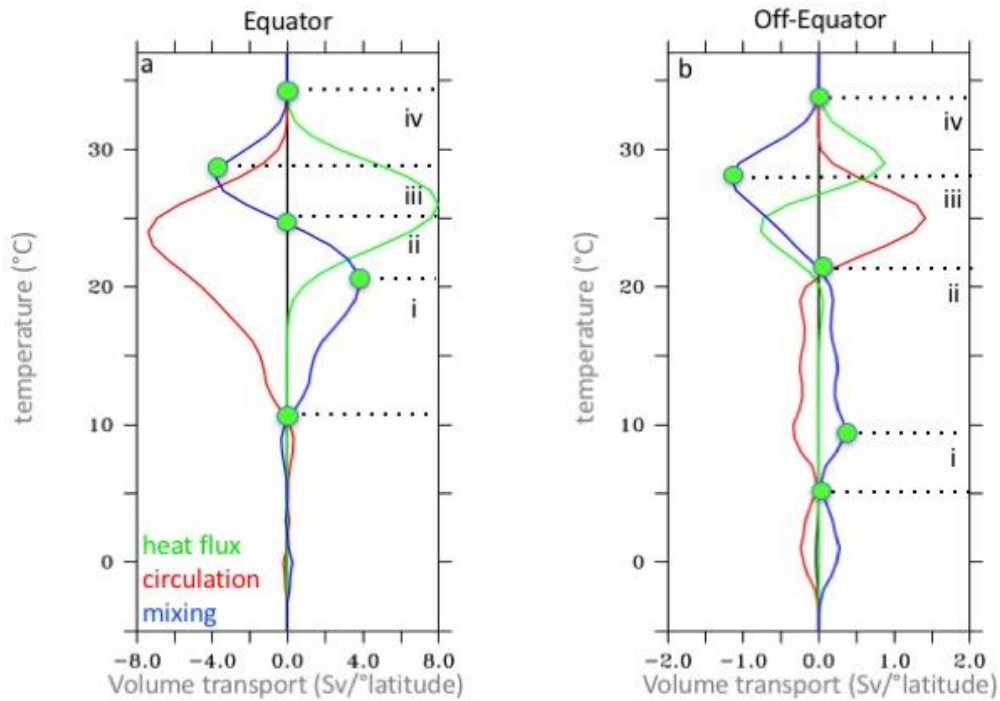
1178 Fig. 9. Transformation streamfunctions (Sv) for the 140-year control simulation in

1179 temperature space (a) time derivative (b) ocean circulation (c) surface heat flux (d) mixing.

1180 Blue (red) shading and arrows indicate clockwise (anticlockwise) transformation. Black lines

1181 show the average SST at each latitude and ± 1 standard deviation.

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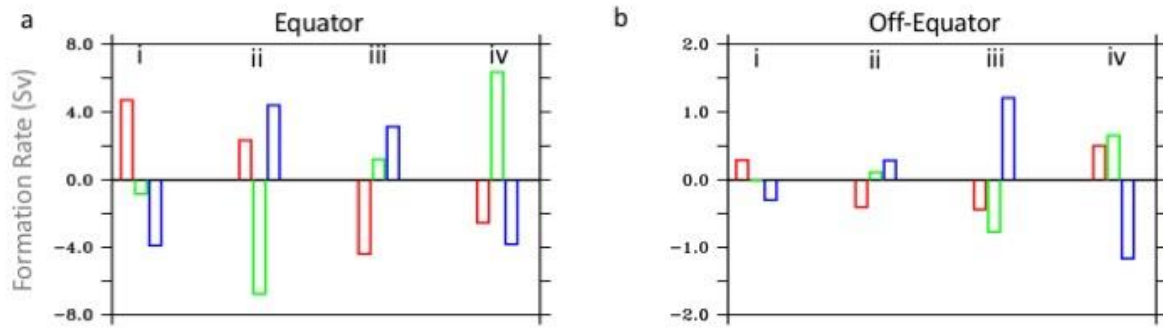


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Fig. 10. (a) Latitudinal divergence of the volume flux ($\text{Sv}/^\circ\text{latitude}$) across isopycnals

1185 (transformation rate) at the Equator (3°S – 3°N) as a function of temperature associated with
1186 surface heat flux (green), overturning (red), mixing (blue). Green circles demarcate the
1187 boundaries between the prevailing thermal regimes. (b) as (a) for the “subtropical” regions
1188 (average of 18°S – 12°S and 12°N – 18°N).

