Has the impact of the Earth's orbital variation at sub-myriadal timescales been underestimated?

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Abstract

It is well established that the Earth's long-term orbital variation affects global climate through insolation changes, particularly with respect to glacial cycles. However, there is little research into the impact of this slowly varying insolation forcing on timescales well below the smallest main periodicity, the precession, of ~ 21,000 years. Here, intermediate complexity climate modelling demonstrates that allowing a smooth annual variation in orbital parameters and atmospheric CO_2 levels through 4,000 years of a glacial climate simulation leads to significantly more variability on a range of timescales in key measures of oceanic climate, compared to simulations with combinations of fixed or variable forcing. This includes a more vigorous ocean overturning circulation on a global basis. These results raise the possibility that past climate change signals, often ascribed to internal natural variability, may instead be initiated by slowly-evolving external orbital and atmospheric CO_2 variation.

Keywords: climate variability, solar insolation, long-term change, ocean overturning

1 Introduction

The Milankovitch cycles in the eccentricity, obliquity and precession of the Earth's orbit are linked to glacial cycling in the Earth's climate (Hayes et al 1976), as well as its modulation over time (Saedeleer et al 2013; Lisiecki 2014). Coupled climate models have shown that this long-term variation in insolation is key to explaining the onset and development of the last glacial cycles (Renssen et al 2005; Berger and Loutre 2010; Ganopolski et al 2010; Singarayer and Valdes 2010; Smith and Gregory 2012; Kessler et al. 2020). However, the dominant periodicities in the Milankovitch cycles are many thousands of years (c. 100,000, 41,000 and 21,000 years; Berger, 1988) so that consideration of the orbital impact on climate has mostly focused on change over tens to hundreds of thousands of years. Nevertheless, there are continuous slow modulations of insolation due to orbital change, both in absolute terms and its geographic spread (Berger 1988). Even relatively small changes in insolation may affect sub-centennial-scale climate variability (Munz et al 2015), and possibly interact with millennial-scale internal oscillations within ice sheet dynamics (Bügelmayer-Blaschek et al 2016) and the climate generally (Renssen et al 2005; Kessler et al 2020). Sensitivity experiments of orbital impacts on climate using climate modelling have focused on these longer time-scales, incorporating slow (Renssen et al 2005; Smith and Gregory 2012; Kessler et al 2020) or step (Singrayer and Valdes 2010) changes in orbital parameters, or concentrating on long-term changes in specific climate factors (Berger and Loutre 2010; Ganopolski et al 2010), or unusual combinations of orbit and albedo (Friedrich et al 2010). Here, we investigate the response of the climate system to these continual slow changes in orbital parameters and atmospheric CO_2 on sub-millennial scales using FRUGAL, an intermediate complexity climate model (Levine and Bigg 2008). The term "sub-myriadal" is used to describe the timescales of centuries to thousands of years that are of interest here, deriving from the Greek word for 10,000, a myriad.

2 Materials and Methods

2.1 Model description

The coupled ocean-atmosphere-sea ice-iceberg intermediate complexity model used has a curvilinear coordinate grid, with its North Pole displaced to central Greenland to increase horizontal resolution in the North Atlantic and Arctic (Wadley and Bigg 2000), in this case to 1° locally but ~7° in the Southern Hemisphere (figure 1). This leads to its name – the Fine ResolUtion Greenland And Labrador (FRUGAL) model. For efficient integration of the variable resolution grid, time-step length is a function of grid spacing (Wadley and Bigg 2000). The ocean model component is a free surface general circulation model, with 19, variable thickness, levels in the vertical. There is temperature and salinity mixing in the horizontal, vertical and along isoneutral surfaces (Levine & Bigg 2008). The topography is averaged from a base 5' data set (ETOPO 1986), with modification for key sill depths, but altered for glacial conditions (Peltier 1994), combined with a 120 m lowering of sea level (Siddall et al 2003). The sea-ice component is a thermodynamic model with simple advection (Wadley and Bigg 2002). The iceberg model is a dynamical and thermodynamical representation of a suite of icebergs whose trajectories are modelled through their lifetime, with a scaling parameter to represent estimates of the fluxes from different calving sources; this is coupled to the ocean component of FRUGAL (Levine and Bigg 2008). While iceberg freshwater contributions are included in the simulations there is no coupling of the model climate to the land ice field, meaning that iceberg fluxes remain the same throughout – any observed variability in model simulations cannot be linked to Heinrich-like variance in iceberg flux.

