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1	Internal ocean dynamics contribution to North Atlantic interdecadal
2	variability strengthened by ocean-atmosphere thermal coupling
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ABSTRACT: Identifying the primary drivers of North Atlantic interdecadal climate variability is 8 crucial for improving climatic prediction over the coming decades. Here the effect of thermal cou-9 pling on the leading energy sources of the interdecadal variability of the ocean-atmosphere system 10 is examined by means of a stochastically-forced idealized coupled model. The effect of coupling 11 is quantified from a comparison of the buoyancy variance budget of coupled and uncoupled model 12 configurations. The simplicity of the model allows us to contrast the effect of coupling between a 13 super-critical regime where the deterministic ocean dynamics drive the variability and a damped 14 regime where noise forcing is central to its existence. The results show that changes in surface 15 buoyancy fluxes act as a sink of temperature variance in the super-critical regime, and only be-16 come a source in the strongly damped regime. By contrast, internal ocean dynamics associated 17 with the interaction of transient buoyancy fluxes with mean buoyancy gradients always act as a 18 source of interdecadal variability. In addition to the reduced thermal damping effect in coupled 19 integrations, thermal coupling with the atmosphere is shown to significantly increase the role of 20 internal ocean dynamics in the variability, in particular in the regime where interdecadal modes 21 are damped. Only for oceanic background states in the strongly damped regime do changes in 22 surface buoyancy fluxes play a leading role in the upper ocean variability. A stochastically-forced 23 coupled box model is proposed that captures the basic effect of thermal coupling on atmospheric 24 and oceanic energy sources of variability. 25

The purpose of this study is to better understand the impact of SIGNIFICANCE STATEMENT: 26 ocean-atmosphere thermal coupling on the leading energy sources of Atlantic interdecadal vari-27 ability. Increasing our understanding of the physical mechanisms driving climate variability at 28 interdecadal timescales is important to improve climate prediction. We show that the effect of 29 ocean-atmosphere thermal coupling is to substantially increase the role of internal ocean dynam-30 ics in the low-frequency variability of the upper ocean heat content and sea surface temperature. 31 Atmospheric stochastic forcing only becomes the primary driver of the oceanic temperature vari-32 ability in the large dissipative limit, when internal ocean modes are strongly damped. 33

#### **1. Introduction**

Multidecadal variability of North Atlantic sea surface temperatures has been observed in both 35 the instrumental record (Delworth et al. 2007; Deser et al. 2010; Tung and Zhou 2013) and a num-36 ber of climatic proxy reconstructions (Mann et al. 1998; Gray et al. 2004; Chylek et al. 2011). 37 Atlantic Multidecadal Variability (AMV, Sutton et al. 2018) is characterized by alternating bas-38 inwide decadal-scale warming and cooling periods with maximum variance at subpolar latitudes. 39 Clement et al. (2015) suggested that the AMV could result from the thermodynamic response to 40 atmospheric stochastic forcing and that ocean circulation variations are unimportant. This mecha-41 nism seems however incompatible with the paleorecord that exhibits statistically significant multi-42 decadal periods above a red-noise background (e.g. Gray et al. 2004; Wang et al. 2017). O'Reilly 43 et al. (2016) showed instead that that ocean circulation variations are key to reproduce the sign 44 of the observed correlation between sea surface temperatures (SST) and the AMV. Zhang et al. 45 (2019) reviewed the observational and modelling evidence for the origins of the AMV and con-46 cluded that internal ocean dynamics is a key driver. The existence of such internal variability was 47 recently questioned by Mann et al. (2021) who found instead a prominent role of volcanic forcing 48 and anthropogenic aerosols. Which of internal ocean dynamics, atmospheric stochastic forcing 49 or external radiative forcing lies at the very origin of the observed AMV remains therefore much 50 debated. The instrumental record is unfortunately too short compared to the timescale of the vari-51 ability and too sparse in terms of spatial coverage to obtain a definitive answer to this question and 52 climate models have yet to reach a consensus (Zhang et al. 2019). The present study will solely 53 focus on the variability generated internally in the ocean-atmosphere system. Even in this con-54

text numerical models are inconclusive, in part because they show that the relative contributions 55 of the ocean and atmosphere to the variability are a strong function of poorly constrained critical 56 parameters, such as turbulent eddy diffusivities associated with unresolved scales (Arzel and Huck 57 2020). In addition ocean-atmosphere interactions certainly have an impact on the amplitude of 58 the variability, at least at low-frequencies (Barsugli and Battisti 1998), but the net effect of those 59 interactions on the leading energy sources of the variability have yet to be found. This study will 60 thus focus on the effect of thermal coupling on the primary drivers of the interdecadal variability 61 of the extratropical ocean-atmosphere system. 62

The effect of ocean-atmosphere coupling on the decadal or longer variability is traditionnaly 63 evaluated by comparing fully coupled ocean-atmosphere simulations to ocean-only experiments 64 where the surface heat, momentum and freshwater fluxes are diagnosed from the coupled run. Un-65 der such flux boundary conditions atmospheric damping on sea surface temperature (SST) anoma-66 lies is reduced to zero and the amplitude of oscillations is maximized. This approach has been used 67 in many modelling studies and across a large diversity of model configurations and complexities. 68 Delworth and Greatbatch (2000) and Gastineau et al. (2018) performed such experiments with 69 comprehensive coupled general circulation models (GCMs) to show that coupling with the atmo-70 sphere was not essential to the variability, and that atmospheric stochastic forcing associated with 71 mid-latitudes disturbances was necessary to sustain interdecadal oscillations against dissipation. 72 One drawback of such an approach is that the surface fluxes used to drive ocean-only integrations 73 contain information about the response of the atmosphere to SST changes. The comparison of 74 those ocean-only runs with coupled integrations does not therefore allow to isolate the effect of 75 the feedback of the atmosphere on the SST anomalies, but instead to assess the entire effect of 76 surface heat flux damping on SST anomalies. Another approach consists in forcing the ocean with 77 the time history of surface air temperatures (SAT) and winds diagnosed from an atmospheric-only 78 integration itself forced by climatological SSTs from the coupled simulation (Barsugli and Battisti 79 1998). These "uncoupled" ocean-only experiments therefore lack the feedback of the atmosphere 80 onto the SST anomalies. A direct comparison with the coupled runs thus allows to unambigously 81 isolate the basic impact of this feedback (i.e. thermal coupling) on the variability of the coupled 82 system. The absence of atmospheric feedback in uncoupled experiments makes the phasing of SST 83 and SAT anomalies less coherent, leading to a smaller persistence of SST anomalies in uncoupled 84

than in coupled integrations. The basic effect of thermal coupling is therefore to enhance the 85 temperature variance in both the ocean and atmosphere, and more specifically at low-frequencies 86 (Barsugli and Battisti 1998). Farneti and Vallis (2009) performed uncoupled atmosphere-only ex-87 periments in the context of interdecadal climate variability and reached the same conclusions. The 88 spectral peak at interdecadal periods in the SAT was shown to be only present when prescribed, 89 time-varying, SST from the coupled run were used, indicating an influence of either an oceanic 90 interdecadal mode or some form of coupled interaction. Wu and Liu (2005) run an uncoupled 91 experiment with their realistic coupled model to show the critical role of coupling in sustaining 92 North Atlantic Ocean decadal variability. 93

Arzel and Huck (2020) used a realistic ocean general circulation model forced by combination of 94 steady surface buoyancy and momentum fluxes and North Atlantic Oscillation-related stochastic 95 forcing to show that internal ocean dynamics plays a leading role in the growth of SST variance in 96 the super-critical regime (i.e. the regime where the multidecadal variability develops without any 97 noise forcing, at relatively low diffusivities), but that the atmospheric stochastic forcing represents 98 the major energy source for the variability in the damped regime (i.e. the regime where the vari-99 ability requires some atmospheric noise to emerge, at relatively high diffusivities). Therefore the 100 leading energy source of the variability is a strong function of critical parameters (such as eddy dif-101 fusivities) and background climate conditions, at least in the forced case. How ocean-atmosphere 102 thermal coupling modifies this behaviour remains to be determined. The present study will assess 103 the impact of thermal coupling on the leading energy sources of the variability and is therefore a 104 natural extension of Arzel and Huck (2020) who used an ocean-only configuration. The aim of the 105 present study is thus to progress in the understanding of the physical processes driving Atlantic 106 climate variability at interdecadal timescales. The main purpose is to quantify the impact of ocean-107 atmosphere thermal coupling on the *drivers* of temperature variance at interdecadal timescales. As 108 such the present study is also a natural extension of Barsugli and Battisti (1998) who focused their 109 analysis on the impact of coupling on the temperature variance of the ocean-atmosphere system. 110 A joint objective is to determine how thermal coupling impact the energy sources of the variability 111 in both the super-critical and damped regimes. Different variability regimes will be obtained by 112 systematically varying the magnitude of the turbulent eddy diffusivity K which has been shown 113 to be one of the most critical parameter to the interdecadal variability of the overturning circula-114

tion (Huck et al. 1999a). For each value of K the effect of thermal coupling is inferred from a 115 comparison of coupled, uncoupled and forced integrations. Given the relatively large number of 116 millenial-scale experiments required to achieve this task a computationally efficient coupled model 117 is required. Following the views Held (2005) we believe that the use of a model hierarchy such 118 as the one proposed here is necessary to help gain understanding of the mechanisms driving cli-119 mate variability at interdecadal timescales. The model comprises a planetary geostrophic oceanic 120 component coupled to a stochastically-forced energy balance model. The geometry is idealized 121 with a two-hemisphere sector of sphere and a southern periodic channel. The model captures the 122 large-scale features of the Atlantic circulation and for some parameter range exhibit interdecadal 123 oscillations under deterministic forcing conditions. The cause of the variability was shown to be 124 a large-scale baroclinic instability of the mean flow driving interdecadal oscillations with the os-125 cillation period set by the zonal transit time of long baroclinic planetary waves (Colin de Verdière 126 and Huck 1999; te Raa and Dijkstra 2002). This mode of variability has been shown to persist in 127 realistic configurations of ocean-only (Sévellec and Fedorov 2013; Arzel et al. 2018) and coupled 128 models (Ortega et al. 2015; Muir and Fedorov 2016; Gastineau et al. 2018). 129

