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# Direct temporal cascade of temperature variance in eddy-permitting simulations of multidecadal variability

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## ABSTRACT

11 The North Atlantic is characterized by basin-scale multidecadal fluctuations  
12 of the sea surface temperature with periods ranging from 20 to 70 years. One  
13 candidate for such a variability is a large-scale baroclinic instability of the  
14 temperature gradients across the Atlantic associated with the North Atlantic  
15 Current. Because of the long time scales involved, most of the studies de-  
16 voted to this problem are based on low resolution numerical models leaving  
17 aside the effect of explicit meso-scale eddies. How high-frequency motions  
18 associated with the meso-scale eddy field affect the basin-scale low-frequency  
19 variability is the central question of this study.

20 This issue is addressed using an idealized configuration of an Ocean General  
21 Circulation Model at eddy-permitting resolution (20 km). A new diagnostic  
22 allowing the calculation of nonlinear fluxes of temperature variance in fre-  
23 quency space is presented. Using this diagnostic, we show that the primary  
24 effect of meso-scale eddies is to damp low frequency temperature variance  
25 and to transfer it to high frequencies.

## 1. Introduction

The existence of basin-scale multidecadal fluctuations of the North Atlantic Sea Surface Temperature (SST) is long established (Bjerknes 1964; Kushnir 1994). It consists of a horseshoe pattern of SST anomaly extending from the subpolar gyre to the tropics and a weaker anomaly of opposite sign south of the equator (Kushnir 1994; Deser et al. 2010; Zhang et al. 2019). Proxy records across the North Atlantic have demonstrated the ubiquitous nature of this Low Frequency (LF) variability (Knudsen et al. 2011) over the past 8000 years. This multidecadal variability has periods in the range 20-70 years (Folland et al. 1984, 1986; Chylek et al. 2011) and is mostly referred to as the Atlantic Multidecadal Variability (AMV) to emphasize the fact that the variability is not an harmonic oscillation at a single frequency but consists of a broader band of low-frequency signals (Zhang 2017).

An AMV index can be defined as a ten-year running mean of linearly detrended SST anomalies averaged north of the equator in the Atlantic (Enfield et al. 2001). Positive phases of this index occurred during the middle of the twentieth century and since 1995, negative phases occurred during the early 20<sup>th</sup> century and during the 1964-1995 period. The large-scale LF variability has significant impacts on the Sahel/Indian summer monsoon rainfall, Atlantic hurricanes frequency and summer climate over western Europe and North America. See Zhang et al. (2019) and references therein for an extensive list of possible impacts of AMV on climate.

Three main mechanisms have been proposed to explain the observed low frequency SST variation. Because of the lack of data on such long time scales, no clear consensus has emerged. The first (controversial) one is linked with radiative forcings from anthropogenic aerosols and greenhouse gases (e.g. Watanabe and Tatebe 2019, and references therein). The second one is related to the integration of the atmospheric white noise by the ocean giving rise to a reddened spectrum

(Hasselmann 1976; Frankignoul and Hasselmann 1977). The third mechanism has dynamical origins and is related to intrinsically unstable multidecadal ocean modes. The relative contributions of these paradigms to the observed low-frequency variability of the North Atlantic climate continues to be fiercely debated (Clement et al. 2015; Zhang et al. 2016; O'Reilly et al. 2016; Gastineau et al. 2018; Zhang et al. 2019). The work presented in this paper focuses on the physics of intrinsic ocean modes.

Using a coupled general circulation model (CGCM), Delworth et al. (1993) were able to reproduce the pattern and time scale of the variability, and concluded that the AMV is intimately related to AMOC changes. Since then, the linkage between multidecadal AMOC variability and AMV has been found in many CGCMs (e.g. Delworth and Mann 2000; Danabasoglu et al. 2012; Keenlyside et al. 2016; Drews and Greatbatch 2017; Kim et al. 2018). Ba et al. (2014) performed a multi-model analysis of the AMV and found that in eight of the ten models they considered, mid-latitude SST variations are correlated with AMOC fluctuations. Numerous other studies employing CGCMs (Delworth and Greatbatch 2000; Cheng et al. 2004; Dong and Sutton 2005) concluded that the variability is due to a damped internal mode of the ocean, forced by atmosphere stochastic forcing. Wills et al. (2019) suggested that the atmospheric response to extra-tropical SST anomalies is essential to explain the AMV amplitude but that it is the ocean circulation that sustains SST anomalies on longtime scales. Observations and high resolution CGCM support the idea that the recent cooling trend in the subpolar North Atlantic upper ocean temperature since 2005 is closely linked to the observed decline in the AMOC (Robson et al. 2016).

A complementary approach to the one employing statistical analyses of CGCM outputs consists of isolating the “simplest” configuration where interdecadal variability exists and decreasing step by step the degree of idealisation (Greatbatch and Zhang 1995; Chen and Ghil 1996; Zu et al. 2013; Colin de Verdière and Huck 1999; Huck et al. 1999, 2001; Te Raa and Dijkstra 2002; Di-

jkstra and Ghil 2005; Arzel et al. 2007; Sévellec and Huck 2015; Arzel et al. 2007; Huck et al. 2015; Jamet et al. 2016; Arzel et al. 2018). The mode's existence was first shown in a rectangular flat-bottomed single-hemispheric basin, with prescribed surface heat fluxes and Planetary Geostrophic dynamics (Greatbatch and Zhang 1995; Colin de Verdière and Huck 1999). In this configuration, a large scale baroclinic instability continuously feeds a large scale mode that gives rise to SST variability which is associated geostrophically with the Meridional Overturning Circulation (Colin de Verdière and Huck 1999; Te Raa and Dijkstra 2002). The period of the mode is explained by the basin transit time of temperature anomalies set by the combination of Rossby waves and geostrophic self-advection. The same mode of variability was later demonstrated to exist in a realistic configuration of the North Atlantic (Sévellec and Fedorov 2013). Arzel et al. (2018) further showed that the internal ocean mode produces maximum SST variance in the western part of the subpolar gyre, in agreement with the observed signature of the AMO in the North Atlantic.

Sensitivity studies carried out at low resolution based on either idealized or realistic oceanic configurations have demonstrated the critical influence of the horizontal and eddy-induced turbulent diffusivities on the properties of the internal ocean mode (e.g. Huck et al. 2001; Arzel et al. 2018). More specifically, the variability disappears for diffusivity values that are typically in the range of observational estimates ( $500 - 1000 \text{ m}^2 \text{ s}^{-1}$ ) casting therefore some doubts on the relevance of this self-sustained internal mode for the observed variability. The same conclusions are found for the influence of the isopycnal diffusivity within the Gent and McWilliams (1990) parameterization in a realistic configuration (Arzel et al. 2018). Moreover, (LaCasce and Pedlosky 2004) have shown that large-scale anomalies associated with Rossby waves are prone to meso-scale baroclinic instability; they further suggested that the presence of Rossby waves in the mid to high latitudes is unlikely because at these latitudes, Rossby wave periods are slow compared to the destabilization

time scale. In contrast, the study of Huck et al. (2015) suggested that the large-scale mode is robust to meso-scale turbulence. Using a series of experiments from coarse up to eddy-resolving (10 km) resolutions, these authors showed that the spatial structure of the mode is modified, in agreement with the changes in the mean state as the resolution is refined, however, the main period and physical mechanism of the mode remains robust. An explanation for this apparent contradiction could be that instead of being a freely evolving mode as in LaCasce and Pedlosky (2004), the large-scale mode is continuously extracting energy from the mean flow through a large-scale baroclinic instability, allowing it to overcome the eddy induced dissipation.

Even if some studies suggest that the mode is robust to meso-scale turbulence, the interaction between the low frequencies and the high frequencies associated with turbulence remains an open question. Indeed Huck et al. (2015) speculated that because the range of vertical diffusivity coefficient ( $K_v$ , hereafter) allowing the existence of the low-frequency mode is larger at eddy-resolving than at low resolution, the eddy field may in fact act as a stochastic excitation of the low-frequency mode. This idea is reminiscent of the studies of Frankcombe et al. (2009); Sévellec et al. (2009); Arzel and Huck (2020) who suggested that the SST multidecadal temperature variability could be explained by a damped oceanic mode forced by atmospheric noise. Juricke et al. (2017) suggested that the low frequency variability at coarse resolution can be better simulated by enhanced subgrid-scale variability, where the latter is included by applying stochastic perturbations to the GM scheme. Additionally, a recent series of studies (Arbic et al. 2012, 2014; Sérazin et al. 2015; O’Rourke et al. 2018; Sérazin et al. 2018) further suggests that meso-scale eddies undergo what has been called a “temporal inverse cascade of kinetic energy”. This cascade is reflected in a transfer of kinetic energy from the high frequency meso-scale eddy field to lower frequencies and seems to support the idea of a possible forcing of the low frequency mode by meso-scale eddies. However there seems to be a contradiction between the idea that diffusion, which parameterizes

the effect of eddies, is a sink of temperature variance and that eddy turbulence might also be a source term for the low frequency mode.

In this work we seek to further investigate the effect of high-frequency motions associated with the meso-scale eddy field on the basin-scale low-frequency motions associated with the internal ocean mode. More specifically, we aim to determine the direction and intensity of these nonlinear temperature variance fluxes in frequency space. We choose here to focus on temperature variance rather than on Available Potential Energy for several reasons. 1) It is a variable that is directly measurable (unlike Available Potential Energy); 2) The effect of AMV is predominantly observed in SST; 3) The temperature variance budget has been extensively used in studies on intrinsic interdecadal variability of the ocean (e.g. Colin de Verdière and Huck (1999); Arzel et al. (2006, 2007); Buckley et al. (2012); Jamet et al. (2016); Arzel et al. (2018); Gastineau et al. (2018)) and has proved to be a powerful tool to understand its underlying mechanism. Here a new method is developed to compute these fluxes and applied to the simplest setup where both the large-scale LF mode and meso-scale eddies are known to exist: a flat-bottom, single hemispheric, rectangular basin forced only by constant heat fluxes at the surface with eddy-permitting resolution. The limited extent of the grid due to the single hemispheric basin allows for the simulation of both low frequency (20 to 50 years) and high frequency variability.

The article is organised as follows: in section 2, the numerical setup, the mean state, and the low and high-frequency variability are described. In section 3, we present a new diagnostic that is used to study the temperature variance fluxes in frequency space. In section 4, we derive an equation for the low-frequency temperature variance, and show in physical space the influence of meso-scale eddy variability on the LF variance. In the last section we conclude and discuss our results.



## 2. Model description, mean flow and variability

### *a. Model description*

We use the MITgcm (Marshall et al. 1997) in a rectangular flat bottom basin with a Cartesian geometry on a  $\beta$ -plane centered at  $40^\circ\text{N}$ . The zonal and meridional extents are respectively  $L_x = 5000\text{ km}$  ( $\approx 47^\circ$  at  $20^\circ\text{N}$ ) and  $L_y = 4500\text{ km}$  ( $\approx 40^\circ$ ), and the Southern boundary is located  $2000\text{ km}$  north of the equator. An eddy-permitting horizontal resolution of  $20\text{ km}$  is used in both directions. Huck et al. (2015) used higher resolution, up to  $10\text{ km}$ , but despite a stronger intensity of the meso-scale eddy field at the highest resolution, no qualitative changes in the characteristics of the multidecadal variability were found between the  $10$  and  $20\text{ km}$  runs. The depth is  $H = 4500\text{ m}$  and there are  $40$  levels in the vertical with grid spacing increasing from  $10\text{ m}$  at the surface to  $400\text{ m}$  at the bottom.