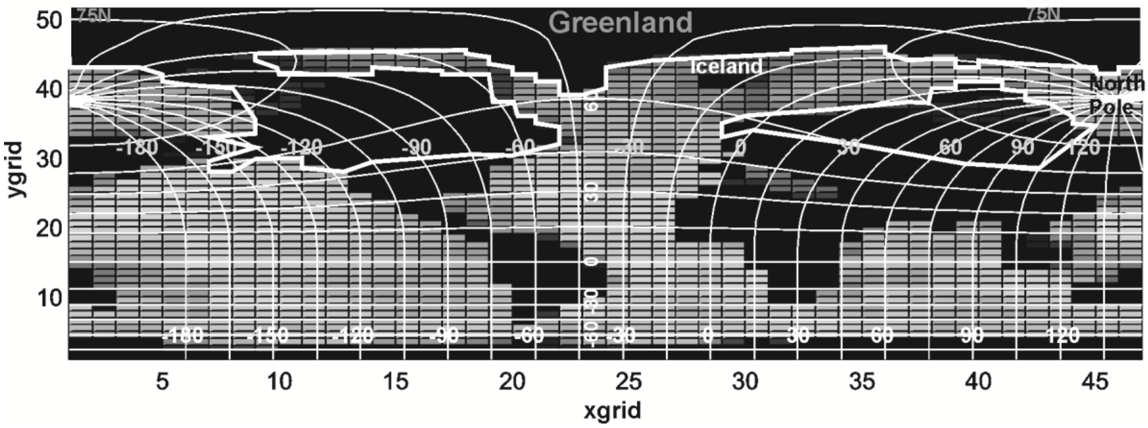

Figure 1. FRUGAL grid, showing latitude and longitude lines, and some location names for orientation. Boundaries of the land ice cover used is shown by the bold white line.

Note that the Strait of Gibraltar between the Atlantic and Mediterranean is explicitly represented in the ocean model. While this leads to a width of several times the actual gap, there is good reproduction of the exchange magnitude in the current climate (Rogerson et al 2012). This is because the exchange is hydraulically controlled and ocean models can reproduce such control even if the gap is too large, as in this case (Wadley and Bigg 1994).

The atmospheric part of FRUGAL is an adaptation of the energy and moisture balance model of the UViC Earth System Model (Fanning and Weaver 1996) that allows for advection of water vapour. The atmospheric model includes parameterizations of clouds, mountains, land-ice, and land hydrology. The simulations discussed in this paper use a monthly varying wind stress from an LGM atmospheric model simulation (Dong and Valdes 1998); this has no feedback from the sea surface temperature field and so remains the same seasonally varying field throughout all simulations.

2.2 Model experiments

To focus on climate variability at sub-myriadal scales dominated largely by orbital change alone, we considered a 4000 year period centred early during the peak of the last glacial period, i.e., from 31-27 ka (thousand years before present), when mean global climate and ice volume changed relatively little (figure 2), but the orbital parameters continued to evolve (figure 3). Four simulations of 4000 years duration each were performed, all started from the same 5000 year spinup with fixed, 31ka orbital parameters and an atmospheric CO_2 concentration of 200 ppm. The first simulation was a Control experiment, with the orbital parameters and atmospheric CO_2 concentration remaining fixed throughout. There is a Total climate simulation, where orbital parameters and atmospheric CO_2 both vary slowly, smoothly and continuously over time, and running from 31 ka to 27 ka. Values for these forcing parameters changed every timestep, with the CO_2 being interpolated between values taken from ice core values every thousand years (table 1), and orbital parameters being calculated afresh each year (Berger 1988), as shown in figure 3. We also did a simulation where atmospheric CO_2 is held fixed at 200 ppm, but the orbital parameters vary smoothly with time as in Total; this is denoted simulation Orbit. Finally, there is a simulation where the orbit is fixed at 31 ka values, but the atmospheric CO_2 varies smoothly as in simulation Total; this is denoted as simulation CO2. As we are here concerned with the sensitivity of the climate system due to the slow evolution of the orbital parameters and atmospheric CO_2, during peak glacial times alone, we keep most model forcing functions the same throughout each simulation. The sea level, topographic grid, ice extent and monthly-varying atmospheric forcing field is as used by a previous simulation of FRUGAL for the Last Glacial Maximum (Levine and Bigg 2008), and did not evolve in the simulations. This is reasonable, as global climate as seen in ice extent, sea level and mean global temperature remains relatively stable through 31-17 ka (Siddall et al 2003; Lüthi et al 2008).