The outline of this paper is as follows. Section 2 provides a description of the idealized cou-130 pled ocean-atmosphere model as well as the experimental procedure used to quantify the effect 131 of ocean-atmosphere thermal coupling. The main characteristics of the variability are presented 132 in section 3, and its sensitivity to horizontal turbulent diffusivity is presented in section 4 for 133 the coupled, uncoupled and forced cases. Section 5 quantifies the leading energy sources of the 134 interdecadal variability and their sensitivity to K using an approach based on the buoyancy vari-135 ance budget. The model hierarchy is finally extended to a coupled stochastically-forced ocean-136 atmosphere dynamical system in section 6 that reproduces with very high fidelity the results ob-137 tained with the 3D model. The results are then summarized and discussed in section 7. 138

# **2. Model and experiments**

# 140 *a. The coupled model*

The 3D ocean model is based on the planetary geostrophic equations, valid for time scales much larger than the inertial period and spatial scales much larger than the internal Rossby radius of deformation (Salmon 1986; Colin de Verdière 1988). The domain is a flat-bottom sector of a



FIG. 1. (a) Zonal (solid) and meridional (dashed) surface wind-stress. (b) Imposed freshwater forcing (evaporation minus precipitation), (c) spatial pattern of the stochastic forcing (c.i.  $10 \text{ W m}^{-2}$ ) entering the atmospheric thermodynamic balance. The amplitude in the subpolar gyre ( $100 \text{ W m}^{-2}$ ) is twice that of the subtropical gyre ( $50 \text{ W m}^{-2}$ ) similar to Herbaut et al. (2002). The sign change in the stochastic forcing roughly occurs at the intergyre boundary at  $47^{\circ}$ N.

sphere, with dimensions appropriate for the Atlantic: it extends from 66°S to 74°N and is 64° 144 wide in zonal extent. A zonal subpolar channel analogous to the Drake Passage in the Southern 145 Ocean is represented by applying cyclic boundary conditions between 66°S and 52°S throughout 146 the entire water column. Lateral boundaries are solid vertical walls where no-slip and no-flux 147 boundary conditions are applied. The horizontal resolution is 2° and there are 15 layers whose 148 vertical thickness increases unevenly from 50 m at the surface to 550 m at the bottom (4500 149 m depth). Static instability is removed by instantaneously restoring vertical density profiles to 150 neutrality. The vertical mixing is fixed at  $10^{-4}$  m<sup>2</sup>s<sup>-1</sup>, in agreement with values inferred from the 151 large-scale abyssal stratification (Munk and Wunsch 1998). The horizontal Laplacian viscosity is 152 fixed at 10<sup>5</sup> m<sup>2</sup> s<sup>-1</sup>. A linear equation of state is used with constant thermal expansion and haline 153 contraction coefficients. The surface wind-stress forcing is distributed over the surface layer (50 m 154 depth), it is purely zonal and follows the analytical profile function of latitude proposed by Weaver 155 and Sarachik (1990) (Fig. 1a). 156

<sup>162</sup> The atmospheric model is a standard, two-dimensional, dry energy balance model. It closely <sup>163</sup> follows the scheme developed by Fanning and Weaver (1996). The freshwater forcing is imposed <sup>164</sup> and follows the latitudinal profile shown in Fig. 1b. The atmospheric eddy temperature diffusivity <sup>165</sup>  $K_a$  is uniform and fixed to  $1.3 \times 10^6$  m<sup>2</sup>s<sup>-1</sup>. The atmosphere-ocean heat exchange coefficient  $\lambda$  falls within the range of observed values and is fixed to 30 W m<sup>-2</sup> K<sup>-1</sup> is all experiments. Both the incoming solar radiation *S* at the top of the atmosphere and the planetary albedo  $\alpha_p$  follow annual mean latitudinal profiles given by North et al. (1981) and Graves et al. (1993), respectively. In the absence of atmospheric dynamics, stochastic forcing is introduced in the atmospheric temperature equation as

$$Q_N(x, y, t) = \sigma_N Q_{NAO}(x, y)\zeta(t), \tag{1}$$

This term parameterizes the stochastic forcing associated with the divergence of eddy temperature 171 fluxes which are typically enhanced at mid-latitudes. There the observed transient sensible and 172 latent heat fluxes amount to about P = 3.5 PW (Peixoto and Oort 1992). Using an eddy length 173 scale  $L_{eddy}$  of 1000 km and a zonal length scale  $L_x$  of 25,000 km for the length of a latitude 174 circle at mid-latitudes, we obtain  $\sigma_N = P/L_x L_{eddy} = O(100)$  W m<sup>-2</sup> for the amplitude of the eddy 175 forcing. We then specify  $\sigma_N = 100 \text{ W m}^{-2}$  in all stochastically-forced experiments. The pattern 176  $Q_{NAO}$  is similar to that used by Herbaut et al. (2002). It mimicks the observed North Atlantic 177 Oscillation pattern with a dipolar structure centered at mid-latitudes and with maximum values 178 in the west of the basin. The random discrete timeseries  $\zeta(t)$  has been built from a first order 179 auto-regressive process with a decorrelation timescale of 10 days. This timescale corresponds to 180 estimates inferred by Feldstein (2000) using daily means from the NCEP-NCAR reanalysis. To 181 assess the effect of noise coherence additional experiments using a white noise forcing and a longer 182 temporal coherence (30 days) have been conducted. The noise forcing has a sampling frequency 183 (0.05 days) corresponding to the oceanic timestep of the model. The variance of  $\zeta(t)$  is 1. Note 184 finally that the stochastic timeseries are stricly identical in all stochastically-forced integrations. 185

## 186 b. Experimental design

Three different model configurations (coupled, uncoupled, forced) are used to assess the impact of ocean-atmosphere thermal coupling on the variability. For each of those three configurations the model is run with and without atmospheric stochastic forcing. The latter case is said to be deterministic. This procedure is repeated for a wide range of oceanic horizontal eddy diffusivities *K* between 500 and 2600 m<sup>2</sup>s<sup>-1</sup>. Those values of *K* span the observed range of eddy diffusivities but do not attempt to capture the strong spatial variations (Abernathey and Marshall 2013). This approach allows us to explore the effect of ocean-atmosphere coupling in two different variability regimes, namely a super-critical regime where the variability spontaneously develops without atmospheric stochastic forcing and a damped regime where the variability requires some noise to emerge. Deterministic runs are integrated for 5000 years and are initialized from a resting ocean with uniform temperature (4°C) and salinity (35 psu). Stochastic runs are integrated for 1000 years and are initialized from the end state of deterministic runs.

199

(1) **Coupled**. The model is integrated in its standard coupled mode (denoted by superscript C) where the atmospheric temperature  $T_a$  and the surface buoyancy (heat and freshwater) flux  $B_O$ toward the ocean are given by

$$C_a dT_a^C / dt = F(T_a^C, T_o^C) + Q_N,$$
<sup>(2)</sup>

$$B_O^C = G(T_a^C, T_o^C), \tag{3}$$

where  $C_a$  is the atmospheric heat capacity and  $T_o$  the sea surface temperature. The operators *F* and *G* include both turbulent and radiative fluxes,

$$F(T_a, T_o) = \lambda(T_o - T_a) + \epsilon_o \sigma T_o^4 - \epsilon_a \sigma T_a^4 - \epsilon_p \sigma T_a^4 + S a_s (1 - \alpha_p) + K_a \nabla^2 T_a,$$
(4)

$$B_O(T_a, T_o) = \lambda (T_a - T_o) - \epsilon_o \sigma T_o^4 + \epsilon_a \sigma T_a^4 + S (1 - a_s)(1 - \alpha_p)$$
(5)

where  $\sigma$  is the Stefan-Boltzmann constant,  $\epsilon_o = 0.96$  is the longwave emissivity of the surface,  $\epsilon_a = 0.85$  is the longwave emissivity of the atmosphere,  $\epsilon_p = 0.63$  is the planetary emissivity to space,  $a_s = 0.3$  is the shortwave absorptivity. The terms on the right hand side of (4) are, from first to last, the sensible heat exchange with the surface, the upward longwave emission from the surface, the downward longwave emission from the atmosphere, the outgoing longwave radiation to space, the absorbed solar radiation within the atmosphere and the diffusive heat transport. The last term on the right hand side of (5) is the absorbed solar flux at the surface.