The ocean is forced by a steady heat flux at the surface. This flux is zonally uniform and decreases linearly with latitude from  $50\text{ W m}^{-2}$  at  $y = 0\text{ km}$  to  $-50\text{ W m}^{-2}$  at  $y = 4500\text{ km}$ , similar to Huck et al. (2015). Since our focus is on the physics of the low-frequency variability forced by prescribed surface heat fluxes, wind-stress forcing is set to zero in our experiments. The effect of the wind-stress forcing has been studied extensively in Quasi-Geostrophic models and was shown to produce sub-decadal gyre variability (e.g. Berloff and McWilliams 1999; Simonnet and Dijkstra 2002; Berloff et al. 2007). How wind-stress forcing affects the results of the present study remains to be addressed. A linear equation of state is used with temperature as the only active tracer and the thermal expansion coefficient is assumed uniform with a value of  $\alpha = 2 \times 10^{-4}\text{ K}^{-1}$ . Horizontal biharmonic eddy diffusivity and viscosity are used with the same uniform value of  $10^{11}\text{ m}^4\text{ s}^{-1}$ . The vertical viscosity is  $\nu_v = 10^{-3}\text{ m}^2\text{ s}^{-1}$ . No Gent-McWilliams parametrization (Gent and McWilliams 1990) is used. Static instability is removed by enhanced mixing of the

water column (to  $100 \text{ m}^2 \text{ s}^{-1}$ ). In this single hemispheric basin, the MOC strength (defined below) is strongly dependent on the value of the vertical diffusivity  $K_v$ , in agreement with the classical geostrophic scaling in the vertical advective-diffusive balance showing a  $K_v^{1/2}$  law for the strength of the MOC under prescribed surface fluxes (Huang and Chou 1994; Huck et al. 1999). Hence, we choose to use  $K_v = 2 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  to obtain a MOC strength close to 10 Sv.

### *b. Time mean circulation and low-frequency variability*

Under this configuration, LF variability of the temperature field spontaneously develops. A detailed description of the variability developing in very similar geometries can be found for instance in Huck et al. (1999); Huck and Vallis (2001); Huck et al. (2015). Here we will only give a short description of its main characteristics. The LF variability of the three dimensional temperature field is well described by the first Complex Empirical Orthogonal Function (CEOF hereafter) that contains 60% of the temperature variance. The CEOF are calculated using 20 days average outputs on a 300 years long simulation. Similar to the widely used empirical orthogonal function, CEOF are the eigenvectors of the complex covariance matrix of a complex temperature anomaly which is calculated using the Hilbert transform of the detrended temperature anomaly (Von Storch and Zwiers 2001). CEOF are more suitable than EOF to describe propagating features: the large-scale mode is indeed contained in a single CEOF while it would require two EOFs to describe the same mode. The temperature anomaly associated to a CEOF can then be reconstructed using the following formula:

$$\theta_{\text{CEOF}}(x, y, z, t) = \text{PC}_{re}(t)\text{CEOF}_{re}(x, y, z) + \text{PC}_{im}(t)\text{CEOF}_{im}(x, y, z) \quad (1)$$

where *re* and *im* stand for respectively the real and imaginary part and PC is the principal component of the corresponding CEOF. Fig. 1 shows that the low-frequency variability takes the form of

a large-scale temperature anomaly propagating northward, located mainly in the northwestern half and in the upper 500 m of the basin with SST anomaly larger than 3 K at some locations. Contours of the time mean temperature are also shown on the same figure; mean temperature decreases northward with isotherms outcropping in the northern half of the basin. In the absence of surface wind-stress forcing, box-ocean models show a single thermally-driven anticyclonic gyre with a western boundary current that remains attached to the coast from tropical to subpolar latitudes. The implied poleward heat transport along the western boundary makes the western part of the subpolar area (typically North of 50 N) always warmer than the eastern part, which is opposed to what is seen in observations at subpolar latitudes. This difference in the SST climatology does not have any consequences for either the energy source (i.e. large-scale baroclinic instability) or the main features of the variability. All these aspects are preserved when using a more realistic geometry (Sévellec and Fedorov 2013; Arzel et al. 2018; Arzel and Huck 2020).

The bottom panel of Figure 1 shows the real and imaginary part of the PC and the time series of the MOC strength. The MOC strength is defined as the vertical and meridional maximum of the time mean overturning streamfunction which is:

$$\Psi(y, z, t) = \int_z^0 \int_{x=0}^L v dx dy, \quad (2)$$

and the MOC strength is then  $\Psi(y_{max}, z_{max}, t)$  with  $y_{max} = 3500\text{km}$  and  $z_{max} = -500\text{m}$  in this setup. Following the time evolution of the PC (bottom panel of Fig. 1), the oscillation consists of four steps separated by one quarter of a cycle:  $re \rightarrow im \rightarrow -re \rightarrow -im \rightarrow re$  where  $re$  and  $im$  are respectively the real and imaginary part of the CEOF. The minimum MOC strength shown on the bottom panel of Fig. 1 is correlated with positive anomaly of SST close to the Northern boundary (the first CEOF real part times  $-1$  on the left panel of figure 1). Maximum values of MOC strength are correlated with negative temperature anomaly being close to the Northern boundary

(the first CEOF real part on the left panel of figure 1). The phase of the oscillation represented by the CEOF is chosen so that the correlation between the CEOF PC real part and the MOC strength is maximum. The real part of the PC follows closely the MOC strength time variation giving us confidence that the leading CEOF adequately represents the low frequency variability of our setup. To estimate the period of this low-frequency mode, we calculate the power spectrum of temperature at each grid point and average them. For this calculation we use a 500 year long simulation made of 1 year time average outputs. The temperature spectrum plotted in Figure 2 has a distinct peak at a period of  $\sim 42$  years which is in the range of that found in previous studies. A secondary peak of smaller intensity is found at a period of  $\sim 21$  years and is attributed to the second harmonic of the leading CEOF. In agreement with the comparison of the PC real part and MOC strength on the bottom of figure 1, the MOC strength frequency spectrum also shows a peak at a period of  $\sim 42$  years (fig. 2).

### c. High-frequency variability

In the absence of wind forcing, Kinetic Energy (KE) can only be forced through Available Potential Energy (APE) to KE conversion. This unique source of KE creates nonetheless a strong eddy field that, as will be shown below, accounts for more than 50% of the total sink of the low-frequency mode temperature variance. The temperature spectrum calculated with 3-day average outputs on a 50-year simulation is shown in Fig. 2. The spectrum follows two different slopes depending on the frequency: a slight slope for frequencies smaller than  $2\pi/3$  months $^{-1}$  and a steep slope for frequencies larger than  $2\pi/3$  months $^{-1}$ . The standard deviation of the Sea Surface Height (SSH) (which is a quantity that is easily observable by satellite altimetry, and can be used to assess eddy activity) with a 2 year running average removed and calculated on the same run, is plotted on the left panel of figure 3. The largest values (around 0.16m) are found close to the

Western boundary and in the Northern half of the basin. These values are in the range of observed values of SSH standard deviation as observed by TOPEX/POSEIDON (Stammer 1997) suggesting that the level of eddy activity in our simulation is realistic. A snapshot of the ratio of the surface relative vorticity ( $\zeta = \partial_x v - \partial_y u$ ) normalized by the Coriolis parameter  $f$  (Fig. 3, right panel) reveals the presence of zonally organised eddying features that can also be seen as zonal jets in the time averaged zonal velocity (Fig. 4). It is interesting to note that the values of this ratio remain much smaller than 1, so that the geostrophic assumption holds even for the meso-scale turbulence.

### 3. Transfer of temperature variance in frequency space

In this section we derive the budget for the transfer of temperature variance in frequency space. We then study the different temperature fluxes of our setup and show in particular that there is a nonlinear flux of temperature variance toward higher frequencies.

#### *a. Temperature variance fluxes in frequency: definition*

Given our setup, the temperature  $\theta$  obeys the following equation:

$$\frac{\partial \theta}{\partial t} + \mathbf{u} \cdot \nabla \theta = D + G \quad (3)$$

where  $\mathbf{u}$  is the 3D velocity  $(u, v, w)$ ,  $D$  represents the parameterised diffusion and the convection and  $G$  is the surface heat flux. To obtain an equation for the temperature variability, we decompose all variables into time mean plus anomaly:  $X = \bar{X} + X'$  with  $X$  being  $\theta, \mathbf{u}, D$ , or  $G$ .  $\bar{(\cdot)}$  is the time mean over the 200 years of the simulation,  $(\cdot)'$  is the anomaly with  $\bar{X}' = 0$ , by construction. Note that  $G$  is time independent and thus equal to its time mean. The time mean of Eq. (3) is:

$$\bar{\mathbf{u}} \cdot \nabla \bar{\theta} + \overline{\mathbf{u}' \cdot \nabla \theta'} = \bar{D} + G, \quad (4)$$

251 The time evolution equation for  $\theta'$  is obtained by subtracting Eq. (4) from Eq. (3):

$$\frac{\partial \theta'}{\partial t} = -\bar{\mathbf{u}} \cdot \nabla \theta' - \mathbf{u}' \cdot \nabla \bar{\theta} - \mathbf{u}' \cdot \nabla \theta' + \overline{\mathbf{u}' \cdot \nabla \theta'} + D'. \quad (5)$$

252 Transfers in the frequency or in the space domain are usually calculated for KE or APE (see for  
 253 instance Arbic et al. 2014, and references therein). The usual way of calculating these QG KE  
 254 or APE transfers is to Fourier transform the vorticity equation multiplied by the conjugate of the  
 255 Fourier transform of the stream function (more frequently in space but it can also be done in  
 256 time). Here we follow this idea but apply it to the temperature and in the frequency domain. The  
 257 temperature anomaly  $\theta'$  is detrended in time and multiplied by a Tukey window, with a value of  
 258 0.2 for the ratio of taper to constant sections as in Arbic et al. (2014) in order to obtain a periodic  
 259 field in time. Temperature spectral transfers are calculated by multiplying the discrete Fourier  
 260 Transform in time of each component of Eq. (5) by the conjugate of the Fourier transform of  $\theta$  at  
 261 every grid point. The result is integrated over the domain to give, in a statistical steady state:

$$0 = \text{Tr}_{\text{mean}}(\omega) + \text{Tr}_{\text{turb}}(\omega) + \text{Tr}_{\text{diss}}(\omega) \quad (6)$$

262 with  $\omega$  the frequency. The expression for  $\text{Tr}_{\text{mean}}(\omega)$  is:

$$\text{Tr}_{\text{mean}}(\omega) = - \int_V \Re \left( \widehat{\theta'}^*(\omega) \left( \widehat{\bar{\mathbf{u}} \cdot \nabla \theta'}(\omega) + \widehat{\mathbf{u}' \cdot \nabla \bar{\theta}}(\omega) \right) \right) dV, \quad (7)$$