Between them, these four simulations explore the sensitivity of the intermediate complexity model with and without the external radiative forcing determined by the orbital parameters and atmospheric CO_2. They also allow determination of whether any sensitivity found depends on orbital changes, CO_2 or requires the contribution of slow variation in both. Each of the three perturbed simulations takes between 500 and 1000 years to reach a new climate equilibrium, so the quantitative elements of the results and discussion sections use the last 2000 years of each simulation for the run comparisons.

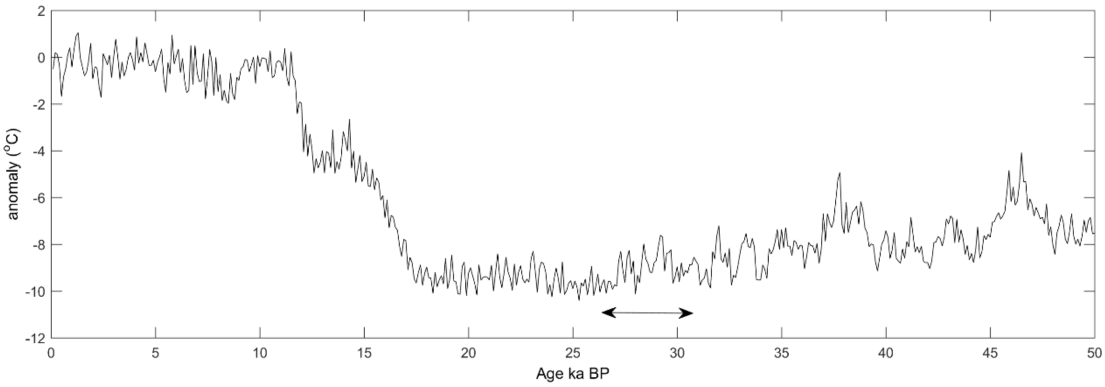

Figure 2. Antarctic Dome C surface air temperature anomaly (°C), relative to the year 2000 (-50 yr BP), over the past 50 ka (Lüthi et al 2008). The horizontal arrow shows the duration of the simulations in this investigation.

Table 1. Atmospheric CO_2 values used in simulations Total and CO2, where this varies with time. These values are in ppm and interpolated from the Vostok ice core record (Lüthi et al 2008).

Approx. age (ka)	Atmos. CO_2 (ppm)
26	188.8
27	191.4
28	194.6
29	197.8
30	200.9
31	204.0