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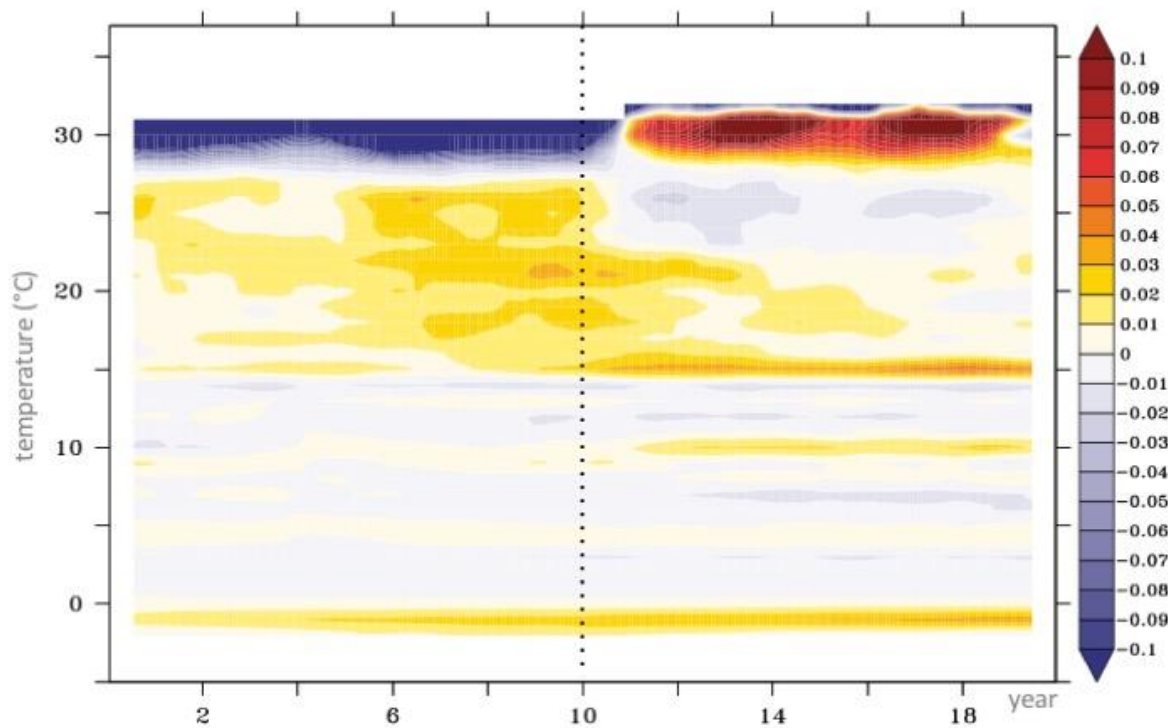
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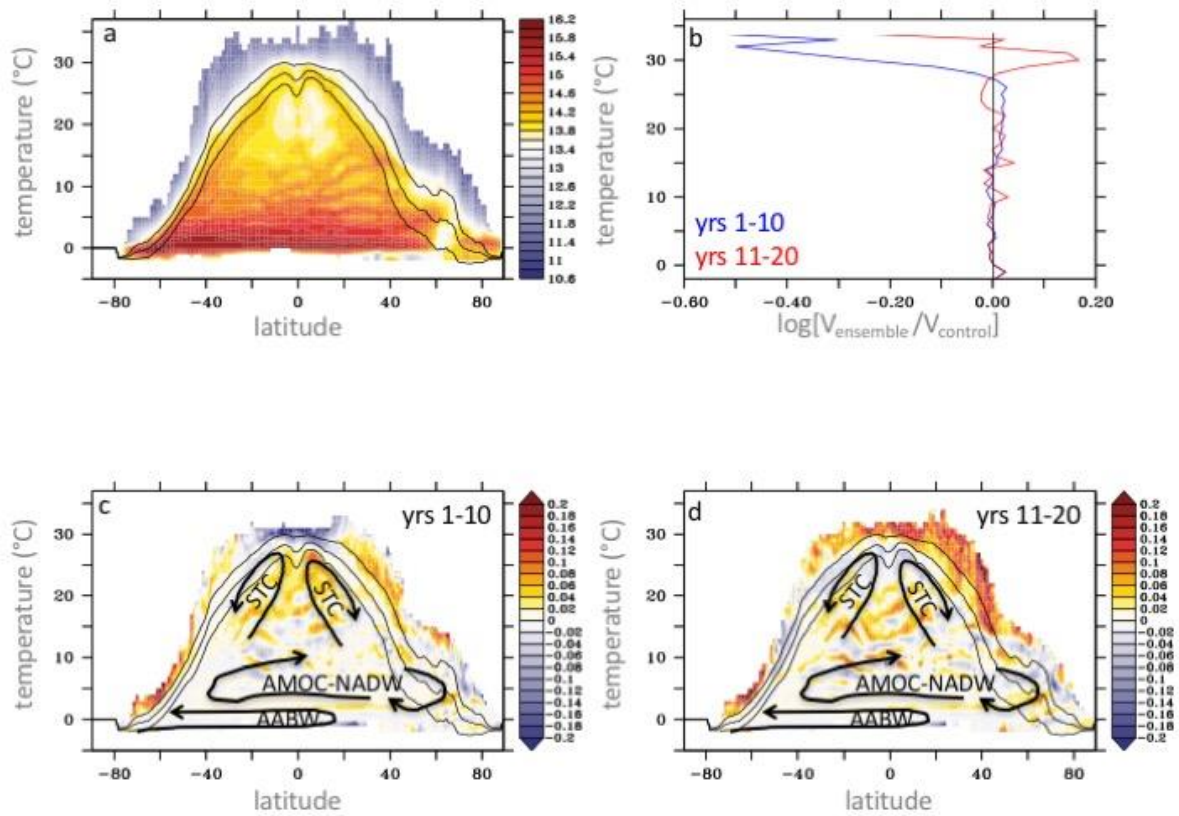
Fig. 11. (a) Formation rate (Sv/°latitude) at the Equator (3°S–3°N) in the control simulation associated with overturning (red), surface heat flux (green), mixing (blue) for the four thermal regimes (i–iv) defined in Fig. 8. (b) as (a) for the “subtropical” regions (average of 18°S–12°S and 12°N–18°N).