263 where  $\Re$  is the real part,  $\omega$  the frequency and  $V$  the volume. The spectral transfers identify  
 264 temperature variance sources (when positive) or sinks (when negative) in frequency space. Note  
 265 that  $\bar{\mathbf{u}} \cdot \nabla \theta'$  and  $\mathbf{u}' \cdot \nabla \bar{\theta}$  are calculated offline from the values of  $\mathbf{u}$  and  $\theta$ .  $\text{Tr}_{\text{mean}}(\omega)$  is interpreted  
 266 as the transfer of temperature variance linked with mean flow anomaly interaction. We will show  
 267 below that this transfer is a source of temperature variance at every frequency. The formula for  
 268 term  $\text{Tr}_{\text{turb}}(\omega)$  is:

$$\text{Tr}_{\text{turb}}(\omega) = - \int_V \Re \left( \widehat{\theta'}^*(\omega) \left( \widehat{\mathbf{u}' \cdot \nabla \theta'}(\omega) \right) \right) dV, \quad (8)$$

and is interpreted as the transfer of temperature variance by turbulent processes. We will show that these turbulent transfers are either a source or a sink of temperature variance depending on the frequency. The last term  $\text{Tr}_{\text{diss}}(\omega)$  is:

$$\text{Tr}_{\text{diss}}(\omega) = \int_V \Re \left( \widehat{\theta'}^*(\omega) \widehat{D'}(\omega) \right) dV \quad (9)$$

and is the diffusive transfer of temperature variance computed from the actual dissipation output of the model, a term which will be shown to be negative for every frequency. The term involving  $\overline{\mathbf{u}' \cdot \nabla \theta'}$  disappears when multiplied by  $\widehat{\theta'}^*$  because its Fourier transform is nonzero only for  $\omega = 0$  and by definition of  $\theta'$ ,  $\widehat{\theta'}(\omega = 0) = 0$ , so that  $\widehat{\theta'}^* \widehat{\mathbf{u}' \cdot \nabla \theta'} = 0$ .

Following other studies (e.g. Scott and Arbic (2007)) we will define and use below spectral fluxes instead of spectral transfers for the three following reasons: 1) fluxes are much less noisy than the transfers, 2) as will become clear below, turbulent transfers are zero when integrated over all frequencies and using fluxes makes this fact directly apparent, 3) our interest is on what happens for a range of frequencies (we will define two ranges below: high-frequencies and low-frequencies) rather than at a local frequency. The drawback of using spectral fluxes rather than transfers is that the sign of the transfer needs to be deduced from the sign of the slope of the fluxes which is less direct than looking directly at the sign of the transfers.

We define the spectral fluxes  $\Pi$  of the spectral transfers  $\text{Tr}$  as:

$$\Pi(\omega) = \int_{\omega}^{\omega_{\max}} \text{Tr}(\omega') d\omega', \quad (10)$$

$\Pi_{\text{mean}}(\omega)$ ,  $\Pi_{\text{turb}}(\omega)$  and  $\Pi_{\text{diss}}(\omega)$  are respectively the input of temperature variance from the mean flow, the turbulent terms, and the diffusivity in the frequency domain defined by all  $\omega'$  with  $\omega' > \omega$ . Thus a positive (negative)  $\Pi(\omega)$  corresponds to a positive (negative) total input of variance between  $\omega$  and  $\omega_{\max}$ . The total transfer between  $\omega_{\min}$  and  $\omega$  is simply:  $\Pi(\omega_{\min}) - \Pi(\omega)$ . If  $\Pi(\omega_{\min}) - \Pi(\omega)$  is positive (negative) the total transfer between  $\omega_{\min}$  and  $\omega$  is positive (negative).

Obtaining the transfers from the flux is straightforward:

$$\frac{d\Pi}{d\omega}(\omega) = -Tr(\omega) \quad (11)$$

The slope of the  $\Pi$  curve is thus equal to minus the transfer. From Eq. (6) and the flux formula (10) we deduce that :

$$\Pi_{\text{mean}}(\omega) + \Pi_{\text{turb}}(\omega) + \Pi_{\text{diss}}(\omega) = 0, \quad (12)$$

i.e. the sum of all fluxes is zero for every frequency.

### *b. Properties of temperature variance fluxes*

In this subsection we explain some properties of the temperature variance flux budget (i.e. eq. (12)) that will be useful to interpret results in the next subsection.

It is straightforward to show that the effect of the turbulent terms integrated over the volume is zero at each time step. Multiplying the term  $\mathbf{u}' \cdot \nabla \theta'$  from eq. (5) by  $\theta'$  and integrating over the volume  $V$  gives:

$$\int_V \theta' \mathbf{u}' \cdot \nabla \theta' dV = \frac{1}{2} \int_V \mathbf{u}' \cdot \nabla \theta'^2 dV = \int_V \nabla \cdot \left( \mathbf{u}' \frac{\theta'^2}{2} \right) dV = \oint_{\partial V} \mathbf{u}' \cdot \mathbf{n} \frac{\theta'^2}{2} dS = 0, \quad (13)$$

where the last equality makes use of the no mass flux condition through the boundaries, and where  $\mathbf{n}$  is the outward normal to the volume  $V$ . Similarly, the advection of the temperature anomaly by the mean flow is zero when integrated on the volume because of the no mass flux condition:

$$\int_V \theta' \bar{\mathbf{u}} \cdot \nabla \theta' dV = 0. \quad (14)$$

The term involving the mean temperature gradient does not a priori vanish when volume integrated, therefore, the integral on the volume of the temperature variance equation is:

$$\frac{1}{2} \frac{\partial \int_V \theta'^2 dV}{\partial t} = - \int_V \theta' \mathbf{u}' \cdot \nabla \bar{\theta} dV + \int_V \theta' D'_\theta dV + \int_V \theta' \bar{\mathbf{u}} \cdot \nabla \theta' dV \quad (15)$$



The last term on the rhs is not zero when volume integrated but its time average is zero (because  $\overline{\theta'} = 0$ ). The second term on the rhs is generally negative because positive anomalies of temperature are correlated with negative values of diffusion. The first term of the rhs can be described as a measure of the turbulent flux of  $\theta'$  in the mean temperature gradient direction. It acts to increase (decrease) the total variance when positive (negative) temperature anomalies are fluxed in the direction opposite to the mean temperature gradient. It is the only source of temperature variance in our setup (when time averaged). This term has been used in many studies to diagnose regions where baroclinic instability adds temperature variance to the flow (see for instance: Colin de Verdière and Huck 1999; Huck et al. 1999).

Using Parseval's theorem, it is possible to link the volume integral of the temperature variance terms with the frequency integral of the corresponding transfers. Indeed, using Eqs. (10) and (13) we have:

$$\Pi_{\text{turb}}(\omega_{\min}) = \int_{\omega_{\min}}^{\omega_{\max}} \text{Tr}_{\text{turb}}(\omega') d\omega' = 0, \quad (16)$$

the total flux of temperature variance made by the turbulent terms is zero. The role of the turbulent terms is to redistribute the temperature variance among different frequencies. From the fact that the term involving the mean temperature does not vanish, we deduce that the total flux made by the mean flow is not zero:

$$\Pi_{\text{mean}}(\omega_{\min}) = \int_{\omega_{\min}}^{\omega_{\max}} \text{Tr}_{\text{mean}}(\omega') d\omega' \neq 0. \quad (17)$$

Then, from Eq. (12) and Eq. (16), the total flux of variance from the mean flow is equal to the opposite of the total diffusive flux:

$$\Pi_{\text{mean}}(\omega_{\min}) = -\Pi_{\text{diss}}(\omega_{\min}) \quad (18)$$

### c. Application to an eddy-permitting experiment

In this section we apply the diagnostics from the two previous subsections to the MITgcm numerical simulations described in Section 2.

To calculate the fluxes we use a 200-year run with time average outputs every 20 days. The output sampling rate needs to be large enough to resolve high frequencies. No significant changes were found between 20 days and higher output frequencies (not shown). Figure 5 shows the temperature variance fluxes  $\Pi_{\text{mean}}(\omega)$ ,  $\Pi_{\text{turb}}(\omega)$  and  $\Pi_{\text{diss}}(\omega)$ . As expected from Eq. (16),  $\Pi_{\text{turb}}(\omega_{\min})$  is zero and  $\Pi_{\text{diss}}(\omega_{\min}) = -\Pi_{\text{mean}}(\omega_{\min})$ .  $\Pi_{\text{mean}}$  is positive and increases toward lower frequencies showing that the mean flow adds temperature variance. On the contrary,  $\Pi_{\text{diss}}(\omega)$  is always negative and decreases toward lower frequencies because dissipation removes temperature variance.  $\Pi_{\text{turb}}(\omega)$  increases toward low-frequencies and then decreases, clearly establishing that nonlinear terms are a sink of temperature variance for frequencies approximately smaller than  $2\pi/3.5 \text{ yr}^{-1}$  and a source of variance for frequencies larger than  $2\pi/3.5 \text{ yr}^{-1}$ . This variation of the direction of the turbulent transfer of temperature with frequency is the central result of this study.

This allows us to define High-Frequencies (HF) as being larger than  $2\pi/3.5 \text{ yr}^{-1}$  and Low-Frequencies (LF) as being smaller than  $2\pi/3.5 \text{ yr}^{-1}$ . This definition allows us to robustly separate the meso-scale eddy turbulence (HF) from the multidecadal mode of variability described in section c.  $\omega_{\text{mid}} \equiv 2\pi/3.5 \text{ yr}^{-1}$  is the maximum of  $\Pi_{\text{turb}}(\omega)$ ; thus by definition, turbulent transfers are a source term in HF while they are a sink in LF. The slope of  $\Pi_{\text{mean}}$  is larger at LF than at HF, suggesting that the flux of temperature variance from the mean flow is larger for LF than for HF. On the contrary, the slope of  $\Pi_{\text{diss}}$  is larger for HF than for LF, suggesting that dissipation of temperature variance is more important for HF than for LF. An excess of temperature variance is created at low frequencies by instability of the mean flow, transferred to high frequen-

cies where it is dissipated. An alternative and equivalent definition for HF and LF could then be that, HF are frequencies for which the temperature variance total transfer from the mean flow (i.e.  $\Pi_{mean}(\omega_{min}) - \Pi_{mean}(\omega_{mid})$ ) is smaller than the removal of temperature variance by the dissipation (i.e.  $|\Pi_{diss}(\omega_{min}) - \Pi_{diss}(\omega_{mid})|$  with  $|\cdot|$  the absolute value) while the opposite is true for LF. Figure 6 is a schematic description of the fluxes shown in Figure 5 between the HF and LF. All fluxes are expressed as a percentage of the total flux from the mean flow i.e.  $\Pi_{mean}(\omega_{min})$ . The transfer of temperature variance from LF to HF by turbulent terms represents 44% of the total flux by the mean flow. 79% of the variance from the mean flow is added at LF while 65% is removed by the dissipation at HF. The main pathway of the temperature variance is clearly from the mean flow LF to the HF dissipation. In other words, since LF are mainly due to the multidecadal mode (see Figure 2) and the source of variability of the mode is the mean temperature gradient, this source is mainly balanced by the meso-scale eddy turbulence acting as a sink of variance and to a lesser extent by LF dissipation. The main source of variance for the meso-scale eddy turbulence is the multidecadal mode and to a lesser extent the mean temperature gradient. These sources of HF temperature variability are balanced by the HF dissipation.