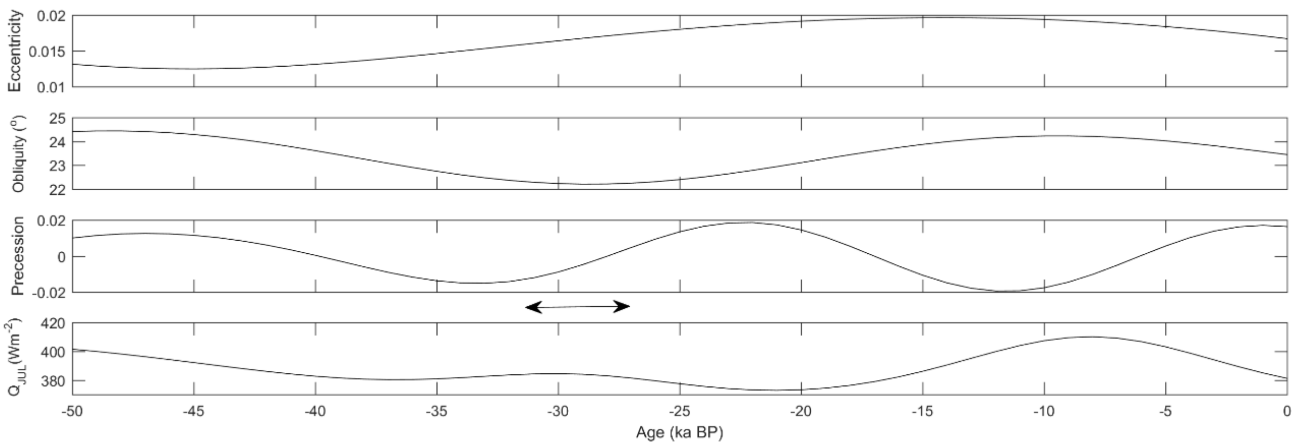

Figure 3. Variation of orbital parameters over the past 50,000 years, with the impact on daily July insolation at the top of the atmosphere at 65°N. The simulation period is marked by the double-headed arrow, with the results of later figures coming from model data of the period 29-27 ka.

3 Results

3.1 Sensitivity experiments

While the orbital parameter values for the Control experiment differ slightly from those used in the spin-up end-state of Levine and Bigg (2008), Bigg et al (2012) and Wilton et al (2021), the basic control ocean and atmospheric circulations are very similar across these FRUGAL experiments. These papers have shown that the FRUGAL model does a reasonable job of reproducing modern and past ocean circulations.The mean values of some large scale ocean flow features of the Control experiment are shown in Table 2, with the measure of the annual range of the seasonal cycle for these quantities shown in Table 3.

The Control run has its main overturning cell in the Southern Ocean (see "S. Hemisphere overturning" in figure 4 and tables 2-3), with smaller, but similar strength, overturning cells in the North Atlantic and North Pacific ("Atlantic MOC" and "Pacific MOC" respectively). These have significant annual cycles, particularly in the North Pacific, where the size of the annual cycle is 50% greater than the mean overturning strength ("Pacific MOC" in table 3). As today, the mean Northern Hemisphere sea-ice area is substantially larger than that in the Southern Hemisphere, but both have large annual cycles, with the Southern Ocean sea-ice almost disappearing in summer, but reaching double its mean area in winter. The various quantities show a degree of interannual variability, but of limited degree, principally because the FRUGAL intermediate complexity climate model tends not to have strong feedbacks within it (Balan et al 2011). This interannual variability typically has a standard deviation of only around 0.5% of the mean.

The sensitivity experiments show interesting comparisons to the Control run (see Tables 2-3). Allowing the atmospheric CO_2 or orbit to vary over time alone leads to only minor changes to the mean flows, annual cycles and the degree of variability (quantified in Tables 2-3 by the standard deviation). However, in simulation Total, where atmospheric CO_2 and orbital forcing evolve over time, there are changes to aspects of the global ocean circulation, and to the degree of interannual variability, the latter of which changes by various amounts between factors of two to ten, depending on the variable concerned. For example, the Pacific MOC changes in magnitude by

~22% (from 15.39 to 19.5 Sv) and the variability increases by a factor of 14 – from 0.02-0.28 Sv. Essentially, allowing the feedbacks between radiational anomalies from insolation with those associated with CO_2 in the model causes a speed-up of the ocean circulation, with enhanced overturning in all basins (figure 4), and stronger inter-basin exchanges between the Pacific and Indian Oceans and between the Mediterranean and Atlantic (figure 5). The latter change leads to a stronger North Atlantic subtropical gyre, despite the wind forcing remaining unchanged. These enhanced ocean flows result in sea-ice declines globally.

Table 2. The mean and standard deviation for major large-scale variables of the simulations, calculated over the last 2000 years of the 4000 year runs. SAT: Surface Air Temperature; MOC: Meridional Overturning Circulation. “Sv” is a Sverdrup and equal to 10^6 m^3 s^{-1}.