212

<sup>213</sup> (2) **Uncoupled**. (a) The atmosphere is first run with the fixed 100 years averaged SST distribution <sup>214</sup> diagnosed from deterministic coupled integrations. This SST field is defined here as the climato-<sup>215</sup> logical SST field  $T_o^{clim}$ . (b) The ocean model is then forced by the surface fluxes computed using <sup>216</sup> the time history of atmospheric temperatures from phase (a) and the actual SST field from the <sup>217</sup> uncoupled integration. Those two phases are run synchronously with the same coupling frequency <sup>218</sup> as the coupled model (0.05 days). Hence

$$C_a dT_a^U / dt = F(T_a^U, T_o^{clim}) + Q_N,$$
(6)

$$B_O^U = G(T_a^U, T_o^U), \tag{7}$$

where the superscript U stands for uncoupled. In those uncoupled runs the atmospheric temper-219 ature can only change in response to the stochastic forcing, all other forcings including the SST 220 field being kept constant in time. The atmosphere therefore does not see the SST variations that 221 can occur in response to either atmospheric stochastic forcing (if any, embedded in  $T_a^U$ ) or changes 222 due to the existence of an intrinsic interdecadal ocean mode. In the coupled integrations by 223 contrast the atmosphere can vary in response to both atmospheric stochastic forcing and changing 224 SSTs. The comparison of the coupled and uncoupled runs therefore allows to unambigously 225 isolate the impact of the atmospheric feedback on SST anomalies. 226

227

<sup>228</sup> (3) **Forced**. The main motivation behind these ocean-only forced experiments is to determine <sup>229</sup> whether the variability obtained in the coupled model can be interpreted as a manifestation of a <sup>230</sup> self-sustained ocean mode or as an excitation of a damped ocean mode. The procedure here is <sup>231</sup> very similar to that used for building uncoupled runs except for the previous phase (b) where the <sup>232</sup> surface heat flux to the ocean  $B_O^F$  is now computed using the climatological SST  $T_o^{clim}$  rather than <sup>233</sup> the actual SST obtained in those ocean-only integrations,

$$B_O^F = G(T_a^U, T_o^{clim}),\tag{8}$$

where the superscript F stands for forced. Because surface heat fluxes are completely decorrelated from SST anomalies in those forced experiments, surface heat flux damping on SST anomalies is reduced to zero and the amplitude of the variability is maximized. Forced experiments therefore lack thermal damping on SST anomalies. Uncoupled experiments by contrast lack thermal coupling and have increased surface heat flux damping on SST anomalies and subsequently a much weaker variability compared to either the forced or coupled runs. As such the forced runs have a



FIG. 2. (top) Time-mean meridional overturning circulation (MOC) in the stochastic coupled integration for  $K = 1000 \text{ m}^2 \text{s}^{-1}$ . (bottom) Time-series of the MOC index in the stochastic experiments. The MOC index is defined as the maximum of the meridional overturning streamfunction north of 30°N and below 850m. The stochastic experiments are run for 1000 years and start from the end state of the corresponding deterministic solutions. These later solutions are obtained from a 5000 years long integration which was necessary to reach statistical equilibrium.

thermal damping on SST anomalies that is at the exact opposite of what is expected when thermal
 coupling with the atmosphere is suppressed.

#### 242 **3. Interdecadal variability**

In what follows we focus on the stochastic integrations and contrast the variability patterns obtained for the coupled, uncoupled and forced systems. The comparison is made for the canonical horizontal eddy diffusivity value  $K = 1000 \text{ m}^2 \text{s}^{-1}$ .

Figure 2 shows the time-mean meridional overturning streamfunction in the stochastic coupled run, as well as the timeseries of the MOC index over a 1000 years period for the coupled, uncoupled and forced systems. The MOC index is defined as the maximum value of the meridional overturning streamfunction north of 30°N and below 850m. Despite the relative simplicity of the model, in terms of both geometry and physics, the observed North Atlantic meridional overturning



FIG. 3. SST standard deviations diagnosed from the annual mean ouput of SST anomalies over a 100 years period for the stochastic case with  $K = 1000 \text{ m}^2 \text{s}^{-1}$  for the coupled (left), uncoupled (middle) and forced (right) configurations.

circulation is well reproduced with a thermally direct cell intensified in the Northern Hemisphere 260 with a peak value close to that observed (18 Sv, Colin de Verdière et al. 2019), and a weaker (5 261 Sv) thermally-indirect cell at depth reminiscent of the Antarctic Bottom Cell. Quasi-regular oscil-262 lations with interdecadal periods ( $\sim 25$  years) are found in all cases, with peak-to-peak amplitudes 263 reaching 38% (56%) of the mean transport in the coupled (forced) case, but only 12% in the un-264 coupled system. Thermal coupling with the atmosphere, as inferred from the comparison between 265 the coupled and uncoupled systems, therefore acts as a strong amplifier of interdecadal oscilla-266 tions in agreement with Farneti and Vallis (2009) and Wu and Liu (2005). As originally stated 267 by Barsugli and Battisti (1998), coupling with the atmosphere makes SST and SAT anomalies to 268 vary more in phase in the coupled case, thereby reducing the damping on SST anomalies com-269 pared to the uncoupled case. In the forced case the damping of SST anomalies is at its minimum 270 (but non-zero because of the presence of mixing processes) leading to much greater variability. 271 This feature can be seen in the standard deviations of SST anomalies in Fig. 3. For all cases 272 SST variance is maximum in the northwest corner of the basin and is about three times greater 273 in the coupled compared to the uncoupled case. There is virtually no variability in the Southern 274 Hemisphere: the presence of the southern channel induces a relatively weak upper circulation that 275

does not favor the baroclinic instability mechanism (Arzel et al. 2007). The spectral peaks (Fig. 276 4) in both the MOC index and SST anomalies (computed as the weighted average in the region 277 of maximum SST variance, 70-50°E, 46°N-74°N) greatly exceed the red-noise background (but 278 barely for the uncoupled case which just passes the 95% confidence level) indicating that inter-279 decadal oscillations cannot be simply interpreted as an integration of the atmospheric noise by the 280 oceanic mixed layer. The first and second leading patterns of variability have been estimated in 281 terms of Empirical Orthogonal Functions (EOFs) and are presented in Fig. 5. The patterns in the 282 coupled and forced systems are very similar but with a greater amplitude for the latter. The first 283 EOF features a monopole intensified in the northwestern corner of the basin, whereas the second 284 EOF features a monopole of oppositve sign centered in the subtropics and with a much weaker 285 amplitude. These two EOFS vary in quadrature (not shown) with the first EOF leading the second 286 one by about 6 years, thereby covering a full cycle in about 25 years. A lagged regression analysis 287 between SST anomalies and the MOC index highlights the emergence and growth of perturbations 288 in the eastward extension of the western boundary current in the northwestern corner of the basin 289 (not shown). The perturbations then propagate westward toward the western boundary where they 290 subduct and quickly dissipate (not shown). This spatio-temporal organization of the variability is 291 similar to several previous idealized ocean-only model studies using fixed flux boundary condi-292 tions without noise forcing (Greatbatch and Zhang 1995; Colin de Verdière and Huck 1999; te Raa 293 and Dijkstra 2002). In the uncoupled case, the NAO signature along with its dipolar structure is 294 clearly evident in the first EOF of SST anomalies (Fig. 5). Therefore thermal coupling does not 295 only increase the amplitude of the mode but also significantly alter the leading pattern of the vari-296 ability, at least for this specific value of  $K = 1000 \text{ m}^2 \text{s}^{-1}$ . The second EOF of the uncoupled case 297 exhibits strong similarities with the leading EOF of the coupled and forced cases, suggesting an 298 implication of internal ocean dynamics in the uncoupled variability. 299

# **4. Sensitivity to** K

Figure 6 presents the sensitivity of some key quantities to the horizontal diffusivity K in the coupled, uncoupled and forced systems, and under both deterministic and stochastic forcing conditions. We first see that the strength of the mean overturning increases with K in all cases. This behaviour sharply contrasts with results obtained with models where the eddy-induced diffusivity



FIG. 4. Power spectra of (top) the MOC index (Sv<sup>2</sup> years) and (bottom) SST anomalies (°C<sup>2</sup> years) timeseries averaged in the northwest corner (70-50°W and north of 46°N) in the stochastic experiments with K = 1000 $m^2s^{-1}$  and for the coupled (blue), uncoupled (red) and forced (black) systems. The calculation uses a multitaper technique with 3 tapers and is based on 1000 years of annual mean model output. The smooth dashed lines indicate the red noise 95% confidence level.

rather the horizontal diffusivity is varied (Marshall et al. 2017; Arzel and Huck 2020). The reason for this discrepancy may be caused by the "Veronis effect" whereby horizontal diffusion induces large diapycnal fluxes once the isopycnals are tilted by coastal upwellings, in particular along the western boundary (Veronis 1975; Huck et al. 1999b). Quite remarkably the strength of the mean overturning for a given value of *K* keeps nearly the same value whatever the configuration used (coupled, forced, uncoupled) and irrespective of the stochastic forcing being present or not. Hence rectification of the mean flow by stochastic forcing does not occur in our simulations as opposed



FIG. 5. First (top) and second (bottom) EOFs of annual mean SST anomalies along with the explained variance (%) in the stochastic experiments with  $K = 1000 \text{ m}^2 \text{s}^{-1}$  and for the coupled (left), uncoupled (middle) and forced (right) cases.

to Frankcombe and Dijkstra (2009). This implies that the different variability characteristics associated with the coupled, uncoupled and forced systems, in particular when stochastic forcing is present, cannot be ascribed to modifications of the mean circulation.