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1201 Fig. 12. Logarithm to base ten of the ratio of ensemble mean to control seawater
1202 volume in temperature classes (1K bins centered on integral values). Evaluated over the
1203 global ocean. The dotted vertical line separates the hiatus period (years 1–10, doubled vertical
1204 diffusivity) from the surge period (years 11–15, standard vertical diffusivity).

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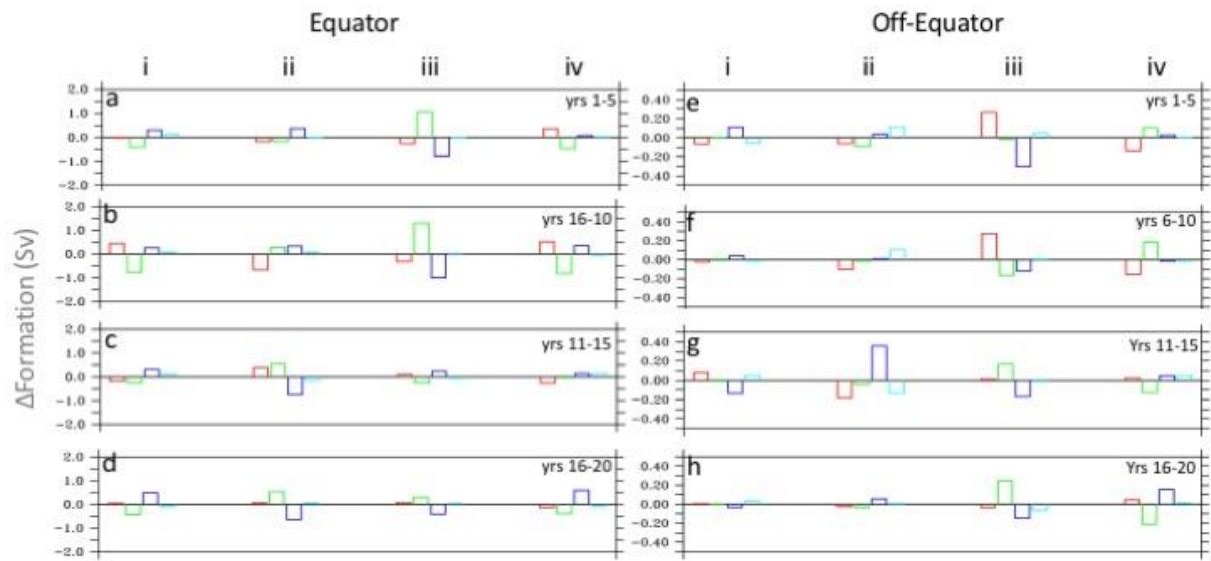
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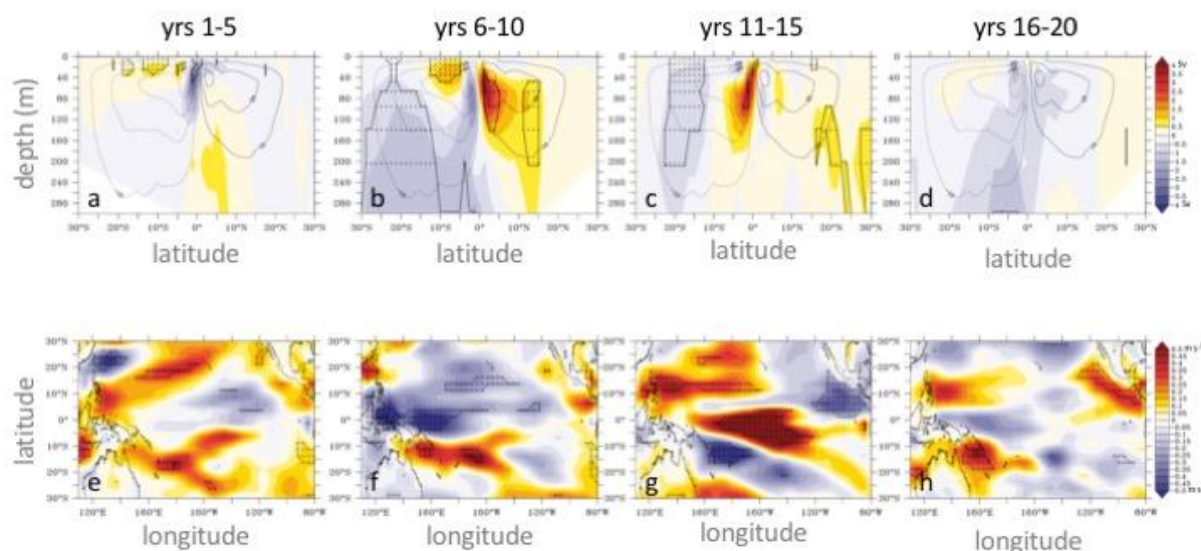
Fig. 13. (a) Logarithm to base ten of the control simulation seawater volume in temperature classes (1K bins centered on integral values) evaluated at each latitude. (b) Logarithm to base ten of the ratio of ensemble mean to control global seawater volume in temperature classes (1K bins centered on integral values). Blue – average over hiatus period; red – average over surge period. Vertical black line indicates a zero value. (c) Logarithm to base ten of the ratio of ensemble mean to control seawater volume in temperature classes (1K bins centered on integral values). Evaluated at each latitude and averaged over the hiatus period, years 1–10 (d) as (c) for the surge period, years 11–20. Black lines show the average SST at each latitude and ± 1 standard deviation.