Variable	Control	CO2	Orbit	Total
Global SAT (°C)	7.26±0.01	7.33±0.00	7.27±0.00	7.30±0.02
Drake Passage (Sv)	95.16±0.57	94.59±0.01	93.46±0.06	100.64±2.24
Indonesian Throughflow (Sv)	13.58±0.05	13.48±0.01	13.31±0.01	14.13±0.55
Strait of Gibraltar (Sv)	1.93±0.02	2.00±0.02	1.95±0.02	2.76±0.04
Atlantic MOC (Sv)	9.83±0.08	9.71±0.08	9.32±0.05	15.00±0.24
Pacific MOC (Sv)	10.37±0.02	9.80±0.03	10.25±0.05	19.78±0.19
S. Hemisphere overturning (Sv)	66.10±0.33	67.34±0.02	65.58±0.08	74.92±0.90
Atlantic subtropical gyre (Sv)	11.76±0.02	11.76±0.02	11.76±0.01	12.51±0.10
Atlantic subpolar gyre (Sv)	4.07±0.01	4.06±0.01	4.09±0.00	4.00±0.01
Pacific subtropical gyre (Sv)	11.48±0.02	11.45±0.01	11.50±0.01	11.65±0.12
Pacific subpolar gyre (Sv)	0.67±0.01	0.67±0.01	0.67±0.01	0.65±0.01
N. Hemisphere sea-ice area (10^6 km^2)	25.23±0.14	25.42±0.02	25.46±0.03	24.15±0.06
N. Hemisphere land ice area (10^6 km^2)	26.32±0.02	26.24±0.01	26.33±0.00	26.17±0.05
S. Hemisphere sea-ice area (10^6 km^2)	8.58±0.14	7.78±0.02	8.44±0.02	7.34±0.26

Table 3. The annual range and standard deviation for the seasonal cycle of major large-scale variables of the simulations, calculated over the last 2000 years of the 4000 year runs. SAT: Surface Air Temperature; MOC: Meridional Overturning Circulation.

Variable	Control	CO2	Orbit	Total
SAT (°C)	0.92±0.00	0.92±0.00	0.92±0.00	0.92±0.00
Drake Passage (Sv)	5.40±0.05	4.90±0.01	5.13±0.03	5.30±0.11
Indonesian Throughflow (Sv)	2.15±0.01	2.16±0.01	2.17±0.01	2.09±0.02
Strait of Gibraltar (Sv)	0.48±0.00	0.46±0.01	0.48±0.00	0.44±0.01
Atlantic MOC (Sv)	4.67±0.08	4.87±0.04	4.82±0.03	3.64±0.05
Pacific MOC (Sv)	15.90±0.02	14.57±0.07	15.69±0.07	19.50±0.28
S. Hemisphere overturning (Sv)	10.80±0.02	10.80±0.01	10.81±0.03	10.01±0.14
Atlantic subtropical gyre (Sv)	4.53±0.03	4.35±0.04	4.50±0.04	4.85±0.03
Atlantic subpolar gyre (Sv)	0.84±0.01	0.88±0.01	0.86±0.01	0.86±0.01
Pacific subtropical gyre (Sv)	13.76±0.03	14.04±0.03	13.88±0.06	12.01±0.21
Pacific subpolar gyre (Sv)	1.56±0.03	1.50±0.03	1.56±0.03	1.56±0.06
N. Hemisphere sea-ice area (Sv)	9.39±0.06	7.93±0.09	9.20±0.07	8.62±0.18
N. Hemisphere land ice area (Sv)	1.6±0.00	1.66±0.00	1.6±0.00	1.52±0.11
S. Hemisphere sea-ice area (Sv)	14.62±0.19	14.09±0.03	13.81±0.11	12.54±0.51