#### 4. Spatial pattern of the temperature variance transfer

In this section we show how the different terms of the temperature budget act on the low-frequency variability. To this end we split the temperature into time mean plus low-frequency ( $\theta_{LF}$ ) plus high-frequency ( $\theta_{HF}$ ) parts:

$$\theta = \bar{\theta} + \theta_{LF} + \theta_{HF} \quad (19)$$

we use a low-pass filter noted  $\langle \cdot \rangle$  so that  $\langle \theta \rangle = \bar{\theta} + \theta_{LF}$  and obtain  $\theta_{HF}$  as  $\theta_{HF} = \theta - \langle \theta \rangle$ .

The cutoff period for the Butterworth low-pass filter is chosen to be 3.5 years to match the results

from the previous section. In this section, the effect of HF on LF is calculated as a remainder, as will become clear below. This remainder allows us to use smaller output frequency (100 days average in this section) which significantly simplifies the computation presented below and allows the study of a longer run (300 years).

#### a. Low-frequency temperature variance budget

To obtain an evolution equation for  $\theta_{LF}$ , we first write the time mean of the temperature equation:

$$\bar{\mathbf{u}} \cdot \nabla \bar{\theta} + \overline{(\mathbf{u}_{LF} + \mathbf{u}_{HF}) \cdot \nabla (\theta_{LF} + \theta_{HF})} = \bar{D} + G \quad (20)$$

Using (19) and subtracting Eq. (20) from the temperature equation (3), applying the low-pass filter, multiplying by  $\theta_{LF}$  and volume integrating gives the low-frequency temperature variance budget :

$$\frac{1}{2} \frac{\partial}{\partial t} \int_V \theta_{LF}^2 dV = V_{\text{mean}} + V_{\text{diss}} + V_{\text{HF}} \quad (21)$$

where:

$$V_{\text{term}} = \int_V v_{\text{term}} dV \quad (22)$$

with term either mean, diss or HF and:

$$v_{\text{mean}} = -\theta_{LF} \mathbf{u}_{LF} \cdot \nabla \bar{\theta} \quad (23)$$

$$v_{\text{diss}} = \theta_{LF} D_{LF} \quad (24)$$

$$v_{\text{HF}} = -\theta_{LF} \langle \mathbf{u}_{LF} \cdot \nabla \theta_{HF} \rangle - \theta_{LF} \langle \mathbf{u}_{HF} \cdot \nabla \theta_{LF} \rangle - \theta_{LF} \langle \mathbf{u}_{HF} \cdot \nabla \theta_{HF} \rangle + \theta_{LF} \overline{(\mathbf{u}_{HF} + \mathbf{u}_{LF}) \cdot \nabla (\theta_{HF} + \theta_{LF})} \quad (25)$$

$V_{\text{diss}}$  is interpreted as the effect of the parameterised diffusion and convection,  $V_{\text{mean}}$  as the mean flow forcing. In  $v_{\text{HF}}$ , terms involving a time mean and a HF term disappear when low-pass filtered. None of the terms of  $v_{\text{HF}}$  can a priori be ruled out; however we have checked, using higher

frequency outputs on a shorter length simulation, that all terms involving only one HF can be neglected so that:

$$V_{HF} \approx \int_V \left( -\theta_{LF} < \mathbf{u}_{HF} \cdot \nabla \theta_{HF} > + \theta_{LF} \overline{\mathbf{u}_{HF} \cdot \nabla \theta_{HF}} + \theta_{LF} \overline{\mathbf{u}_{LF} \cdot \nabla \theta_{LF}} \right) dV. \quad (26)$$

$V_{HF}$  is thus interpreted as the influence of high-frequency motions on low-frequency temperature variance. In practice  $V_{HF}$  is computed as the remainder of all other terms,  $v_{mean}$  is calculated from the values of  $\theta_{LF}$ ,  $\mathbf{u}_{LF}$  and  $\bar{\theta}$  and  $v_{diss}$  from the values of  $D_{LF}$  and  $\theta_{LF}$ . Each term of Eq. (21) is shown in figure 7. The main equilibrium is between the forcing by  $V_{mean}$  and the dissipation by both  $V_{HF}$  and  $V_{diss}$ . High levels of LF forcing are correlated with high levels of LF dissipation: the (time) correlation coefficient is -0.87 between  $V_{mean}$  and  $V_{diss}$  and -0.85 between  $V_{mean}$  and  $V_{HF}$ . Note that the two correlations are similar but most of the variability is taken up by  $V_{HF}$ . The negative correlation between  $V_{mean}$  and  $V_{diss}$  and between  $V_{mean}$  and  $V_{HF}$  is because LF temperature gradients and meso-scale activity increase with LF forcing. On the same figure is shown the real and imaginary part of the temperature leading CEOF PC calculated in section 3. The correlation coefficient between the imaginary part PC and  $V_{diss}$ ,  $V_{mean}$  and  $V_{HF}$  is small: respectively -0.16, 0.01, and 0.09 while it is much larger with the PC real part: respectively 0.66, -0.62 and 0.60. The real part of the PC and by construction the AMOC strength time variation are thus approximately in phase with the LF dissipation, the LF NL transfer toward HF and the LF mean flow forcing. This is explained by the fact that a stronger AMOC is associated with increased temperature gradients and thus with a larger LF mean flow forcing, LF dissipation and with a stronger eddy field. The lagged correlation (see Fig. 8) between  $PC_{re}$  and  $V_{HF}$ ,  $V_{mean}$  and  $V_{diss}$  shows that  $PC_{re}$  is close to be in phase with the three latter. There is however a small time lag of respectively -1.1 and -0.6 year between  $V_{diss}$  and  $PC_{re}$  and between  $V_{mean}$  and  $PC_{re}$  while  $V_{HF}$  is almost exactly in phase with  $PC_{re}$ . The fact that  $V_{HF}$  is lagging the two other terms from 0.6 to 1.1 year is not surprising

because of the finite growth time-scale of eddy turbulence compared to the instantaneous action of the parameterized dissipation.

To better understand the spatial distribution of the  $V_{HF}$  term, we first calculate the vertical average over the water column of  $v_{HF}$ , that is:

$$V_{HF}^{int} = \frac{1}{H} \int_{z=-H}^0 v_{HF} dz \quad (27)$$

where  $H$  is the total depth. Then, to analyse its temporal variation, we time average  $V_{HF}^{int}$  following the sign of the PC's real part. The PC's real part is used rather than the imaginary part because it correlates (positively) with  $V_{HF}$  (see last paragraph). The time average of  $V_{HF}^{int}$  on all  $t$  satisfying  $PC_{re}(t) \geq 0$  and all  $t$  satisfying  $PC_{re}(t) < 0$  is shown on the two panels of figure 9. Note that  $PC_{re}(t)$  correlates very well with the MOC strength so that positive values of  $PC_{re}(t)$  correspond to a positive MOC anomaly (see subsection 3.c). Large negative values of  $V_{HF}^{int}$  are mostly located in the North-West quadrant of the basin and follow large gradients of temperature created by the low-frequency variability. Temperature fronts created by the low-frequency variability are unstable and eroded by the subsequent eddies associated with high-frequencies. As first shown in the last section, the high-frequency part of the temperature variance is mostly forced by the low-frequency part (44% vs 21% in figure 6), explaining why the spatial location of  $V_{HF}^{int}$  is moving with time (Fig. 9). Indeed if the only source of meso-scale eddies was the time mean temperature gradients, the location and intensity of the meso-scale field would remain constant in time. Volume-integrating the time-average of  $V_{HF}^{int}$  reveals that the quantity of eddy induced dissipation occurring when  $PC_{re} > 0$  or  $PC_{re} < 0$  is of the same order of magnitude with respectively 43% and 57% of the total eddy induced dissipation.

## b. Propagation of the temperature anomaly

The propagation of the temperature anomaly is usually explained (e.g. Sévellec and Fedorov (2013)) as the result of the advection of the time mean temperature by the anomalous velocities. The purpose of this subsection is to test this explanation and more broadly to understand what terms drive the oscillation captured by the temperature leading CEOF shown in figure 1. In particular, we want to understand if the HF field plays a role in the LF temperature propagation. To this end, we derive below a diagnostic that allows us to evaluate the contribution of each term from the temperature anomaly budget (5) in the propagation of the leading CEOF temperature anomaly. We first calculate the linear regression of the LF temperature budget (Eq. (5)) on the real and imaginary parts of the leading CEOF's PC (see formula (1)) to obtain a separate equation for the time evolution of the real ( $\theta_{LF}^{re}$ ) and imaginary ( $\theta_{LF}^{im}$ ) part of the CEOF:

$$\left[\frac{\partial \theta_{LF}}{\partial t}\right]^{re} = -[\bar{\mathbf{u}} \cdot \nabla \theta_{LF}]^{re} - [\mathbf{u}_{LF} \cdot \nabla \bar{\theta}]^{re} - [\mathbf{u}_{LF} \cdot \nabla \theta_{LF}]^{re} + [\mathbf{D}_{LF}]^{re} + [G_{HF}]^{re} \quad (28)$$

where  $[\cdot]^{re}$  is the linear regression against the real part of the first CEOF PC and  $G_{HF}$  is the effect of high frequencies on low frequency temperature ( $\int_V \theta_{LF} G_{HF} dV = V_{HF}$ ). The linear regression is given by:

$$[A(x, y, z, t)]^{re} = \frac{\int_0^T A(x, y, z, t) \text{PC}_{re}(t) dt}{\int_0^T (\text{PC}_{re}(t))^2 dt} \quad (29)$$

where  $A$  is any of the terms in Eq. (28) and  $T$  the length of the time series. We proceed the same way for the imaginary part:

$$\left[\frac{\partial \theta_{LF}}{\partial t}\right]^{im} = -[\bar{\mathbf{u}} \cdot \nabla \theta_{LF}]^{im} - [\mathbf{u}_{LF} \cdot \nabla \bar{\theta}]^{im} - [\mathbf{u}_{LF} \cdot \nabla \theta_{LF}]^{im} + [\mathbf{D}_{LF}]^{im} + [G_{HF}]^{im} \quad (30)$$

where  $[\cdot]^{im}$  is the regression against the imaginary part of the first CEOF PC, using Eq. (29) but with  $\text{PC}_{im}$  instead of  $\text{PC}_{re}$ . Following figure 1 the oscillation steps are  $re \rightarrow im \rightarrow -re \rightarrow -im \rightarrow re$ . The two last transitions are the same as the first two except for their sign and we can thus restrict

ourselves to the former. To understand what term is a positive or a negative contributor to  $re \rightarrow im$  ( $im \rightarrow -re$ ) we volume-average the product of  $[\frac{\partial \theta_{LF}}{\partial t}]^{re}$  and each term of Eq. (28) ( $[\frac{\partial \theta_{LF}}{\partial t}]^{im}$  and each term of Eq. (30)). When the volume-average of the product is positive (negative) the term positively (negatively) contributes to the oscillation. All values are normalized by the volume average of  $\left([\frac{\partial \theta_{LF}}{\partial t}]^{re}\right)^2$  for Eq. (28) and by the volume average of  $\left([\frac{\partial \theta_{LF}}{\partial t}]^{im}\right)^2$  for Eq. (30) and shown in table 1.