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1223 Fig. 14. Ensemble mean anomalous formation rate ($\text{Sv}/^\circ\text{latitude}$) at the Equator (3°S –
1224 3°N) associated with overturning (red), surface heat flux (green), mixing (blue) and volume
1225 change (cyan) for the four thermodynamic regimes (i–iv) defined in Fig. 8. (a) years 1–5 (b)
1226 year 6–10 (c) years 11–15 (d) years 16–20. (e)–(h) as (a)–(d) for the “subtropical” regions
1227 (average of 18°S – 12°S and 12°N – 18°N).

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1231 Fig 15 (a) Ensemble mean anomalous overturning streamfunction (Sv) in the
 1232 subtropics (colour shading) for years 1–5 (b) year 6–10 (c) years 11–15 (d) years 16–20 (e)
 1233 ensemble mean anomalous zonal wind ($m\ s^{-1}$) for years 1-5 (f) year 6–10 (g) years 11–15 (h)
 1234 years 16–20. Thick black contour and dotted regions denote anomalies significantly different
 1235 from the control simulation at the 95% confidence level. Thin black contours in (a)-(d) denote
 1236 the control mean streamfunction (Sv).

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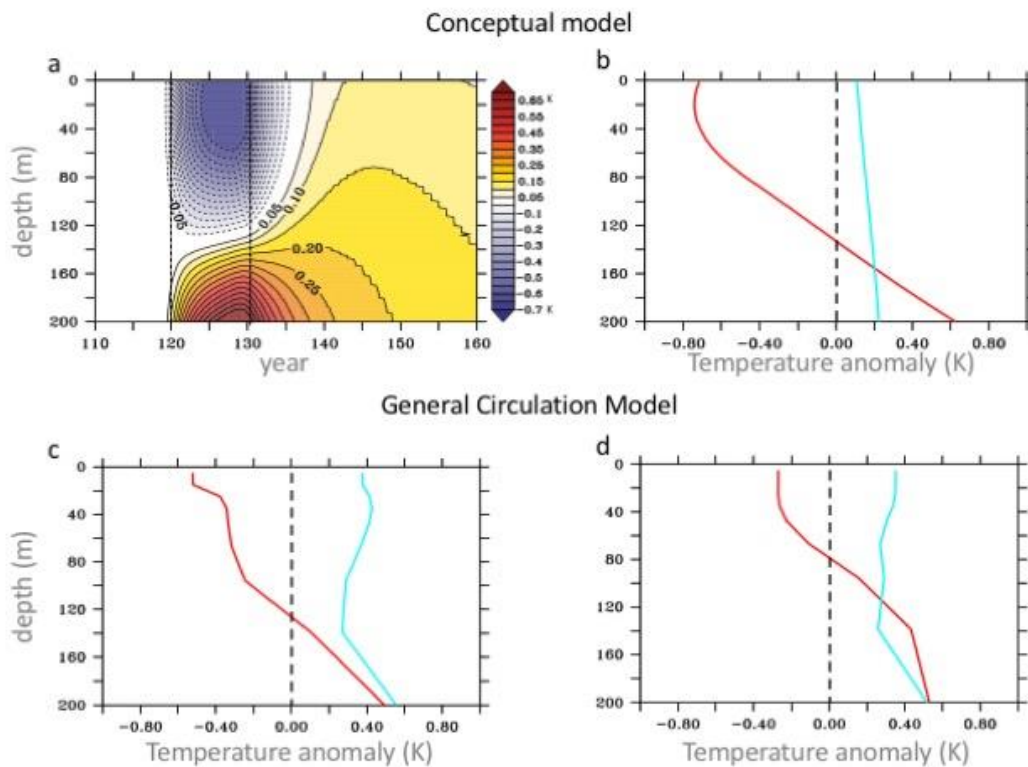