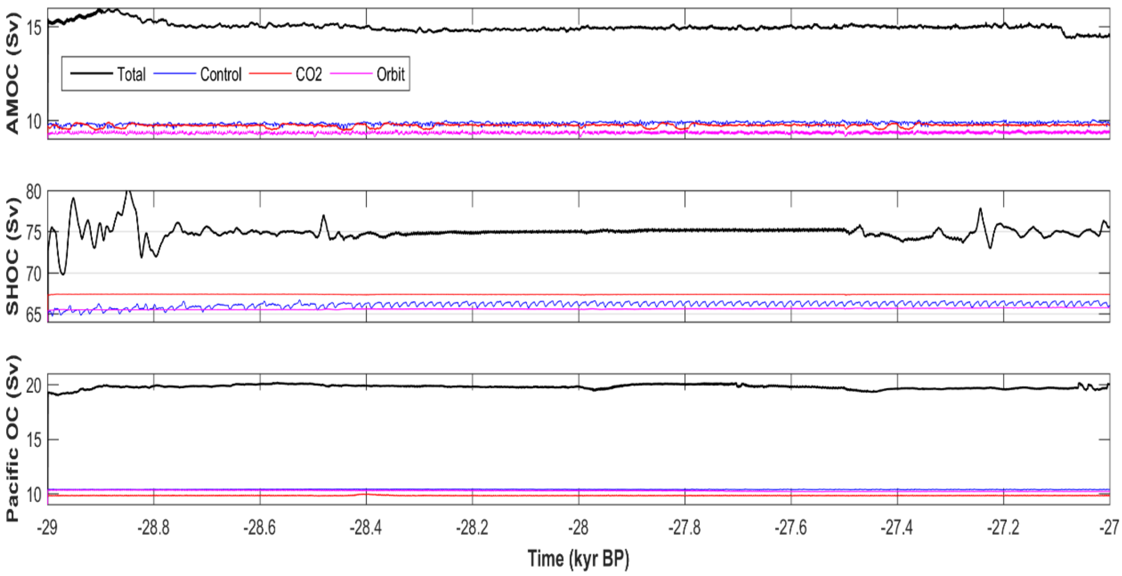

Figure 4. Annual mean of main ocean basin overturning circulations for last 2000 years of each simulation (see table 2). Top panel – Atlantic, centre – Southern Ocean, lower – Pacific.

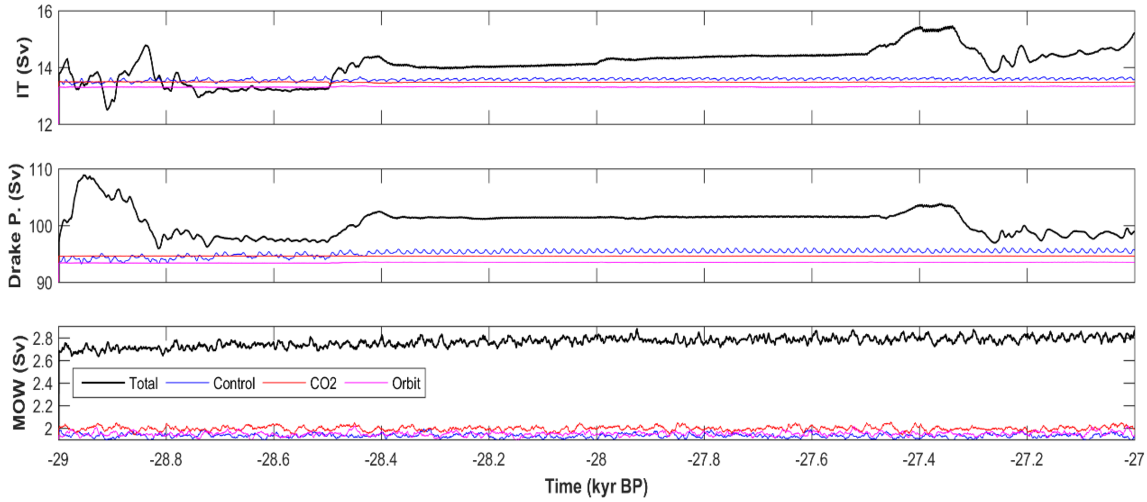

Figure 5. Annual means of inter-basin exchanges for last 2000 years of each simulation (see Table 1). Top panel – Indonesian Throughflow, centre – Drake Passage, lower – Mediterranean Outflow.