In agreement with previous descriptions of this LF mode (for instance in Sévellec and Fedorov (2013)), the propagation of the large-scale temperature anomaly is mainly due to the term  $\mathbf{u}_{LF} \cdot \nabla \bar{\theta}$ . The LF temperature anomaly is associated with anomalous LF velocities that in turn advect mean temperature in a way that creates this propagation. We show, in figure 10, that  $\bar{\mathbf{u}} \cdot \nabla \theta_{LF}$  resists the propagation. To understand why the term involving  $\bar{\mathbf{u}} \cdot \nabla \theta_{LF}$  is opposite to  $\mathbf{u}_{LF} \cdot \nabla \bar{\theta}$  we show, in figure 10, the two terms regressed against the real and imaginary parts of the 1st CEOF PC. It is seen that, for both real and imaginary parts we have:

$$\mathbf{u}_{LF} \cdot \nabla \bar{\theta} \propto -\bar{\mathbf{u}} \cdot \nabla \theta_{LF}, \quad (31)$$

where  $\propto$  is the proportionality symbol. This is due to the “non-Doppler shift” (Rossby 1939; Held 1983; Killworth et al. 1997; Liu 1999) which states that the geostrophic self advection and the mean advection exactly cancel each other when the mean flow and the anomaly have the same vertical structure.

The contributions from the remaining terms ( $\mathbf{u}_{LF} \cdot \nabla \theta_{LF}$ , diffusion and eddy turbulence) are much smaller. Thus, eddy turbulence that appears when eddy-permitting resolution are used does not significantly modify the oscillatory mechanism that was previously described in low resolution studies (Te Raa and Dijkstra 2002; Sévellec and Fedorov 2013).



## 5. Conclusion

In this article we seek to better understand the interaction between high and low-frequency variability in the ocean, i.e., between a multidecadal mode of variability and meso-scale eddy turbulence. To this end, we study an idealized configuration of the North Atlantic using an OGCM at eddy-permitting resolution (20km) that allows the coexistence of a multidecadal mode of variability and high-frequency meso-scale variability. In agreement with many other studies of this mode (Colin de Verdière and Huck 1999; Sévellec and Huck 2015), the multidecadal mode is sustained by temperature variance drawn from the time mean circulation. This mode has a dominant period of 42 years and consists of large-scale temperature anomalies propagating northward in the Northern half of the basin. The high-frequency variability is associated with meso-scale eddy turbulence driven by temperature variance drawn from both the time-mean flow and the low-frequency large-scale mode. A new diagnostic of temperature variance fluxes in frequency space reveals that high-frequencies are a sink of temperature variance for the low-frequencies. At low and high frequencies the positive flux of temperature variance from the mean flow is not equal to the opposite of the diffusive flux: a turbulent flux of temperature variance from LF to HF balances the budget for the LF and HF temperature variance reservoirs. Indeed, in our setup, the temperature variance flux from the mean flow is more than two times larger than the dissipation at low frequencies. Writing the temperature variance equation for the LF allows visualization of the term contributing to the high and low frequency variance. This term that we attribute to the overall effect of meso-scale eddies, follows the LF temperature gradient in the western half of the basin showing that HF erodes the LF fronts. Regarding the propagation of the LF temperature anomaly we have shown that the eddy term plays no significant role. The main driver of the propagation is the advection of time mean temperature by LF anomalous velocities:  $\mathbf{u}_{LF} \cdot \nabla \bar{\theta}$  mainly counter-

acted by the advection of anomalous LF temperature by time mean velocities:  $\bar{\mathbf{u}} \cdot \nabla \theta_{LF}$ . Results from O'Rourke et al. (2018); Sérazin et al. (2018); Martin et al. (2020) have shown that surface KE undergoes an inverse cascade of energy in both spatial and temporal space. Our results are not necessarily in contradiction with these studies because under the QG approximation, the temperature variance behaves closely to QG Available Potential Energy which is known to follow a direct cascade in both spatial and temporal space (Arbic et al. 2014). However in our simulation the QG approximation does not hold (because isotherms outcrop at the surface in the Northern half of the basin) and the direction of the APE cascade therefore remains unclear. The large-scale low-frequency variability is primarily seen in temperature and has very little KE (because its scale is much larger than the deformation radius) which makes it more relevant to an analysis in terms of temperature variance.

LaCasce and Pedlosky (2004); Isachsen et al. (2007) have shown that propagating Rossby waves are prone to meso-scale instabilities. At high latitudes, where Rossby waves phase speed is slower than at lower latitudes, their studies suggest that meso-scale eddies prevent Rossby waves from reaching the western boundary. Given that the large scale temperature anomaly that we see in our simulation could be described as a Rossby wave modified by the effect of the mean flow, the same mechanism is at play in our study except that the large scale variability is continuously forced by the large scale baroclinic instability (Colin de Verdière and Huck 1999). This baroclinic instability compensates for the transfer to HF made by meso-scale eddies (as shown by the schematic in figure 6). As explained in the introduction, it was found in Huck et al. (2015), that the range of  $K_v$  allowing a low frequency oscillation is larger at high than at low resolution. Huck et al. (2015) then suggested that this could be explained by the fact that meso-scale eddies were acting as a forcing for the LF mode at high resolution. Because we showed that the temperature variance transfer is directed toward high frequencies, this wider range of  $K_v$  is however probably due to a

smaller effective diffusivity in the high resolution run rather than due to a forcing by meso-scale eddies. Indeed the mode is known to be sensitive to the amount of diffusivity (Huck and Vallis 2001) and the total effective diffusivity is hard to assess at eddy-permitting resolution.

As shown by previous studies (e.g. Huck et al. (2015)) the use of a single hemispheric basin allows us to capture the essential features of the large scale mode in a simple framework. Indeed, the vertical structure, energy source, period of the large scale mode found in this study are also seen in realistic configuration of the North Atlantic and world ocean (Sévellec and Fedorov 2013; Muir and Fedorov 2017; Arzel et al. 2018).

However, several limitations arise from the omission of the full basin. Indeed it is now conjectured that the effect of wind over the Southern Ocean needs to be taken into account to obtain a realistic stratification as well as realistic values of  $K_v$  (Toggweiler and Samuels 1998; Gnanadesikan 1999; Vallis 2000; Nikurashin and Vallis 2012). The effect of  $K_v$  on our results will be reported in a separate study where we show that increasing  $K_v$  strengthens the eddy field and thus the transfer of LF temperature variance to HF. The mode eventually disappears for unrealistically large values of  $K_v$  ( $\approx 1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$ ). In our setup, the eddy-field is of primary importance for the large-scale low-frequency mode because it represents more than 50% of the LF temperature variance destruction. With higher resolution and subsequent lower parameterised diffusion, virtually all the low frequency dissipation may be performed by the eddy field. Thus, future studies should adequately simulate eddies by allowing a more realistic stratification through the effect of wind forcing. Also, wind forcing over the North Atlantic creates its own kind of low frequency variability (Berloff and McWilliams 1999; Berloff et al. 2007) that might interact with the low frequency mode described in this study and might modify substantially the eddy field and potentially the dissipation of the low frequency mode. We do not expect a fundamental change when using eddy-resolving rather than eddy-permitting resolution. Indeed, Huck et al. (2015) have shown that if the eddy field is

stronger at 10 km than at 20 km horizontal resolution, the large scale mode period, spatial scale and mechanism remain almost unchanged. However, we can expect that the intensity of the turbulent transfer of temperature variance from LF to HF will increase with increased resolution until all the input of LF temperature variance by the mean flow is removed by the turbulent transfer from LF to HF (i.e. the flux due to the parameterised dissipation disappears). Sérazin et al. (2018) compared the surface kinetic energy inverse cascade in a  $1/4^\circ$  and in a  $1/12^\circ$  global simulation. They found that the  $1/4^\circ$  simulation resolves the main non-linear oceanic processes generating the KE inverse cascade. We are thus confident that our 20 km resolution simulation also captures the essential features of the non-linear oceanic processes.

Other limitations come from the omission of active salt tracer and freshwater forcing though it has been shown in a realistic setup (Sévellec and Fedorov 2013) that the mode is largely controlled by temperature variation in the upper ocean. The simulations shown here cannot be compared with observations because of the idealized forcing and geometry of the basin. However, low resolution simulations (Sévellec and Fedorov 2013; Buckley et al. 2012, e.g.) have shown that the low frequency variability still exists in more realistic setups. The intensity and location of the meso-scale field and of the large scale mode is believed to strongly depend on the details of the forcing bathymetry and coastline. The next step will thus be to study the interaction of low and high frequencies in a realistic configuration of a high resolution OGCM.

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## References

- Arbic, B. K., M. Müller, J. G. Richman, J. F. Shriver, A. J. Morten, R. B. Scott, G. Sérazin, and T. Penduff, 2014: Geostrophic turbulence in the frequency–wavenumber domain: Eddy-driven low-frequency variability. *Journal of Physical Oceanography*, **44** (8), 2050–2069.
- Arbic, B. K., R. B. Scott, G. R. Flierl, A. J. Morten, J. G. Richman, and J. F. Shriver, 2012: Non-linear cascades of surface oceanic geostrophic kinetic energy in the frequency domain. *Journal of Physical Oceanography*, **42** (9), 1577–1600.
- Arzel, O., and T. Huck, 2020: Contributions of atmospheric stochastic forcing and intrinsic ocean modes to north atlantic ocean interdecadal variability. *Journal of Climate*, **33** (6), 2351–2370.
- Arzel, O., T. Huck, and A. Colin de Verdière, 2006: The different nature of the interdecadal variability of the thermohaline circulation under mixed and flux boundary conditions. *Journal of physical oceanography*, **36** (9), 1703–1718.
- Arzel, O., T. Huck, and A. Colin de Verdière, 2018: The internal generation of the atlantic ocean interdecadal variability. *Journal of Climate*, **31** (16), 6411–6432.
- Arzel, O., A. C. d. V. Verdière, and T. Huck, 2007: On the origin of interdecadal oscillations in a coupled ocean atmosphere model. *Tellus A: Dynamic Meteorology and Oceanography*, **59** (3), 367–383.
- Ba, J., and Coauthors, 2014: A multi-model comparison of atlantic multidecadal variability. *Climate dynamics*, **43** (9-10), 2333–2348.
- Berloff, P., A. M. C. Hogg, and W. Dewar, 2007: The turbulent oscillator: A mechanism of low-frequency variability of the wind-driven ocean gyres. *Journal of Physical Oceanography*, **37** (9), 2363–2386.