The transition from a steady circulation to an interdecadal oscillatory behaviour was shown to occur through a Hopf bifurcation as the horizontal diffusivity (Colin de Verdière and Huck 1999) or ocean-atmosphere heat exchange coefficient (Chen and Ghil 1996) is decreased. Here Hopf bifurcations under deterministic conditions are successively found at K = 210, 1140 and 1560  $m^2s^{-1}$  for the uncoupled, coupled and forced cases respectively (note that only the last two bifurcations are shown in Fig. 6). This sequence of bifurcations is consistent with the magnitude of

atmospheric damping acting on SST anomalies which is at its largest in the uncoupled case and at 329 its lowest in the forced case, as discussed in Section 2b. The suppression of variability for large 330 K is not caused by changes in the circulation, which favors increasingly baroclinically unstable 331 situations as K increases (Fig. 6a), but instead by the direct damping effect of dissipation on per-332 turbations. When stochastic forcing is present interdecadal variability exists over the full range of 333 diffusivities explored here, but only as long as the noise forcing has some temporal coherence, in 334 agrement with Frankcombe and Dijkstra (2009) and Arzel and Huck (2020). Additional spectral 335 analyses of the MOC and SST (averaged in the region  $70 - 50^{\circ}$ W,  $46 - 74^{\circ}$ N) indices show that the 336 interdecadal variability is always statistically significant in both fields, even for the most diffusive 337 cases. The effect of stochastic forcing on the variability is large in the damped regime (for diffusiv-338 ity values larger than the critical value at bifurcation) and much weaker in the super-critical regime 339 (for diffusivity values weaker than the critical value at bifurcation). The comparison of coupled 340 and uncoupled configurations of the model clearly shows that thermal coupling between the ocean 341 and atmosphere increases the temperature variance in both fluids, in agreement with Barsugli and 342 Battisti (1998), as well as the amplitude of ocean circulation changes. The novel aspect is that the 343 amplifying effect of thermal coupling on the variability is much stronger in the super-critical than 344 in the damped regime. For SST for instance, thermal coupling typically quadruples the amplitude 345 of the variability in the super-critical regime (from 0.55 to  $2.31^{\circ}$ C for  $K = 500 \text{ m}^2 \text{s}^{-1}$ ) but less than 346 doubles it in the damped regime (from 0.28 to 0.45°C for  $K = 2600 \text{ m}^2 \text{s}^{-1}$ ). Let us finally mention 347 that the amplitude of the changes in surface heat flux (diagnosed from a composite analysis identi-348 cal to that used for SST and SAT in Fig. 6) in the region of maximum SST variance are relatively 349 independent of K and amount to about 25 W m<sup>-2</sup> (not shown), an amplitude comparable to the 350 observed annual mean changes in turbulent fluxes associated with the NAO (e.g. Fig. 1 in Arzel 351 and Huck 2020). 352

# **5. Energy sources**

We now turn to the analysis of the effect of thermal coupling on the leading energy sources of the variability. These energy sources are identified as the major terms driving the growth of buoyancy variance in the ocean against all sources of damping. As such the buoyancy variance budget provides a quantitative estimate of the contribution of the ocean and atmosphere to the



Fig. 6. Statistics of key indices as a function of the horizontal eddy diffusivity K under deterministic and 353 stochastic forcing conditions for the coupled, uncoupled and forced cases. (a) Time-mean MOC index. The 354 index is defined as the maximum value of the meridional overturning streamfunction north of 30°N and below 355 850m. (b) Amplitude of MOC variations (Sv). (c) Amplitude of SST changes (°C) averaged in the north-356 western corner  $(70 - 50^{\circ}\text{W}, 46 - 74^{\circ}\text{N})$ , which corresponds to the region of largest temperature changes in all 357 experiments. (d) Same as (c) but for the SAT (°C). The amplitude of the variability in (b-c-d) has been estimated 358 from a composite analysis of the last 1000 years of each experiment where the max (min) of the timeseries is 359 computed as the time average of the values greater (smaller) than the mean plus (minus) one standard deviation. 360 Note that only the changes greater than zero are shown. The uncoupled deterministic system does not have 361 variability in the range of diffusivity values shown here. When the stochastic forcing is white (orange dotted 362 line) the amplitude of the variability in the damped regime is greatly reduced compared to a situation where a 363 noise forcing has a finite temporal coherence. 364

growth of perturbations. Such an approach has been previously and successfully applied to the interdecadal climate variability problem in either stand-alone ocean models (Colin de Verdière and Huck 1999; Arzel et al. 2006, 2018) or coupled models (Arzel et al. 2007, 2012; Buckley et al. 2012; Jamet et al. 2016; Gastineau et al. 2018) with complexities ranging from idealized to fully coupled and realistic.

#### 375 a. Method

<sup>376</sup> The linearized buoyancy variance equation is given by

$$\frac{1}{2}\frac{\overline{\partial b'^2}}{\partial t} = -\overline{\mathbf{u}'_{\mathbf{h}}b'}.\nabla_h \overline{b} - \overline{w'b'}\partial_z \overline{b} - \frac{1}{2}\overline{\mathbf{u}}.\nabla \overline{b'^2} + \overline{b'Q'_b} + \overline{b'D'_b},\tag{9}$$

where the overbar denotes a time average and the prime the perturbation. Integrated over the do-377 main and weighted appropriately, it provides an equation for the available potential energy in the 378 quasi-geostrophic approximation (Huang 1998). The third-order term associated with advection 379 of buoyancy variance by the disturbed flow is between one and three orders of magnitude smaller 380 than  $-\overline{\mathbf{u}'b'}$ .  $\nabla \overline{b}$  for all values of K (not shown) and has been dropped during the linearization pro-381 cedure. The first term on the right hand side is a source of buoyancy variance when transient 382 buoyancy fluxes  $\overline{\mathbf{u}'_{\mathbf{h}}b'}$  are oriented down the mean buoyancy gradient, where  $\mathbf{u}_{\mathbf{h}}$  is the horizontal 383 Eulerian velocity. This configuration is typical of baroclinic instability for which potential energy 384 is extracted from the mean stratification. This term has been pinpointed as the primary source of 385 the variability in many ocean-only and coupled models (see references above). Baroclinic insta-386 bility is associated with a conversion of potential to kinetic energy of perturbations through the 387 positive exchange term  $\overline{w'b'}$ . Under such unstable conditions, the second term in (9) is always 388 negative (provided  $\partial_z \overline{b} > 0$  in stably stratified waters) and is therefore a sink of buoyancy variance. 389 The third term represents the spatial redistribution of buoyancy variance by the three dimensional 390 background flow  $\overline{\mathbf{u}}$ . It plays an important role at the regional scale by decreasing or increasing the 391 variance, but cannot be at the very origin of the variability at the basin scale since its global average 392 is identically zero. The fourth term is a source of buoyancy variance when the surface buoyancy 393 anomalies and the surface buoyancy flux anomalies  $Q'_{h} = g_{0}\alpha_{T}Q'/C_{0}$  (with  $g_{0}$  the acceleration of 394 gravity at the sea surface,  $C_0$  the specific heat capacity of the forcing layer and Q' the anomalous 395 surface heat flux) are positively correlated. The dissipation term  $\overline{b'D'_{b}}$ , which contains contribu-396 tions from horizontal and vertical mixing processes (including convective mixing), is a sink of 397 buoyancy variance at the basin scale and will not be considered. In what follows, we will restrict 398 our attention to the two only terms in (9) that can take positive values at the basin scale, namely 399 the atmospheric energy source associated with surface buoyancy fluxes anomalies and the oceanic 400

energy source associated with the interaction of transient buoyancy fluxes with mean horizontal
 buoyancy gradients.

# 403 b. Results

The buoyancy variance budget is applied to the northwestern corner of the basin (70-50 $^{\circ}$ W, 46-404 74°N) which roughly coincides with the region of maximum temperature (buoyancy) changes in 405 all experiments (Fig. 3). The buoyancy changes are largely controlled by temperature variations so 406 that the buoyancy variance budget presented hereafter is to a large extent identical to the temper-407 ature variance budget (not shown). We define the quantities  $S_A = \langle \overline{b'Q'_h} \rangle$ ,  $S_O = -\langle \overline{u'_h b'}, \nabla_h \overline{b} \rangle$ 408 where the brackets denote volume averaging. The analysis entirely focuses on the budget within 409 the uppermost 250 m of the ocean model. Because our region of averaging encompasses the bulk 410 of the variability the advection of buoyancy variance into or out of the region by the mean flow is 411 always an order of magnitude smaller than  $S_O$  (not shown). 412

Figure 7ab presents the sensitivity of  $S_O$  and  $S_A$  to K for the coupled, uncoupled, and forced 425 systems, and under both deterministic and stochastic forcing conditions. We first see that  $S_{O}$  is 426 positive for all model configurations and for all values of K. Consistent with the amplitude of 427 the variability, the largest values of  $S_{O}$  are obtained for the forced system, and the lowest values 428 for the uncoupled one. The oceanic energy source term in the coupled case is about an order of 429 magnitude larger than that in the uncoupled one. The effect of thermal coupling is therefore to 430 significantly increase the creation of buoyancy variance by internal ocean dynamics for all values 431 of K, with the strongest impact in the super-critical regime (Fig. 7c). 432

Focusing now on  $S_A$  in the coupled case, we see that this term is strongly negative in the super-433 critical regime and even beyond (up to  $K = 1400 \text{ m}^2 \text{s}^{-1}$ , Fig. 7b). In this range of K values, surface 434 heat flux anomalies therefore act to damp the variability. For larger K values,  $S_A$  becomes positive, 435 which along with S<sub>O</sub> contributes to the production of buoyancy variance in the northwestern corner 436 of the domain. The change in the sign of  $S_A$  with K can be rationalized as follows. Density 437 anomalies are largely controlled by temperature changes so that  $b' \propto T'$ . Using this result, and 438 assuming that the upwelling and downwelling longwave radiative fluxes nearly cancel out, the 439 covariance term between the surface buoyancy flux anomalies and the buoyancy anomalies can be 440 approximated as  $S_A \propto \langle \lambda(\overline{T'T'_a} - \overline{T'^2}) \rangle$ , where  $\lambda$  is the air-sea heat exchange coefficient (recall 441