3.2 Radiational feedbacks

How does the combined insolation and CO_2 feedback lead to, in many areas, enhanced variability, and ultimately to a different, more vigorous ocean circulation, given that the Global SAT is essentially identical in all sensitivity tests (table 2)? Hints to the answer can be seen in example spectra of the monthly timeseries of some major integrative ocean circulation measures (figure 6). The Control, CO2 and Orbit runs all have flat spectra for the interannual period timescales, suggesting similar behaviour between the simulations. Individually changing the CO_2 or orbital parameters continually appears to have similar impacts on the size of the response of the ocean system compared to the natural internal variability of this model's Earth system. However, the Total simulation is notable by showing enhanced power in the century or longer timescales. This is more the case in large-scale measures of ocean circulation rather than those at local scale or directly determined by fluxes without much feedback with other processes (compare figures 6 and 7).

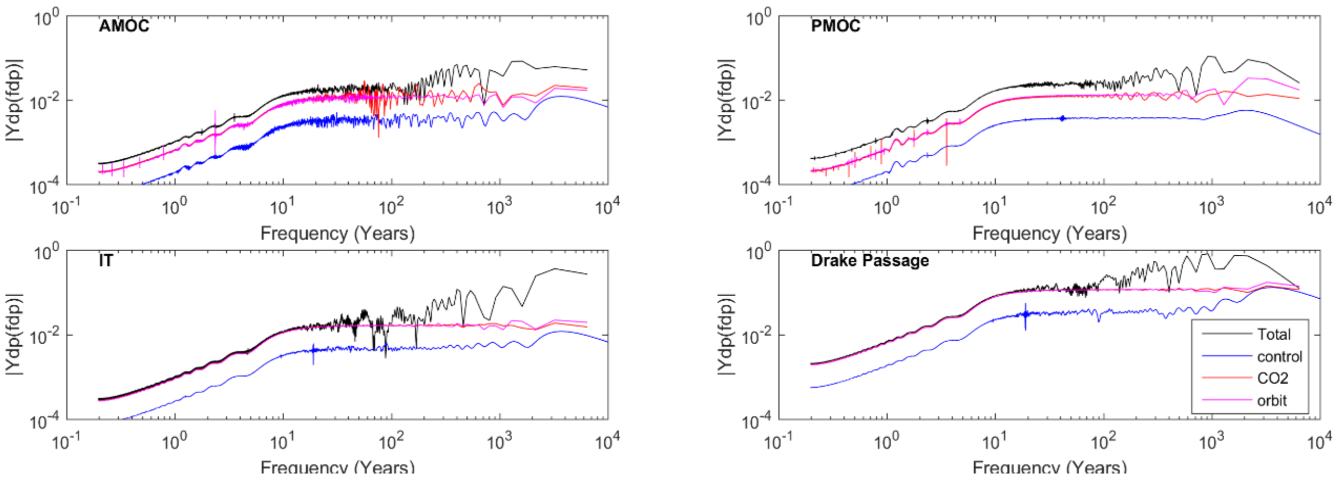

Figure 6. Fast Fourier Transform spectra for the monthly time series of the principal ocean overturning and basin exchange fluxes of the simulations: a) Atlantic overturning, b) Pacific overturning, c) Indonesian Throughflow, d) Drake Passage.

The combination of radiational forcing from outside the Earth (insolation) with perturbation to that for air-sea exchange (CO_2) leads to enhanced feedbacks within the large ocean basin scale, and therefore the exchanges between oceans, as seen in the enhanced power at the centennial and longer scales in figure 6.

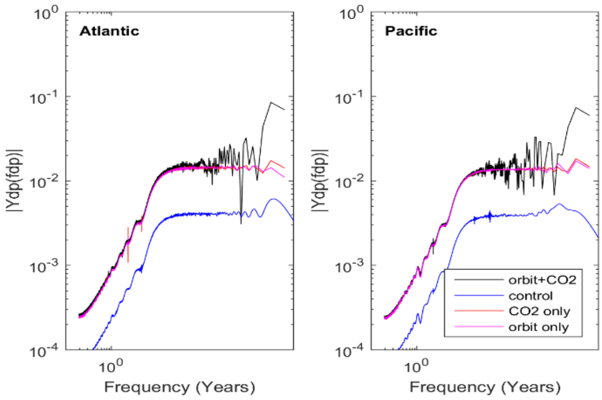

Figure 7. Fast Fourier Transform spectra for the monthly time series of the Northern Hemisphere a) Atlantic and b) Pacific sub-tropical gyres of the simulations