- 580 Berloff, P. S., and J. C. McWilliams, 1999: Large-scale, low-frequency variability in wind-driven  
581 ocean gyres. *Journal of Physical Oceanography*, **29** (8), 1925–1949.
- 582 Bjerknes, J., 1964: Atlantic Air-Sea Interaction. *Advances in Geophysics*, H. E. Landsberg, and  
583 J. Van Mieghem, Eds., Vol. 10, Elsevier.
- 584 Buckley, M. W., D. Ferreira, J.-M. Campin, J. Marshall, and R. Tulloch, 2012: On the relation-  
585 ship between decadal buoyancy anomalies and variability of the atlantic meridional overturning  
586 circulation. *Journal of Climate*, **25** (23), 8009–8030.
- 587 Chen, F., and M. Ghil, 1996: Interdecadal variability in a hybrid coupled ocean-atmosphere model.  
588 *Journal of physical oceanography*, **26** (8), 1561–1578.
- 589 Cheng, W., R. Bleck, and C. Rooth, 2004: Multi-decadal thermohaline variability in an ocean–  
590 atmosphere general circulation model. *Climate dynamics*, **22** (6-7), 573–590.
- 591 Chylek, P., C. K. Folland, H. A. Dijkstra, G. Lesins, and M. K. Dubey, 2011: Ice-core data ev-  
592 idence for a prominent near 20 year time-scale of the Atlantic Multidecadal Oscillation. *Geo-*  
593 *physical Research Letters*, **38** (13).
- 594 Clement, A., K. Bellomo, L. N. Murphy, M. A. Cane, T. Mauritsen, G. Radel, and B. Stevens,  
595 2015: The Atlantic Multidecadal Oscillation without a role for ocean circulation. *Science*,  
596 **350** (6258), 320–324.
- 597 Colin de Verdière, A., and T. Huck, 1999: Baroclinic instability: An oceanic wavemaker for  
598 interdecadal variability. *Journal of Physical Oceanography*, **29** (5), 893–910.
- 599 Danabasoglu, G., S. G. Yeager, Y.-O. Kwon, J. J. Tribbia, A. S. Phillips, and J. W. Hurrell,  
600 2012: Variability of the atlantic meridional overturning circulation in ccsm4. *Journal of cli-*  
601 *mate*, **25** (15), 5153–5172.

- Delworth, T., S. Manabe, and R. J. Stouffer, 1993: Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model. *Journal of Climate*, **6** (11), 1993–2011.
- Delworth, T. L., and R. J. Greatbatch, 2000: Multidecadal thermohaline circulation variability driven by atmospheric surface flux forcing. *Journal of Climate*, **13** (9), 1481–1495.
- Delworth, T. L., and M. E. Mann, 2000: Observed and simulated multidecadal variability in the northern hemisphere. *Climate Dynamics*, **16** (9), 661–676.
- Deser, C., M. A. Alexander, S.-P. Xie, and A. S. Phillips, 2010: Sea surface temperature variability: Patterns and mechanisms. *Annual review of marine science*, **2**, 115–143.
- Dijkstra, H. A., and M. Ghil, 2005: Low-frequency variability of the large-scale ocean circulation: A dynamical systems approach. *Reviews of Geophysics*, **43** (3).
- Dong, B., and R. T. Sutton, 2005: Mechanism of interdecadal thermohaline circulation variability in a coupled ocean–atmosphere gcm. *Journal of climate*, **18** (8), 1117–1135.
- Drews, A., and R. J. Greatbatch, 2017: Evolution of the atlantic multidecadal variability in a model with an improved north atlantic current. *Journal of Climate*, **30** (14), 5491–5512.
- Enfield, D. B., A. M. Mestas-Núñez, and P. J. Trimble, 2001: The Atlantic Multidecadal Oscillation and its relation to rainfall and river flows in the continental U.S. *Geophysical Research Letters*, **28** (10), 2077–2080, doi:10.1029/2000GL012745, URL <http://doi.wiley.com/10.1029/2000GL012745>.
- Folland, C., D. Parker, and F. Kates, 1984: Worldwide marine temperature fluctuations 1856–1981. *Nature*, **310** (5979), 670.
- Folland, C. K., T. N. Palmer, and D. E. Parker, 1986: Sahel rainfall and worldwide sea temperatures, 1901–85. *Nature*, **320** (6063), 602.

- Frankcombe, L. M., H. A. Dijkstra, and A. Von der Heydt, 2009: Noise-induced multidecadal variability in the North Atlantic: Excitation of normal modes. *Journal of Physical Oceanography*, **39** (1), 220–233.
- Frankignoul, C., and K. Hasselmann, 1977: Stochastic climate models, Part II Application to sea-surface temperature anomalies and thermocline variability. *Tellus*, **29** (4), 289–305.
- Gastineau, G., J. Mignot, O. Arzel, and T. Huck, 2018: North atlantic ocean internal decadal variability: Role of the mean state and ocean-atmosphere coupling. *Journal of Geophysical Research: Oceans*, **123** (8), 5949–5970.
- Gent, P. R., and J. C. McWilliams, 1990: Isopycnal mixing in ocean circulation models. *Journal of Physical Oceanography*, **20** (1), 150–155.
- Gnanadesikan, A., 1999: A simple predictive model for the structure of the oceanic pycnocline. *Science*, **283** (5410), 2077–2079.
- Greatbatch, R. J., and S. Zhang, 1995: An interdecadal oscillation in an idealized ocean basin forced by constant heat flux. *Journal of climate*, **8** (1), 81–91.
- Hasselmann, K., 1976: Stochastic climate models part I. Theory. *tellus*, **28** (6), 473–485.
- Held, I. M., 1983: Stationary and quasi-stationary eddies in the extratropical troposphere: Theory. *Large-scale dynamical processes in the atmosphere*, 127–168.
- Huang, R. X., and R. L. Chou, 1994: Parameter sensitivity study of the saline circulation. *Climate Dynamics*, **9** (8), 391–409.
- Huck, T., O. Arzel, and F. Sévellec, 2015: Multidecadal variability of the overturning circulation in presence of eddy turbulence. *Journal of Physical Oceanography*, **45** (1), 157–173.



- Huck, T., A. Colin de Verdière, and A. J. Weaver, 1999: Interdecadal variability of the thermohaline circulation in box-ocean models forced by fixed surface fluxes. *Journal of physical oceanography*, **29** (5), 865–892.
- Huck, T., and G. K. Vallis, 2001: Linear stability analysis of the three-dimensional thermally-driven ocean circulation: application to interdecadal oscillations. *Tellus A*, **53** (4), 526–545.
- Huck, T., G. K. Vallis, and A. Colin de Verdière, 2001: On the robustness of the interdecadal modes of the thermohaline circulation. *Journal of climate*, **14** (5), 940–963.
- Isachsen, P., J. H. LaCasce, and J. Pedlosky, 2007: Rossby wave instability and apparent phase speeds in large ocean basins. *Journal of physical oceanography*, **37** (5), 1177–1191.
- Jamet, Q., T. Huck, O. Arzel, J.-M. Campin, and A. Colin de Verdière, 2016: Oceanic control of multidecadal variability in an idealized coupled GCM. *Climate Dynamics*, **46** (9-10), 3079–3095, doi:10.1007/s00382-015-2754-3, URL <http://link.springer.com/10.1007/s00382-015-2754-3>.
- Juricke, S., T. N. Palmer, and L. Zanna, 2017: Stochastic subgrid-scale ocean mixing: impacts on low-frequency variability. *Journal of Climate*, **30** (13), 4997–5019.
- Keenlyside, N. S., J. Ba, J. Mecking, N.-E. Omrani, M. Latif, R. Zhang, and R. Msadek, 2016: North atlantic multi-decadal variability—mechanisms and predictability. *Climate change: Multidecadal and beyond*, World Scientific, 141–157.
- Killworth, P. D., D. B. Chelton, and R. A. de Szoeke, 1997: The speed of observed and theoretical long extratropical planetary waves. *Journal of Physical Oceanography*, **27** (9), 1946–1966.

- Kim, W. M., S. Yeager, P. Chang, and G. Danabasoglu, 2018: Low-frequency north atlantic climate variability in the community earth system model large ensemble. *Journal of Climate*, **31** (2), 787–813.
- Knudsen, M. F., M.-S. Seidenkrantz, B. H. Jacobsen, and A. Kuijpers, 2011: Tracking the Atlantic Multidecadal Oscillation through the last 8,000 years. *Nature communications*, **2**, 178.
- Kushnir, Y., 1994: Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. *Journal of Climate*, **7** (1), 141–157.
- LaCasce, J. H., and J. Pedlosky, 2004: The instability of Rossby basin modes and the oceanic eddy field. *Journal of physical oceanography*, **34** (9), 2027–2041.
- Liu, Z., 1999: Planetary wave modes in the thermocline: Non-doppler-shift mode, advective mode and green mode. *Quarterly Journal of the Royal Meteorological Society*, **125** (556), 1315–1339.
- Marshall, J., A. Adcroft, C. Hill, L. Perelman, and C. Heisey, 1997: A finite-volume, incompressible Navier Stokes model for studies of the ocean on parallel computers. *Journal of Geophysical Research: Oceans*, **102** (C3), 5753–5766.
- Martin, P. E., B. K. Arbic, A. McC. Hogg, A. E. Kiss, J. R. Munroe, and J. R. Blundell, 2020: Frequency-domain analysis of the energy budget in an idealized, coupled, ocean-atmosphere model. *Journal of Climate*, (2020).
- Muir, L. C., and A. V. Fedorov, 2017: Evidence of the amoc interdecadal mode related to westward propagation of temperature anomalies in cmip5 models. *Climate Dynamics*, **48** (5-6), 1517–1535.

- Nikurashin, M., and G. Vallis, 2012: A theory of the interhemispheric meridional overturning circulation and associated stratification. *Journal of Physical Oceanography*, **42** (10), 1652–1667.
- O'Reilly, C. H., M. Huber, T. Woollings, and L. Zanna, 2016: The signature of low-frequency oceanic forcing in the atlantic multidecadal oscillation. *Geophysical Research Letters*, **43** (6), 2810–2818.
- O'Rourke, A. K., B. K. Arbic, and S. M. Griffies, 2018: Frequency-domain analysis of atmospherically forced versus intrinsic ocean surface kinetic energy variability in GFDL CM2-O model hierarchy. *Journal of Climate*, **31** (5), 1789–1810.
- Robson, J., P. Ortega, and R. Sutton, 2016: A reversal of climatic trends in the north atlantic since 2005. *Nature Geoscience*, **9** (7), 513–517.
- Rossby, C., 1939: Relationship between variations in the intensity of the zonal variation and the displacement of the semipermanent centers of action. *Journal of Marine Research*, **2**, 38–55.
- Scott, R. B., and B. K. Arbic, 2007: Spectral energy fluxes in geostrophic turbulence: Implications for ocean energetics. *Journal of physical oceanography*, **37** (3), 673–688.
- Sérazin, G., T. Penduff, S. Grégorio, B. Barnier, J.-M. Molines, and L. Terray, 2015: Intrinsic variability of sea level from global  $1/12^\circ$  ocean simulations: Spatiotemporal scales. *Journal of Climate*, **28** (10), 4279–4292.
- Sévellec, F., and A. V. Fedorov, 2013: The leading, interdecadal eigenmode of the Atlantic meridional overturning circulation in a realistic ocean model. *Journal of Climate*, **26** (7), 2160–2183.
- Sévellec, F., and T. Huck, 2015: Theoretical investigation of the atlantic multidecadal oscillation. *Journal of Physical Oceanography*, **45** (9), 2189–2208.