Fig. 7. Buoyancy variance budget in the northwestern corner of the domain (70-50°W, 46-74°N) and in the 413 uppermost 250m as a function of the horizontal diffusivity K. Top panels show the oceanic  $S_O = -\langle \overline{\mathbf{u}'_h b'}, \nabla_h \overline{b} \rangle$ 414 (left) and atmospheric  $S_A = \langle \overline{b' Q'_h} \rangle$  (right) energy source terms under both deterministic and stochastic bound-415 ary conditions and for the coupled, uncoupled and forced systems. The uncoupled case under deterministic 416 forcing conditions does not exhibit variability in the range of diffusivities considered here which explains why 417 it is absent. The bottom left panel shows the changes (coupled minus uncoupled) in atmospheric and oceanic 418 source terms caused by coupling with the atmosphere. Note that a log vertical scale is used for both posi-419 tive and negative values in the upper and bottom left panels. The bottom right panel shows the contribution 420  $R = S_O/(S_O + S_A^+)$  (%) of internal ocean dynamics to the production of buoyancy variance in the northwestern 421 corner of the domain for the coupled  $(R_C)$ , uncoupled  $(R_U)$  and forced  $(R_F)$  cases, with  $S_A^+ = S_A$  if  $S_A > 0$  and 0 422 otherwise (see text for details). In all panels the vertical dashed lines indicate the position of the Hopf bifurcation 423 of the coupled (blue,  $K = 1140 \text{ m}^2\text{s}^{-1}$ ) and forced (black,  $K = 1560 \text{ m}^2\text{s}^{-1}$ ) cases. 424

that both the solar and freshwater forcings are steady). Now in the super-critical regime, SST anomalies are relatively large implying that the covariance term  $S_A$  is dominated by  $-\lambda < \overline{T'^2} >$  which is always negative. Larger negative values of  $S_A$  are obtained in the coupled case in the super-critical regime because of the much greater oceanic temperature variance  $\overline{T'^2}$  compared to the uncoupled case. In the damped regime, SST anomalies have a much weaker magnitude and the leading term in  $S_A$  is  $\lambda < \overline{T'T'_a} >$  which is positive since both SST and SAT anomalies tend to vary in phase. In the uncoupled system, the same behaviour occurs but the change in  $S_A$  occurs for lower values of K, consistent with the lower SST variance compared to the coupled case.

Interestingly the change in the sign of  $S_A$  in the uncoupled case occurs near the Hopf bifurcation of the coupled system. Whether this occurs by pure coincidence or not remains to be found. Consistent with the larger SST variance in the coupled system compared to the uncoupled one, thermal coupling acts to increase the sink of buoyancy variance associated with surface fluxes across a large range of *K* values, up to 1700 m<sup>2</sup>s<sup>-1</sup> beyond the bifurcation at 1140 m<sup>2</sup>s<sup>-1</sup>, but increase the production of buoyancy variance for the most diffusive cases (Fig. 7c).

In the forced case surface flux anomalies are always a source of buoyancy variance (Fig. 7). 456 We also note that  $S_A$  in this case is nearly independent on K despite the significant changes in the 457 amplitude of the variability between the super-critical and damped regimes. An explanation for 458 this behaviour was provided by Arzel and Huck (2020). In the damped regime, the kinetic energy 459 variability is much weaker than in the super-critical regime. Low anomalous oceanic advection 460 tends to keep the noise-forced SST anomalies in the forcing region, a process that favors relatively 461 high correlations between the forcing and the SST field. The increase of the correlation with K 462 compensates for the decrease in the SST variance leading to an almost unchanged covariance term 463  $S_A$  as K increases. Compared to the forced system, the coupled system decreases the production 464 of buoyancy variance associated with both the surface fluxes and internal ocean dynamics. 465

Which of the ocean or atmosphere represents the major energy source for the variability? How 469 does the contribution of the atmosphere and ocean to the production of temperature variance de-470 pend on thermal coupling and oceanic eddy diffusivities? To answer these questions we now 471 compute the ratio  $R = S_O/(S_O + S_A^+)$  where  $S_A^+ = S_A$  if  $S_A > 0$  and equal to 0 otherwise. This 472 ratio thus provides the fraction of the production of buoyancy variance associated with  $S_O$  and  $S_A$ 473 explained by internal ocean dynamics. If  $S_A \leq 0$  internal ocean dynamics (S<sub>O</sub>) explains 100% of 474 the production of the buoyancy variance associated with the ensemble  $S_{O}$  and  $S_{A}$ . The results are 475 presented in Fig. 7d. Quite clearly the leading role of internal ocean dynamics in the variability ex-476



FIG. 8. Leading patterns of SST variability (°C) and explained variance (%) in the damped regime shown here for the specific diffusivity value  $K = 1400 \text{ m}^2 \text{s}^{-1}$ . Shown are the first two EOFs and the first EOF in the stochastic coupled and uncoupled systems, respectively.

tends beyond the super-critical regime to a much larger range of K values when thermal coupling 477 with the atmosphere is enabled. Of course stochastic forcing is required to generate variability 478 in the damped regime, but the growth of upper ocean heat content (250 m) anomalies is however 479 mostly explained by changing ocean currents rather than by the direct thermodynamical response 480 of the mixed layer to the noise forcing. Importantly this behaviour occurs despite the fact that the 481 leading pattern of the variability in the damped regime bears some resemblance with the imposed 482 NAO forcing (Fig. 8). The imprint of the internal ocean mode in this damped regime is clearly 483 seen in the second EOF of SST anomalies (Fig. 8) which appears to be similar to the leading 484 pattern obtained in the super-critical regime (Fig. 5). We further note that the pattern correlation 485 between the coupled and uncoupled systems in the damped regime is high (Fig. 8), a behaviour 486 that sharply contrasts with that obtained in the super-critical regime (Fig. 5). In the forced config-487 uration, internal ocean dynamics is the primary driver of the upper ocean heat content (0-250 m) 488 variability for all values of K (Fig. 7d). 489

The results obtained here have been obtained using a very specific timescale of noise coherence of 10 days. Additional experiments not presented here reveal that increasing this timescale to 30 days has the effect of increasing the production of buoyancy variance by both surface fluxes ( $S_A$ ) and internal ocean dynamics ( $S_O$ ), the effect being much more pronounced in the damped regime.



FIG. 9. Contribution  $R = S_O/(S_O + S_A^+)$  (%, see text for details) of internal ocean dynamics to the production of buoyancy variance in the northwestern corner of the domain (70-50°W, 46-74°N) as a function of the horizontal diffusivity *K*. The calculation is done for the stochastic coupled ( $R_C$ ) case only and in the upper 50, 250 and 1150m. The vertical dashed (dotted) line indicates the corresponding position of the Hopf bifurcation at K =1140 m<sup>2</sup>s<sup>-1</sup> whereas the horizontal dashed line corresponds to the pivotal value where  $S_A = S_O$ .

As a whole the increase in  $S_A$  is larger than the increase in  $S_O$  implying that a more coherent noise tends to increase (decrease) the role of the surface fluxes (internal ocean dynamics) in the variability. However the ocean is still the dominant contributor to the growth of buoyancy variance in the northwestern corner of the basin and in the upper 250m for diffusivity values up to 2200  $m^2s^{-1}$  (not shown).

Figure 9 illustrates finally how the contribution of the ocean and atmosphere to the tempera-504 ture variability of the coupled system changes as the averaging depth increases from 50 m (the 505 depth of the forcing layer) to 1150 m (roughly the thermocline depth), the 250 m averaging depth 506 chosen previously being intermediate. Quite clearly the greater the averaging depth the stronger 507 the contribution of internal ocean dynamics to the variability, irrespective of the values of K. For 508 instance changes in surface heat fluxes represent the leading energy source of the SST variability 509 in the strongly damped regime, for  $K > 1700 \text{ m}^2 \text{s}^{-1}$ . Upper ocean heat content variability (0-1150 510 m) in this range is however mostly constrained by the production of temperature variance asso-511 ciated with internal ocean dynamics. The reason for this behaviour is that the  $S_A$  term is only 512 nonzero in the forcing layer (upper 50 m) whereas  $S_O$  monotonically decreases with depth, con-513

sistent with the result that the temperature anomalies mostly project onto the first baroclinic mode whose vertical extent is O(1000)m. Performing the variance budget over a greater depth has the effect of decreasing  $S_O$  but at a much lower rate than  $S_A$ , implying a greater role of internal ocean dynamics in the variability.

#### **6.** A coupled ocean-atmosphere box model

To better understand the way thermal coupling reinforces the internal oceanic contribution to the variability we propose herein a conceptual two degree of freedom dynamical system first introduced by Colin de Verdière and Huck (2000) to study interdecadal oscillations of the oceanic overturning under deterministic conditions. The strength of this conceptual model resides in the fact that it captures the behaviour obtained in GCMs, and in particular the Hopf bifurcation at the transition between the steady and oscillatory regimes (Arzel et al. 2018). This model is modified here to include atmospheric stochastic forcing (Fig. 10).