4 Discussion and Conclusion

The intermediate complexity model simulations shown here demonstrate the enhanced sensitivity of the climate system to the combination of the slow evolution of atmospheric CO_2 concentration and the slow evolution of the Earth's orbital parameters. Allowing the orbit and atmospheric CO_2 level to evolve annually leads to significantly higher centennial to sub-myriadal scale variability compared to a climate with either a fixed orbit or fixed atmospheric CO_2 (figure 6), but also enhanced inter-annual variability (table 3). It could be argued that this sensitivity to orbital and CO_2 parameters is a function of the model used. In particular, intermediate complexity earth system models tend to under-estimate natural variability, as has been seen in a number of studies comparing the responses of a suite of climate models to common forcing (Balan et al 2011; Flato et al 2013; Weaver et al 2012; Nikolva et al 2013). However, the FRUGAL intermediate complexity model has been shown to behave in similar ways to other climate models under contemporary forcing change (Balan et al 2011), as well as producing previous glacial simulations that are compatible with palaeodata (Levine and Bigg 2008; Bigg et al 2010; Bigg et al 2012; Wilton et al 2021). Indeed, the smoothing of natural variability in intermediate complexity models enhances the potential significance of the enhanced variability due to combined orbital and CO_2 variation found here.

These intermediate complexity climate model results suggest, therefore, that global radiational perturbation from atmospheric CO_2 variability, combined with slow evolution of insolation strength and geographic pattern, can lead to the initiation of internal variability within the climate system. This seems particularly centred on the sub-polar convection zones. Its impact is enhanced, due to this influence being in critical areas for producing long-lasting change through incorporation in the ocean thermohaline circulation. Occasional generation of relatively "abrupt change" in inter-basin exchanges is also seen (figure 5), presumably when some convection measure is pushed past a local tipping point in the phase space of ocean response.

Climate is not fixed, even during periods of relative stability like the Holocene. Internal variability due to stochastic forcing has been suggested as a reason for variation in the past (Khan and Ahmed 2015; Rypdal 2016), or interaction between different components of the climate system (Bond et al 1999; Bügelmayer-Blaschek et al 2016). These mechanisms for forcing climate variability seem likely. However, the continuing presence of slowly varying changes, and interactions, of the dominant solar forcing of the whole climate system through orbital changes with global atmospheric CO_2 levels is also a feasible means for promoting climate variability. The results of this modelling study suggest it largely acts through this global radiational change particularly modifying the sub-polar ocean convection regimes, an area also noted as being important for millennial-scale internal variability by Renssen et al (2005) and Kessler et al (2020). The peak of the last glacial cycle was a time of fairly small orbital change (figure 3), with at most ~ 20 W m^{-2} variation in summer insolation at 65°N, and minor atmospheric CO_2 level (Lüthi et al 2008; table 1). However, since the switch in glacial cycling to roughly 100,000 years around 800 ka BP, there have been frequent periods with either significantly larger orbital change (Bigg 2016) and/or atmospheric CO_2 (Lüthi et al 2008) over similar periods of ~15,000 years. Strong variability associated with polar fluxes of ice-rafted debris can be seen at a number of sites over this period (Bigg 2016). While they may have had other causes, and the model of Kessler et al (2020) suggests these sorts of changes are largely driven by the orbital changes alone, the results here suggest that these and other centennial-sub-myriadal climate signals are likely to be enhanced or initiated by the combined slow evolution of global CO_2 levels and orbital forcing. This enhancement of variability due to changes in “external” global forcing found here may also have implications for the robustness of the findings of a range of previous millennial-scale climate simulations and tests of perturbation responses. However, it should be noted that this enhanced model variability is centennial to sub-myriadal scale (figure 6), so has little implication for recent global warming attribution.

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