- Sévellec, F., T. Huck, M. Ben Jelloul, and J. Vialard, 2009: Nonnormal multidecadal response of the thermohaline circulation induced by optimal surface salinity perturbations. *Journal of Physical Oceanography*, **39** (4), 852–872.
- Simonnet, E., and H. A. Dijkstra, 2002: Spontaneous generation of low-frequency modes of variability in the wind-driven ocean circulation. *Journal of physical oceanography*, **32** (6), 1747–1762.
- Stammer, D., 1997: Global characteristics of ocean variability estimated from regional topex/poseidon altimeter measurements. *Journal of Physical Oceanography*, **27** (8), 1743–1769.
- Sérazin, G., T. Penduff, B. Barnier, J.-M. Molines, B. K. Arbic, M. Muller, and L. Terray, 2018: Inverse Cascades of Kinetic Energy as a Source of Intrinsic Variability: A Global OGCM Study. *Journal of Physical Oceanography*, **48** (6), 1385–1408.
- Te Raa, L. A., and H. A. Dijkstra, 2002: Instability of the thermohaline ocean circulation on interdecadal timescales. *Journal of physical oceanography*, **32** (1), 138–160.
- Toggweiler, J. R., and B. Samuels, 1998: On the ocean large-scale circulation near the limit of no vertical mixing. *Journal of Physical Oceanography*, **28** (9), 1832–1852.
- Vallis, G. K., 2000: Large-scale circulation and production of stratification: Effects of wind, geometry, and diffusion. *Journal of Physical Oceanography*, **30** (5), 933–954.
- Von Storch, H., and F. W. Zwiers, 2001: *Statistical analysis in climate research*. Cambridge university press.
- Watanabe, M., and H. Tatebe, 2019: Reconciling roles of sulphate aerosol forcing and internal variability in Atlantic multidecadal climate changes. *Climate Dynamics*, **53** (7-8), 4651–4665, publisher: Springer.

- Wills, R. C., K. C. Armour, D. S. Battisti, and D. L. Hartmann, 2019: Ocean–atmosphere dynamical coupling fundamental to the atlantic multidecadal oscillation. *Journal of Climate*, **32** (1), 251–272.
- Zhang, R., 2017: On the persistence and coherence of subpolar sea surface temperature and salinity anomalies associated with the Atlantic multidecadal variability. *Geophysical Research Letters*, **44** (15), 7865–7875.
- Zhang, R., R. Sutton, G. Danabasoglu, T. L. Delworth, W. M. Kim, J. Robson, and S. G. Yeager, 2016: Comment on “the atlantic multidecadal oscillation without a role for ocean circulation”. *Science*, **352** (6293), 1527–1527.
- Zhang, R., R. Sutton, G. Danabasoglu, Y.-O. Kwon, R. Marsh, S. G. Yeager, D. E. Amrhein, and C. M. Little, 2019: A Review of the Role of the Atlantic Meridional Overturning Circulation in Atlantic Multidecadal Variability and Associated Climate Impacts. *Reviews of Geophysics*, **0** (ja), doi:10.1029/2019RG000644, URL <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2019RG000644>.
- Zu, Z., M. Mu, and H. A. Dijkstra, 2013: Optimal nonlinear excitation of decadal variability of the North Atlantic thermohaline circulation. *Chinese journal of oceanology and limnology*, **31** (6), 1368–1374.

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$\theta_{LF}$ budget term	re $\rightarrow$ im	im $\rightarrow$ -re
$\mathbf{u}_{LF} \cdot \nabla \bar{\theta}$	+219%	+213%
$\bar{\mathbf{u}} \cdot \nabla \theta_{LF}$	-129%	-153%
$\mathbf{u}_{LF} \cdot \nabla \theta_{LF}$	+14%	+24%
$G_{HF}$	-0%	+14%
$D_{LF}$	-4%	+2%
$\Sigma$	+100%	+100%

TABLE 1. Contributions to the LF evolution of temperature, computed as the volume integrated correlations between  $[\frac{\partial \theta_{LF}}{\partial t}]^{re}$  and the rhs terms of (28), and  $[\frac{\partial \theta_{LF}}{\partial t}]^{im}$  and Eq. (30) (“im $\rightarrow$ -re” column). For each column, values are expressed as a percentage of the sum of all terms.

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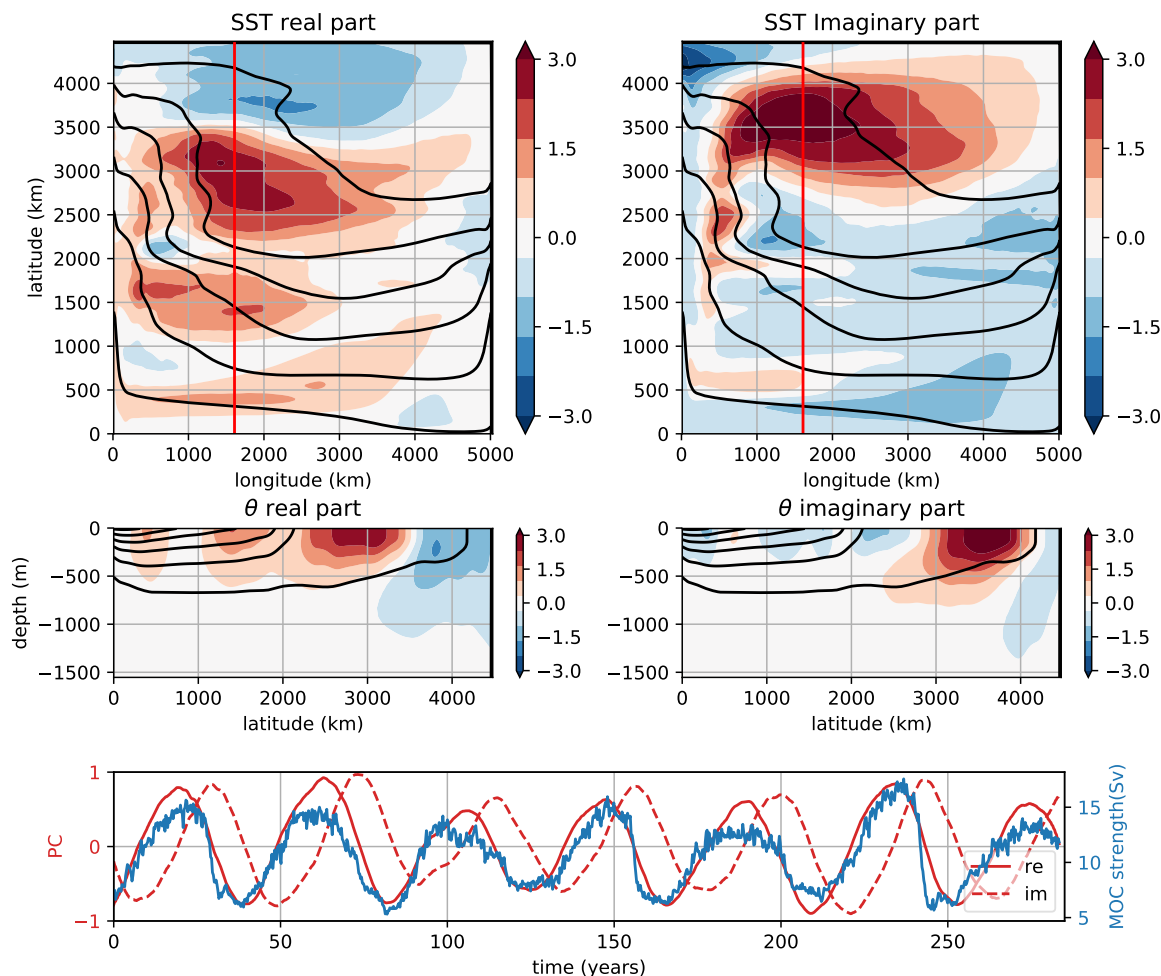


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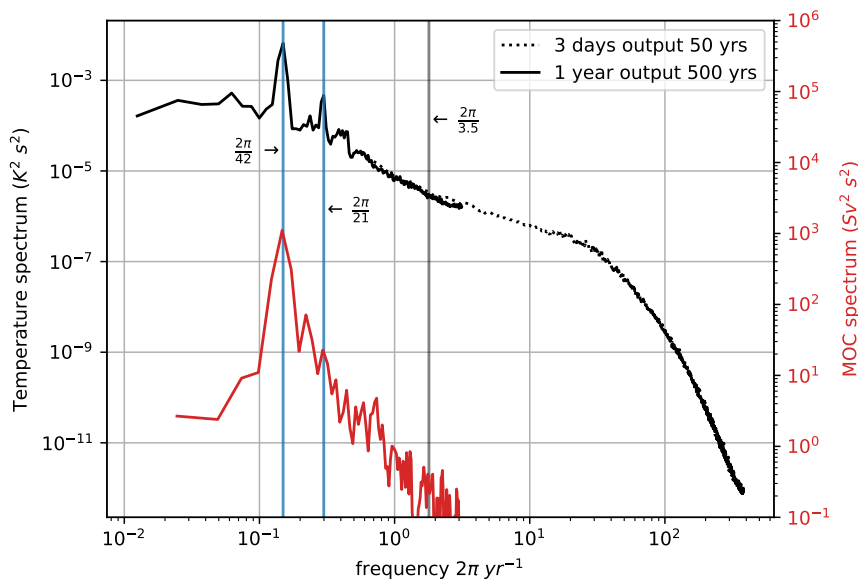


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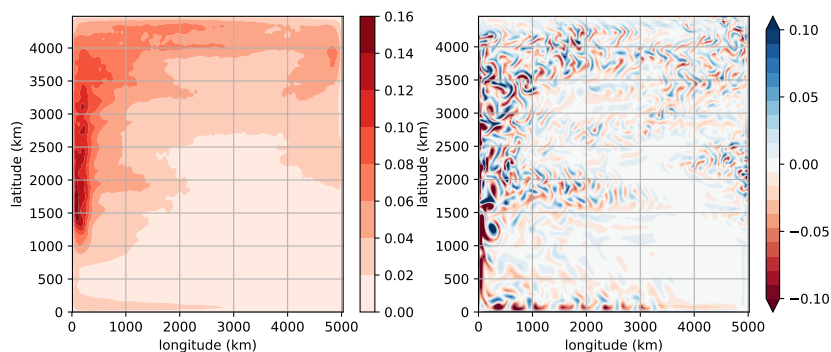


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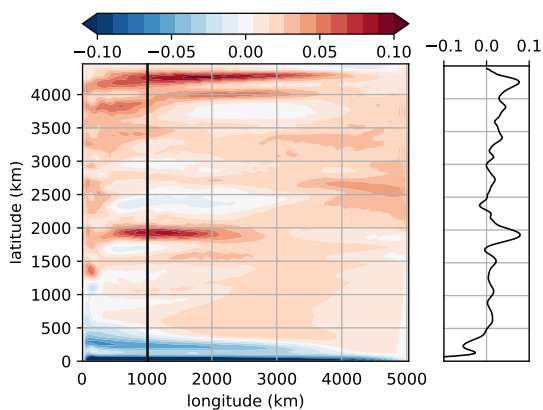


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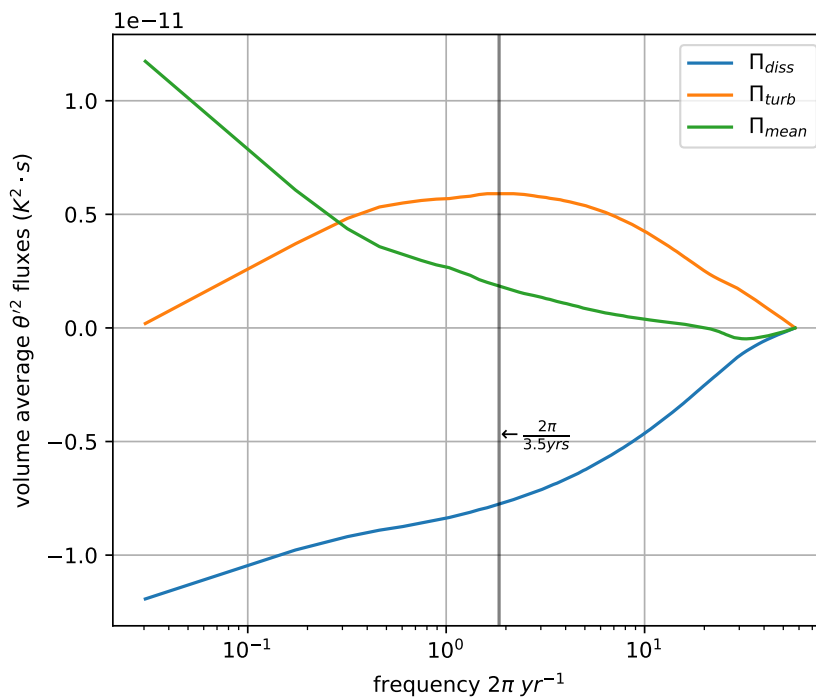