## <sup>533</sup> a. Model description

A detailed derivation of the model from first principles can be found in the Appendix. We sim-534 ply provide here a brief description of the model and governing equations before going on to its 535 analysis in the following sections. The model represents advection, dissipation and exchange of 536 heat within and between the ocean and atmosphere in a single-hemisphere configuration (Fig. 10). 537 The meridional redistribution of heat by advective processes between the tropical and polar boxes 538 in the ocean is assumed to be entirely accomplished by the meridional overturning circulation. The 539 oscillations we are studying here arise under constant surface wind-stress forcing. As such, heat 540 transport by the gyre circulation does not play a fundamental role in the variability and will there-541 fore be discarded, as opposed for instance to Marshall et al. (2001). The overturning circulation is 542 assumed to vary in quadrature with the anomalous temperature contrast, in agreement with results 543 from previous studies (Huck et al. 1999a; te Raa and Dijkstra 2002). This feature is also clearly 544 apparent here with the rate of change in the MOC being highly correlated (r = 0.75) with the 545 anomalous SST contrast between the subtropics (20°N-48°N) and the subpolar area (48°N-74°N) 546 in all model configurations, irrespective of the oceanic state being in the damped or super-critical 547 regime. The time lag between meridional circulation anomalies and changes in the meridional 548



<sup>526</sup> FIG. 10. Geometry of the single-hemisphere coupled ocean-atmosphere box model. Two oceanic boxes are <sup>527</sup> coupled to two atmospheric boxes in the meridional plane. Radiative fluxes at the top of the atmosphere are <sup>528</sup> the solar forcing  $Q_S$  and outgoing longwave radiation  $Q_L$ . The ocean and atmosphere exchange heat through <sup>529</sup> turbulent sensible fluxes  $Q_{AO}$ . The term *F* acting on the atmospheric layer represents the dynamical part of the <sup>530</sup> forcing and is taken to be stochastic. Mixing by turbulent motions between the tropical and polar boxes within <sup>531</sup> the atmosphere and ocean is taken into account. Advective heat exchange between the two oceanic boxes is <sup>532</sup> entirely accomplished by the meridional overturning circulation  $\Psi$ .

temperature contrast represents the delay associated with westward propagating planetary waves. 549 Turbulent mixing by mesoscale eddies in the ocean and by synoptic-scale disturbances in the at-550 mosphere between the tropical and polar boxes is included. Stochastic forcing is applied to the 551 atmospheric layer with the same decorrelation time ( $\tau_N = 10$  days) and amplitude ( $\sigma_N = 100$  W 552  $m^{-2}$ ) as the 3D model. Oceanic baroclinic instability is parameterized through a linear growth rate 553  $\mu$  for the strength of the meridional circulation. This parameter constitutes the most important pa-554 rameter of the box model because it tells us whether the oceanic state belongs to the super-critical 555 or damped regime depending on the choice of oceanic and atmospheric turbulent diffusivities and 556 other thermal damping coefficients. A dynamical ocean is thus added in the simpler way to the 557 Hasselmann (1976)'s formulation. The nondimensional model equations (see Appendix, equations 558

## <sup>559</sup> A5 and A7) are given by

$$\dot{x'} = -2\psi' - \delta x' + \beta N_0, \tag{10a}$$

$$\dot{\psi}' = \kappa x' + \mu \psi' - \gamma \psi'^3, \tag{10b}$$

where x' is the perturbation oceanic temperature difference between the tropical and polar boxes 560 and  $\psi'$  the perturbation oceanic meridional overturning circulation. There is no evolution equation 561 of the atmospheric temperature since on the long, decadal, timescales of interest here, the atmo-562 sphere is in instantaneous equilibrium with its fluxes. Under this approximation, the atmospheric 563 temperature becomes diagnostic and the atmosphere enters the problem only through the parame-564 ters  $\delta$ ,  $\beta$  and  $N_0$ . More specifically,  $\delta$  sums up the oceanic and atmospheric contributions to oceanic 565 thermal damping (oceanic and atmospheric eddy diffusivity, turbulent air-sea heat exchange, long-566 wave radiative feedback) and  $\beta$  represents the amplitude of the atmospheric stochastic forcing N<sub>0</sub> 567 seen by the ocean. The last cubic term has been introduced to stabilize the system at large am-568 plitudes. We choose  $\gamma = 300$  in all experiments. Following the previous approach, three different 569 configurations of the box model impacting the coefficients  $\delta$  and  $\beta$  are considered. The first one is 570 the standard case where ocean-atmosphere thermal coupling is enabled, with coefficients 571

$$\delta = \alpha (2K_O + \lambda(1 - \sigma))$$
 and  $\beta = \alpha \sigma \sigma_N / \overline{x}$  (11)

with  $\alpha = \tau/C_O$  and  $\sigma = \lambda/(\lambda + B + 2K_A)$ . Here  $\tau = 1$  year is the scale for time,  $C_O = 4 \times 10^9$  J K<sup>-1</sup> m<sup>-2</sup> is the heat capacity of the thermocline whose depth is fixed to 1000 m,  $\bar{x} = 20^{\circ}$ C is the mean meridional temperature contrast between the tropical and polar boxes,  $\lambda = 30$  W m<sup>-2</sup> K<sup>-1</sup> is the air-sea heat exchange coefficient,  $K_O = 1$  W m<sup>-2</sup> K<sup>-1</sup> (corresponding to 1000 m<sup>2</sup>s<sup>-1</sup>),  $K_A = 2.275$ W m<sup>-2</sup> K<sup>-1</sup> is the atmospheric eddy diffusivity (corresponding to 1.3×10<sup>6</sup> m<sup>2</sup>s<sup>-1</sup>), and B = 1.7W m<sup>-2</sup> K<sup>-1</sup> is the longwave feedback at the top of the atmophere. The second case is the forced ocean-only case where

$$\delta = 2\alpha K_0$$
 and  $\beta = \alpha \sigma \sigma_N / \overline{x}$  (12)



FIG. 11. Standard deviation of the anomalous meridional temperature contrast in the box model. For each value of the baroclinic growth rate  $\mu$  the coupled system is integrated for 10,000 years using a fourth order Runge-Kutta scheme and the statistics are computed over the second half of the integration. The integration is done for the coupled, uncoupled and forced cases under both deterministic and stochastic conditions. Hopf bifurcations occur successively at  $\mu = 0.016$ , 0.057 and 0.25 for the forced, coupled and uncoupled cases. Over the range of values of  $\mu$  considered here, the deterministic uncoupled system is always stable.

The third case is the uncoupled one where the feedback of the atmosphere on oceanic temperature anomalies is suppressed where

$$\delta = \alpha (2K_O + \lambda)$$
 and  $\beta = \alpha \sigma \sigma_N / \overline{x}$  (13)

#### <sup>581</sup> b. Bifurcations, oscillation period and power spectra

Using the values of  $\delta$  for the coupled, uncoupled and forced cases, we see that as  $\mu$  increases, 582 oscillations first emerge in the forced case ( $\mu_C = 0.016$ ), then in the coupled case ( $\mu_c = 0.057$ ), and 583 lastly in the uncoupled case ( $\mu_C = 0.25$ ). The transition from the non-oscillatory to the oscillatory 584 regime occurs through a genuine Hopf bifurcation (see Appendix for details). This sequence of 585 bifurcations is presented in Fig. 11 for the amplitude of the changes in the meridional temperature 586 contrast, as obtained from a 10,000 years numerical integration of the box model. This sequence of 587 bifurcations is consistent with the results from the 3D model (Fig. 6), demonstrating the relevance 588 of this box model to study the physics of interdecadal oscillations of the overturning circulation 589 and the effect of ocean-atmosphere thermal coupling as we shall now see. 590



FIG. 12. Theoretical power spectra (A8) of temperature anomalies for  $\mu = 0$  and for the coupled, uncoupled and forced cases (solid lines). When circulation changes are inhibited ( $\psi' = 0$ ) the canonical red noise response is obtained (dashed lines) with a temperature spectra given by  $|\hat{X}(\nu)|^2 = \beta^2 |\hat{N}(\nu)|^2 / (4\pi^2 \nu^2 + \mu_c^2)$ . All the spectra are normalized by the peak value of the forced case.

Figure 12 shows the theoretical power spectra (A8) for the meridional temperature contrast at 601  $\mu = 0$  for the coupled, uncoupled and forced cases and for cases where the circulation is free to 602 interact with the temperature field (solid lines) and prescribed to its background value (dashed 603 lines). If the circulation is not allowed to vary, technically by setting  $\psi' = 0$  in (10a), the canonical 604 response of Hasselmann (1976) is obtained: a red spectrum at high frequencies that levels out when 605  $\omega < \mu_c$ . Circulation changes are thus essential in producing a preferred timescale in the system, 606 consistent with realistic ocean-only simulations of the MITgcm (Arzel and Huck 2020). The 607 effect of thermal coupling is to reduce the damping of oceanic temperature anomalies compared 608 to the uncoupled case. There is therefore more variability in the coupled case compared to the 609 uncoupled case. This result is independent of the circulation being prescribed or not. The forced 610 response exhibits the highest variance since the damping of temperature anomalies is even weaker 611 in this case, with an effective damping timescale  $\tau \delta^{-1}$  of 63 years (compared to 18 and 4 years 612 in the coupled and uncoupled cases respectively). These results are in good agreement with those 613 deduced from the 3D model (Fig. 4). 614

#### 615 *c. Energy sources*

Following previous studies and the present approach, insight into the physical mechanisms driving the variability can be obtained by multiplying the temperature equation by x' and averaging over a long time period (denoted by an overbar),

$$\frac{1}{2}\overline{\dot{x'}^2} = -2\overline{\psi'x'} + \beta\overline{x'N_0} - \delta\overline{x'^2}$$
(14)

Similar to (9), the growth of temperature variance can either originate from internal ocean dynam-619 ics  $(-2\overline{\psi'x'})$  or from the direct effect of stochastic forcing on the oceanic temperature  $(\beta \overline{x'N_0})$ . 620 The last term is always negative and represents a sink of temperature variance. The results for 621 the coupled and uncoupled cases are presented in Fig. 13. As can be seen the box model repro-622 duces with very high fidelity the results obtained with the full 3D model (Fig. 7). The major 623 points are that thermal coupling (1) significantly increases the production of temperature variance 624 by internal oceanic processes, (2) significantly increases the destruction of temperature variance 625 by air-sea heat fluxes, but in the most weakly unstable cases where the effect of coupling is to 626 reinforce the production of temperature variance by air-sea heat fluxes, (3) and as a whole signifi-627 cantly reinforces the role of internal ocean dynamics in the variability, in particular in the damped 628 regime. 629

# **7. Summary and discussion**

The impact of ocean-atmosphere thermal coupling on the primary drivers of interdecadal variability has been investigated using an idealized coupled model. The main advantage of such a model is its ability to explore a greater parameter regime than that would be possible with a comprehensive coupled GCM. The effect of thermal coupling on the primary drivers of the variability was quantified across a wide range of oceanic eddy diffusivities from a comparison of the buoyancy variance budget of coupled and uncoupled integrations.