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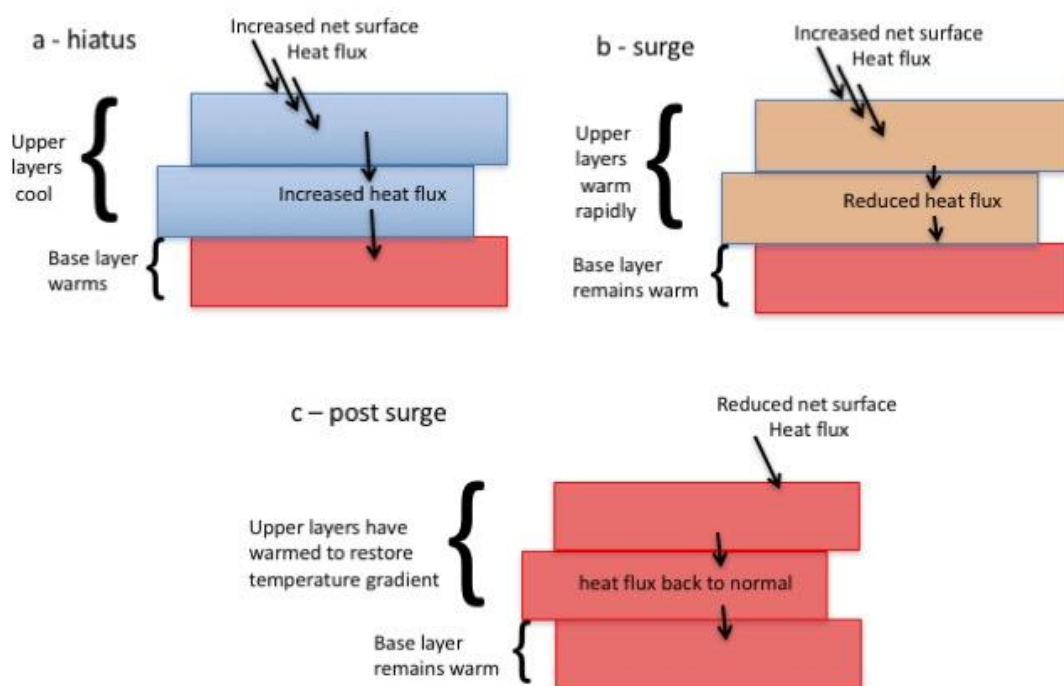
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Fig. 16. (a) Temperature anomaly (K) predicted by the conceptual model. The dotted vertical line separates the hiatus period (years 121–130, higher vertical diffusivity) from the surge period (years 131–160, standard vertical diffusivity). (b) anomaly-depth profile predicted by the conceptual model at hiatus year 125 (red); surge year 145 (cyan). (c) Ensemble mean zonally averaged Equatorial (3°S–3°N) temperature anomaly profile from the coupled climate model experiment as a function of depth in the Pacific Ocean averaged over the hiatus period (red); first year of the surge (cyan); (d) as (c) except for the off-equatorial regions (average of 18°S–12°S and 12°N–18°N). Cyan line shows average over year 3 of the surge.

Hiatus to surge mechanism



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1261 Fig. 17. Schematic of the hiatus to surge mechanism described in this paper (a)
1262 conditions during hiatus (b) conditions during surge (c) conditions after surge.