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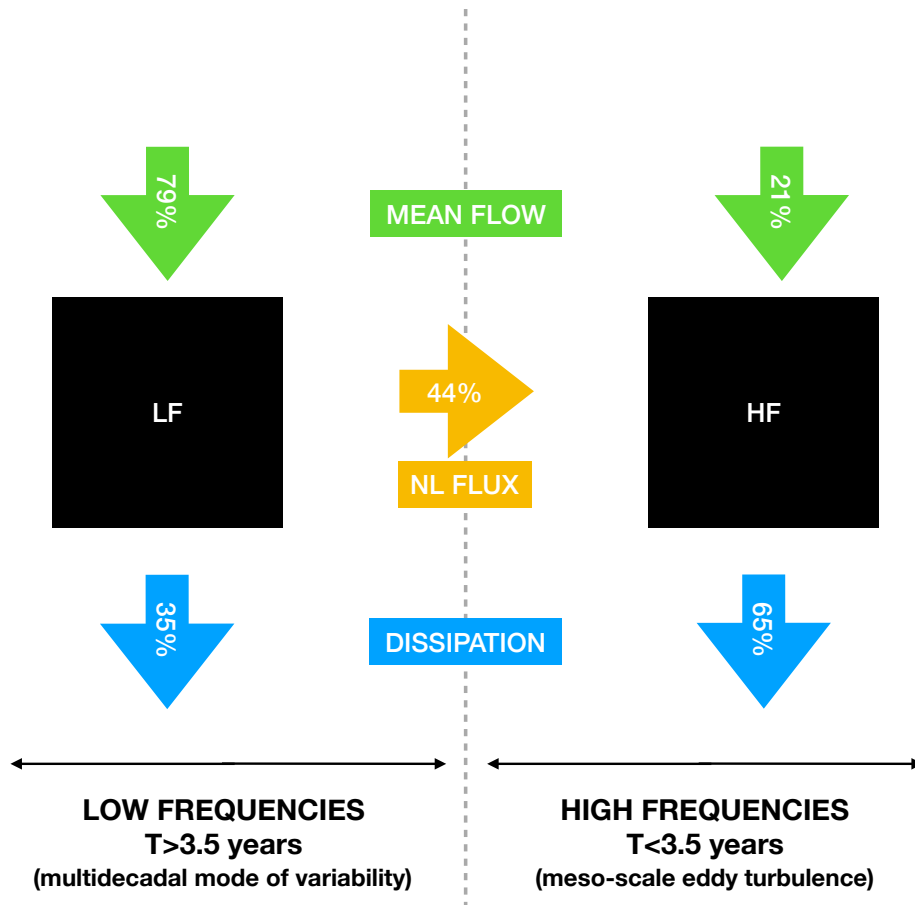


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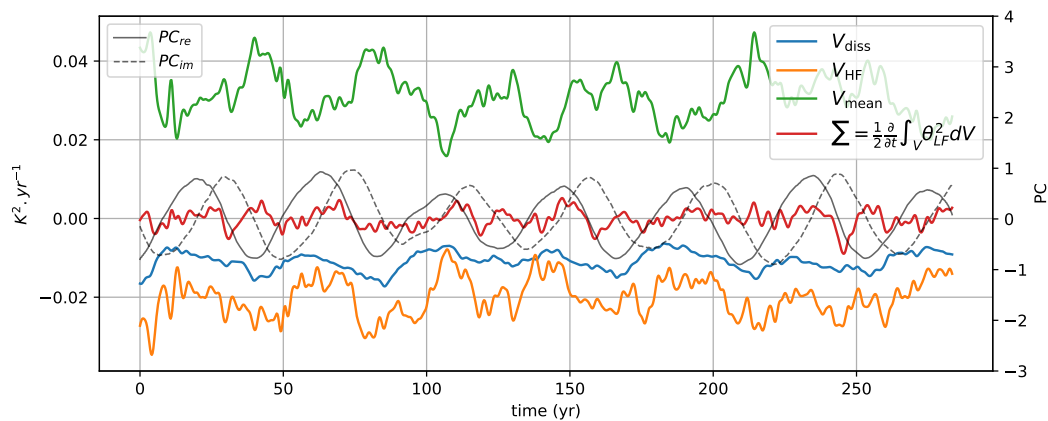


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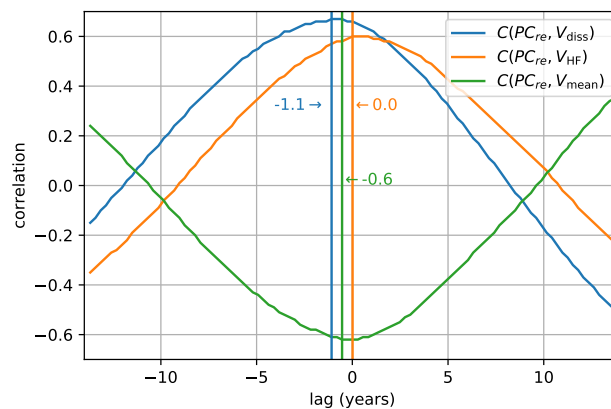


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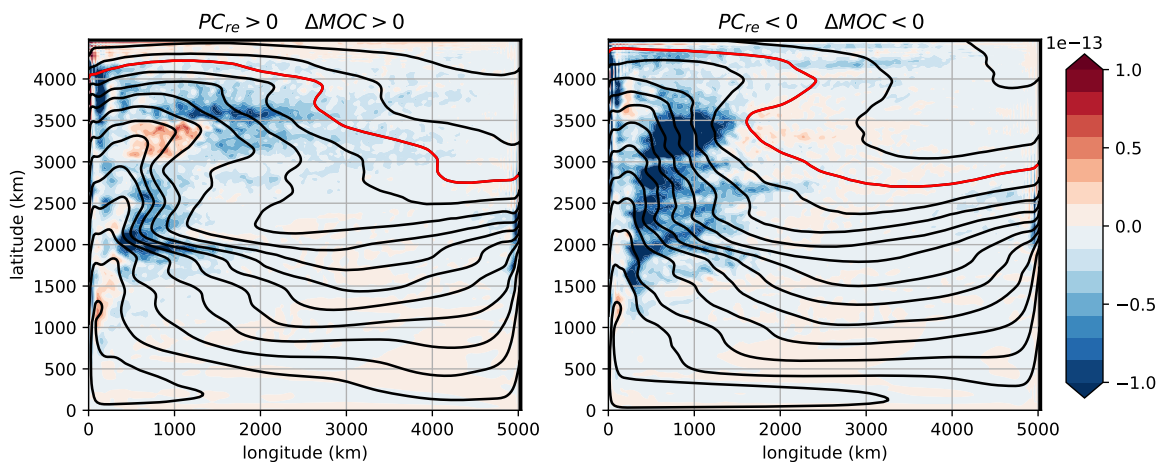


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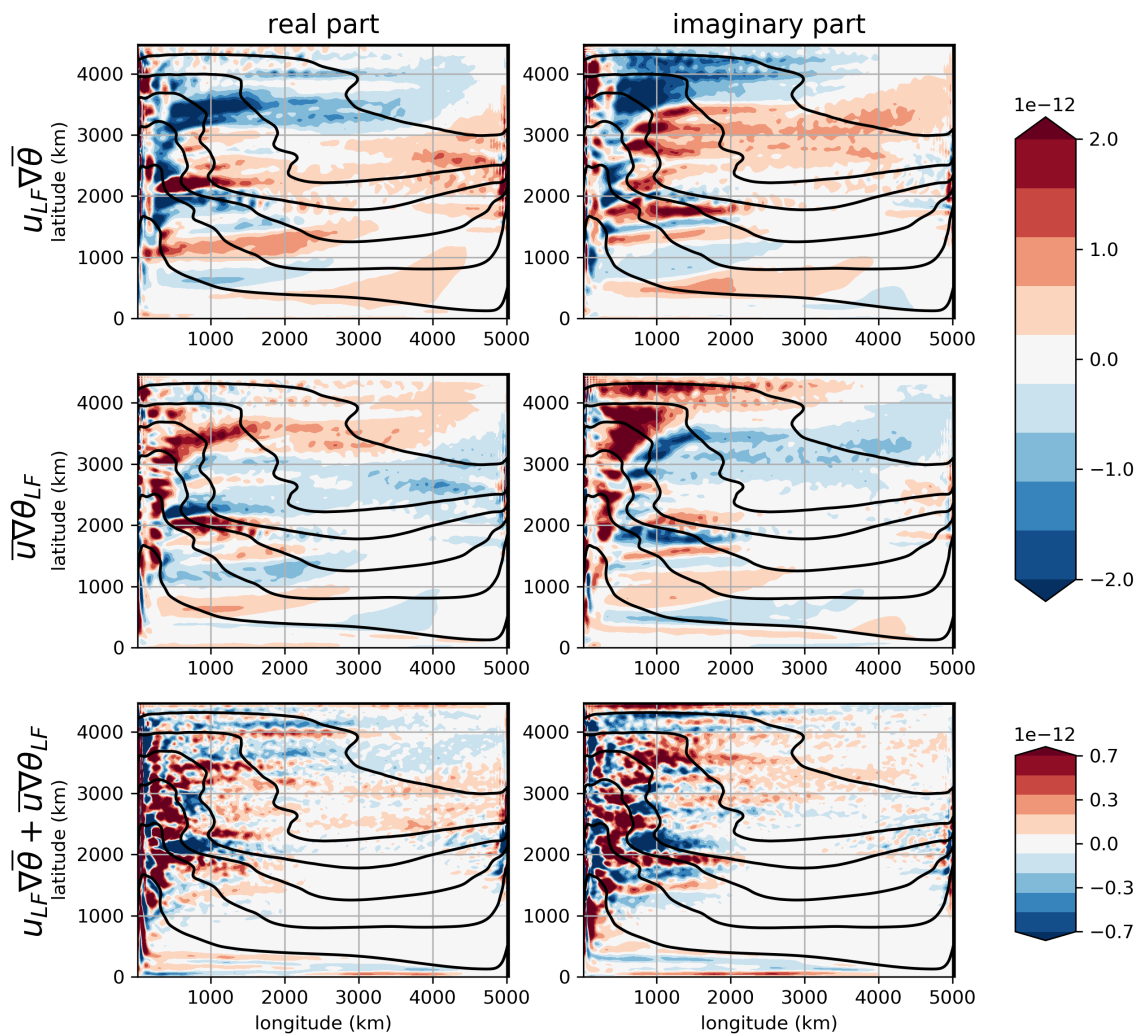


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