The results first confirm that the primary effect of thermal coupling is to reduce the internal damping of temperature anomalies due to surface heat fluxes (Barsugli and Battisti 1998). This reduced damping leads to a greater thermal variance in both the ocean and atmosphere as well as greater ocean circulation changes in coupled integrations compared to uncoupled ones, in agree-



FIG. 13. Temperature variance budget in the coupled-ocean-atmosphere box model, as obtained from a 10,000 years long numerical integration. Shown are the oceanic  $(S_O = -2\overline{x'\Psi'})$  and atmospheric energy source  $(S_A = \beta \overline{x'N_0})$  terms for both the coupled and uncoupled systems under both deterministic and stochastic forcing conditions as a function of the growth rate  $\mu$ , the differences in  $S_O$  and  $S_A$  between the coupled and uncoupled systems and the ratio  $R = S_O/(S_O + S_A^+)$  measuring the fraction of the total production of temperature variance explained by internal ocean dynamics. In the calculation of R only positive values of  $S_A$  are retained so that  $S_A^+ = 0$  wherever  $S_A < 0$ .

<sup>648</sup> ment with Wu and Liu (2005). One new aspect is that the amplifying effect of thermal coupling on <sup>649</sup> the variability is much stronger in the super-critical than in the damped regime. The presence of <sup>650</sup> an internal ocean mode in the super-critical regime therefore provides a significant positive feed-<sup>651</sup> back on the amplitude of the variability through reduced thermal damping by surface heat fluxes. <sup>652</sup> Thermal coupling is also shown to significantly alter the spatial pattern of the variability, in partic-<sup>653</sup> ular in the super-critical regime. When the interdecadal internal ocean mode is damped however, <sup>654</sup> the pattern correlation between the coupled and uncoupled leading EOFs is large. The analysis presented by Wu and Liu (2005) falls within this latter regime suggesting that the internal ocean
 mode in their model is damped.

The buoyancy variance budget then reveals that internal ocean dynamics always acts as a source 657 of temperature variance, irrespective of the coupling with the atmosphere being present or not 658 or the oceanic state being in the super-critical or damped regimes. On the contrary changes in 659 surface buoyancy fluxes always act to damp the variability in the super-critical regime. Only 660 for the most diffusive, weakly unstable states, does the atmosphere act as a source of energy 661 for the variability through the stochastic forcing exceeding surface heat flux damping. These 662 behaviours are present in both the coupled and uncoupled configurations of the model, but with 663 different magnitudes. Overall thermal coupling with the atmosphere is shown to significantly 664 reinforces the role of internal ocean dynamics in the interdecadal variability of the coupled ocean-665 atmosphere system, in particular in the damped regime. This result constitutes the major finding of 666 our study. Specifically it is shown that the leading role of internal ocean dynamics in the variability 667 extends beyond the super-critical regime to a much larger range of diffusivity values when thermal 668 coupling with the atmosphere is enabled. Put another way, thermal coupling with the atmosphere 669 significantly widens the range of diffusivity values over which internal ocean dynamics drives the 670 variability. Of course stochastic forcing is required to generate variability in the damped regime. 671 The growth of upper ocean heat content anomalies in this regime is however mostly constrained 672 by changing ocean currents associated with the noise excitation of the internal ocean mode rather 673 than by the direct thermodynamical response of the mixed layer to the noise forcing. Importantly 674 this behaviour occurs despite the fact that the leading pattern of the variability in the damped 675 regime bears some resemblance with the imposed NAO forcing. The same conclusion applies to 676 the underlying mechanism driving interdecadal sea surface temperature variability. These results 677 are in line with the more general statement that the ocean drives the mid-latitude North Atlantic 678 variability on decadal and longer timescales (Bjerknes 1964; Gulev et al. 2013; O'Reilly et al. 679 2016). Our results are also consistent with Garuba et al. (2018) who showed on the basis of 680 coupled and partially-coupled integrations of a realistic coupled climate model that the AMV is 681 mostly driven by ocean circulation variability. 682

A stochastic coupled ocean-atmosphere model is finally proposed that captures the basic effect of ocean-atmosphere thermal coupling, as obtained in the 3D model. The box model builds upon

a stochastically-forced atmospheric component coupled to a dynamical ocean. The presence of 685 the latter significantly modifies the response of the coupled system to stochastic forcing, first by 686 introducing an additional and significant source of temperature variance in the system, and second 687 by selecting an interdecadal timescale as opposed to pure thermodynamical models (Hasselmann 688 1976; Barsugli and Battisti 1998; Clement et al. 2016). The fact that the effect of coupling is 689 captured by a simple two-degree of freedom dynamical system suggests that the proposed mech-690 anism does not depend on the details of the 3D model and is therefore a robust feature of the 691 ocean-atmosphere system. 692

There are of course several fundamental aspects of the real climate system that are missing from 693 our idealized framework. Perhaps the most critical is the absence of a dynamical atmospheric 694 component. With a full dynamical atmosphere, NAO related stochastic forcing would adjust its 695 amplitude and spatial pattern to the changing ocean surface conditions and feed back onto the SST 696 field such that coupled air-sea modes or resonant behaviours could emerge (Weaver and Valcke 697 1998; Wu and Liu 2005; Ortega et al. 2015). The absence of a seasonal cycle and the limited 698 poleward extent of the ocean domain does not allow the emergence of sea ice although a sim-699 ple one-layer thermodynamic sea ice component identical to that used by Colin de Verdière and 700 te Raa (2010) is present. The variability studied here is typically enhanced at mid-latitudes along 701 the North Atlantic Current (Ortega et al. 2015; Arzel et al. 2018; Gastineau et al. 2018; Arzel and 702 Huck 2020; Gastineau et al. 2018) so that direct sea ice effects are unimportant. It would however 703 be interesting to assess the effect of thermal coupling on the variability of coupled climate mod-704 els presenting a source of variability related to Arctic-Atlantic interactions (Jungclaus et al. 2005; 705 Escudier et al. 2013). The lack of oceanic turbulence is another limitation of our model. With 706 resolved mesoscale eddies the explicit horizontal diffusivity used here becomes irrelevant and an-707 other critical parameter, such as vertical diffusivity used by Farneti and Vallis (2009) for instance, 708 would be required to explore the physics of the variability in both the damped and super-critical 709 regimes. The presence of oceanic mesoscale eddies does not modify the generic mechanism of 710 baroclinic instability captured in our simple model (Huck et al. 2015), but simply acts as a sink of 711 temperature variance at interdecadal timescales (Hochet et al. 2020, 2022). As such the leading 712 role of internal ocean dynamics in the variability advocated in the present study is probably overes-713 timated compared to a situation where mesoscale eddies are resolved. Additional studies based on 714

strongly eddying quasi-geostrophic coupled models point to a strong coupling between the ocean 715 and atmosphere on decadal and longer timescales (Kravtsov et al. 2007; Martin et al. 2021). These 716 quasi-geostrophic models cannot capture the large-scale baroclinic instability mechanism that is at 717 play in our model, because it develops in regions of steep isopycnal slopes. Future work based on 718 primitive ocean models at eddying resolutions will need to be pursued to assess which mechanism 719 among the few listed here is the most relevant for driving temperature variance from mid to subpo-720 lar latitudes as well as the role of thermal coupling which is still debated (e.g. Weaver and Valcke 721 1998; Timmermann et al. 1998; Delworth and Greatbatch 2000; Dong and Sutton 2005; Gastineau 722 et al. 2018). The generic large-scale baroclinic instability mechanism driving internal oceanic vari-723 ability in the planetary geostrophic ocean component of the present model (Colin de Verdière and 724 Huck 1999; Huck et al. 2001) has been shown to be robust in comprehensive ocean climate models 725 (Sévellec and Fedorov 2013; Ortega et al. 2015; Arzel et al. 2018; Gastineau et al. 2018). We thus 726 feel confident that the mechanisms presented here might play a role in more realistic contexts. As 727 a final note we wish to stress the extreme usefulness of the buoyancy variance budget combined 728 with the coupled/uncoupled modelling approach to identify the primary drivers of the variability 729 and unravel the role of thermal coupling. Applying such a budget in coupled and uncoupled con-730 figurations of comprehensive GCMs and realistic climate models will undoubtedly improve our 731 understanding of the mechanisms driving climate variability on interdecadal timescales in those 732 models. 733

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Data availability statement. The data that support the findings of this study are stored on IFRE MER (Brest, France) servers and can be made available upon reasonable request to the corre sponding author.

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# APPENDIX

#### The coupled ocean-atmosphere box model

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#### 743 a. Model equations

Radiative imbalance at the top of the atmosphere (Fig. 10) arises due to differences between in-744 coming solar flux  $Q_S$  (assumed constant) and outgoing longwave radiation  $Q_L$  which is linearized 745 around the atmospheric temperature  $T_a$ , i.e.  $Q_L = A + BT_a$ . The term F acting on the atmospheric 746 layer represents the dynamical part of the forcing which is taken to be stochastic. The turbulent 747 surface heat flux  $Q_{AO}$  (assumed positive downward) between the ocean and atmosphere is equal to 748  $\lambda(T_a - T_o)$  where  $\lambda$  is the air-sea heat exchange coefficient and  $T_o$  the oceanic temperature. Using 749  $\lambda = 30 \text{ W m}^{-2} \text{ K}^{-1}$ , an atmospheric heat capacity  $C_a = \rho_a C_{pa} h_a = 7 \times 10^6 \text{ J m}^{-2} \text{K}^{-1}$  (assuming 750 an atmosperic scale height  $h_a = 7$  km, and standard values for air density  $\rho_a$  and specific heat 751 capacity of air  $C_{pa}$ ), we obtain an atmospheric damping timescale  $C_a/\lambda$  of O(1) day. Since we 752 are interested in timescales of the order of 10 years or longer, it is legitimate to assume that the 753 atmosphere is in instantaneous equilibrium with the ocean and the stochastic forcing (Saravanan 754 and McWilliams 1998). The heat budget of the atmospheric box *i* then reads 755

$$Q_{S}^{i} - Q_{L}^{i} - Q_{ao}^{i} + K_{A}(T_{a}^{j} - T_{a}^{i}) + F^{i} = 0$$
(A1)

where j = 3 - i and  $K_A$  parameterizes the turbulent heat exchange by large-scale atmospheric eddies between the tropical (i = 1) and the polar (i = 2) boxes. The heat budget of the oceanic box i is

$$C_O \frac{\partial T^i}{\partial t} = Q_{ao}^i + \frac{C_O}{a_i} \psi(T^j - T^i) + K_O(T^j - T^i), \tag{A2}$$

<sup>758</sup> where  $\psi$  (m<sup>3</sup> s<sup>-1</sup>) is the overturning strength,  $C_O = \rho_o C_{po} h$  is the heat capacity of the ocean where <sup>759</sup>  $\rho_o$ ,  $C_{po}$  and h are the density, specific heat of water and ocean depth respectively, and  $K_O$  parame-<sup>760</sup> terizes the turbulent heat exchange by mesoscale eddies between the two oceanic boxes. Following <sup>761</sup> Colin de Verdière and Huck (2000) we concentrate on the variability of the temperature differences <sup>762</sup> between the two boxes. Introducing  $x = T^1 - T^2$  for the ocean and  $y = T_A^1 - T_A^2$  for the atmosphere <sup>763</sup> and assuming equal areas  $a_i$  ( $a_i = a = 4 \times 10^6$  km<sup>2</sup>) for the two boxes allows us to write

$$S - By - \lambda(y - x) - 2K_A y + N = 0 \tag{A3}$$

764 and

$$\dot{x} = \alpha \lambda (y - x) - 2\psi x - 2K_0 \alpha x, \tag{A4}$$

where  $\alpha = \tau/C_0$  and  $\psi$  is scaled by  $ha/\tau$  where  $\tau = 1$  year. The terms *S* and *N* are the differences of solar flux and noise forcing between the tropical and polar boxes respectively. The noise forcing is written as  $N = \sigma_N N_0$  where  $\sigma_N$  is the amplitude of the forcing and  $N_0$  obeys to a first order autoregressive process with standard deviation equal to 1 and decorrelation time  $\tau_N = 10$  days. Linearizing now the thermodynamic equations (A3 and A4) around a time mean state (denoted by an overbar), substituing (A3) into (A4), and nondimensionalizing the oceanic temperature anomaly x' by  $\overline{x}$  the mean meridional temperature contrast, we obtain

$$\dot{x'} = -2\psi' - \delta x' + \beta N_0, \tag{A5}$$

where  $\psi'$  is the anomalous overturning circulation. It has been shown from 3D numerical experiments (Huck et al. 1999a) that temperature anomalies are mostly constrained by anomalous advection rather than mean flow effects. Hence the term  $-2\overline{\psi}x'$  is neglected. The coefficients  $\delta$ and  $\beta$  are given by

$$\delta = \alpha (2K_O + \lambda (1 - \sigma))$$
 and  $\beta = \alpha \sigma \sigma_N / \overline{x}$  (A6)

with  $\sigma = \lambda/(\lambda + B + 2K_A)$ . Using B = 1.7 W m<sup>-2</sup> K<sup>-1</sup>,  $K_A = 2.275$  W m<sup>-2</sup> K<sup>-1</sup> (equivalent to an 776 atmospheric eddy diffusivity  $K_A \times a/C_a = 1.3 \times 10^6 \text{ m}^2 \text{s}^{-1}$ ),  $K_O = 1 \text{ W m}^{-2} \text{ K}^{-1}$  (equivalent to an 777 oceanic eddy diffusivity of  $K_O \times a/C_O = 1000 \text{ m}^2 \text{s}^{-1}$ , with  $C_O = 4 \times 10^9 \text{ J m}^{-2} \text{ K}^{-1}$ , for a 1000m 778 ocean depth), we obtain a damping timescale  $\tau \delta^{-1}$  of oceanic temperature anomalies of about 18 779 years. This timescale is much longer than the typical e-folding times of O(1) year of unstable 780 planetary waves computed from OGCM studies (Arzel et al. 2018) suggesting the potential for 781 spontaneous oscillations developping in ocean-only integrations to survive to coupling with the 782 atmosphere. Notice that the magnitude of the noise forcing seen by the ocean is proportional to 783  $\sigma$  which decreases with decreasing  $\lambda$  values and increasing  $K_A$  values. To close the system, an 784 equation for the anomaly in the strength of the circulation is needed. With an attempt to encap-785 sulate the ideas developed in the previous sections, at least two features must be represented. The 786 first one is the baroclinic instability mechanism, which is parameterized here through the use of a 787 linear growth rate  $\mu$ . The second one is the apparent phase lag that exists between the meridional 788

temperature gradient and the MOC. These aspects encourage us to follow Colin de Verdière and
Huck (2000) and use the same dynamics, that is

$$\dot{\psi}' = \kappa x' + \mu \psi' - \gamma \psi'^3,\tag{A7}$$

where the last cubic term has been introduced to stabilize the system at large amplitudes. The sensitivity of model solutions is studied in terms of the baroclinic growth rate  $\mu$ .

# 793 b. Properties of the solutions

In the deterministic case ( $\beta = 0$ ) and in the small amplitude limit, it can easily be shown (Colin de Verdière and Huck 2000) that the eigensolutions are unstable for  $\mu > \mu_c = \delta$  provided that  $\delta < \sqrt{2\kappa}$ . At  $\mu = \mu_c$  a super-critical Hopf bifurcation occurs. Physically this means that when the growth rate of pertubations becomes larger than all sources of thermal damping, the oscillations can grow and eventually settle into a finite amplitude limit cycle. In the stochastic regime, the statistics of the variability can be deduced from the temperature spectra in the small amplitude limit,

$$|\hat{X}(\omega)|^{2} = \frac{\beta^{2} |\hat{N}_{0}(\omega)|^{2}}{\omega^{2} a(\omega)^{2} + b(\omega)^{2}},$$
(A8)

where  $\hat{N}_0(\omega)$  is the Fourier transform of the noise forcing and

$$a(\omega) = 1 - \frac{2\kappa}{\omega^2 + \mu^2} \tag{A9a}$$

$$b(\omega) = \mu_c - \frac{2\kappa\mu}{\omega^2 + \mu^2} \tag{A9b}$$

For timescales much longer than O(10) days, the stochastic forcing is essentially white and  $\hat{N}_0(\omega) = 1$ . In this case an analytical expression for the most energetic timescale  $T_0 = 2\pi\omega_0^{-1}$ of the variability can be deduced from (A8) and will be valid as long as it is much longer than the atmospheric spin down timescale,

$$T_0 = 2\pi [2\sqrt{\kappa}(\mu^2 - \mu\mu_c + \kappa)^{1/2} - \mu^2]^{-1/2},$$
(A10)

while the amplitude of temperature variations is given by  $(\int |\hat{X}(\omega)|^2 d\omega)^{1/2}$ . With  $\kappa = 0.0315$ , a 805 period of 25 years is obtained in the coupled, uncoupled and forced cases, similar to the 3D model. 806 It is readily seen that the period in the deterministic case  $2\pi [2\kappa - (\mu + \mu_c)^2/4]^{-1/2}$  closely follows 807 that obtained under stochastic forcing as long as  $\mu < \mu_c \ll 1$ . At bifurcation  $\mu = \mu_c$  the period of 808 the variability predicted by the deterministic solution is exactly recovered by (A10) but the linear 809 assumption breaks down since  $|\hat{X}(\omega_0)|^2$  becomes infinite. The stochastic linear solutions become 810 invalid when approaching the Hopf bifurcation and are mostly useful in the most diffusive, weakly 811 unstable cases where  $\mu \ll \mu_c$ . 812

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