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Dansgaard-Oeschger events in climate models: Review and baseline MIS3 protocol

Irene Malmierca-Vallet[1], **Louise C. Sime**[1], and **the D-O community members**[+]

[1]British Antarctic Survey, Cambridge, U.K.

[+]A full list of authors appears at the end of the paper

Correspondence: Irene Malmierca-Vallet (irealm37@bas.ac.uk)

Abstract.

Dansgaard-Oeschger (D-O) events, millennial-scale climate oscillations between stadial and interstadial conditions (of up to 10-15°C in amplitude at high northern latitudes), occurred throughout the Marine Isotope Stage 3 (MIS3; 27.8 – 59.4 ka) period. The climate modelling community up to now has not been able to answer the question: Are our climate models too stable to simulate D-O events? To address this, this manuscript lays the ground-work for a MIS3 D-O protocol for general circulation models which are used in the International Panel for Climate Change (IPCC) assessments. We review: D-O terminology, community progress on simulating D-O events in these IPCC-class models (processes and published examples), and evidence about the boundary conditions under which D-O events occur. We find that no model exhibits D-O like behaviour under pre-industrial conditions. Some, but not all, models exhibit D-O like oscillations under MIS3 and/or full glacial conditions. Greenhouse gases and ice-sheet configurations are crucial. However most models have not run simulations of long enough duration to be sure which models show D-O like behaviour, under either MIS3 or full glacial states. We propose a MIS3 baseline protocol at 38 ky (38 to 32 ky) period, which (1) shows a regular sequence of D-O events, and (2) features the intermediate ice-sheet configuration and medium-to-low MIS3 greenhouse gas values which our review suggests are most conducive to D-O like behaviour in models. We also provide a protocol for a second “kicked Heinrich meltwater” experiment, since previous work suggests that this variant may be helpful in preconditioning a state in models which is conducive to D-O events. This review and protocol is intended to provide modelling groups investigating MIS3 D-O oscillations with a common framework.

1 Introduction

During a Dansgaard-Oeschger (D-O) event, Greenland transitions between cold stadial (GS) and warmer Greenland Interstadial (GI) conditions. The warming can occur within a decade (Kindler et al., 2014; Huber et al., 2006), whilst cooling occurs over a much longer period that is typically several centuries in length. During a warming phase, surface air temperatures over Greenland increase by 10-15°C (Andersen et al., 2006; Kindler et al., 2014; Huber et al., 2006). D-O events are best documented during Marine Isotope Stage 3 (MIS3; between 27.8 – 59.4 thousand of years BP, hereafter ka Goni and Harrison, 2010), including being recorded in several ice cores from Greenland (Fig. 1 Johnsen et al., 2001). Whilst the D-O event

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

recorded in these cores are renowned, the events are global in nature (Voelker et al., 2002; Sanchez Goñi and Harrison, 2010; Sánchez Goñi et al., 2017), with known climate signatures including imprints in surface temperature and the hydrological cycle at high northern latitudes (Andersen et al., 2004; Thomas et al., 2009; Seierstad et al., 2014), in the tropics (Deplazes et al., 2013; Baumgartner et al., 2014; Adolphi et al., 2018), in Eurasia (Genty et al., 2003; Wang et al., 2008; Jacobel et al., 2017; Rousseau et al., 2017), and in North and South America (Wang et al., 2004; Wagner et al., 2010; Asmerom et al., 2010; Deplazes et al., 2013; Vanneste et al., 2015). While there are no Greenland ice core records of the previous glacial (MIS6 around 140-190 ka), speleothems and Antarctic ice cores indicate that it is extremely likely that D-O events also occurred during MIS6 and earlier glacial periods (Lang et al., 1999; Uriarte, 2019; Landais et al., 2004; Turner and Marshall, 2011; Barker et al., 2011; Lambert et al., 2012). This observational evidence shows that D-O millennial timescale D-O events do not occur under interglacial or full Last Glacial Maximum conditions (Galaasen et al., 2014; Tzedakis et al., 2018; Galaasen et al., 2020).

In 2011, Valdes (2011) argued that climate models used in the assessments of the Intergovernmental Panel on Climate Change (IPCC) have not proved their ability to simulate D-O events. This has several implications for the delivery of accurate projections of climate change, within the context of tipping points and abrupt climate change (Brovkin et al., 2021). Whilst in the intervening years a number of models have captured key features of D-O events through AMOC hysteresis behaviour and/or produced D-O type millennial-scale variability under a range of forcings (Brown and Galbraith, 2016; Galbraith and de Lavergne, 2019; Klockmann et al., 2018; Peltier et al., 2020; Armstrong et al., 2021; Zhang et al., 2021; Vettoretti et al., 2022), we still do not know if climate models are too stable because too few models have run an appropriate simulation. This deficiency is related to both the computational expense which prevents models from being run for the longer time periods needed for investigating D-O events and to the lack of an agreed appropriate experimental set-up. The limited knowledge of pre-Last Glacial Maximum (LGM) boundary conditions, in particular in the case of the ice sheet height and distribution, makes it challenging to generate an appropriate MIS3 experimental set-up.

Whether models can simulate abrupt changes is a crucial research question: if the current IPCC-class models are too stable to simulate D-O events, their ability to predict future abrupt transitions, and their use in identifying tipping points is doubtful. For example, a tipping point may have been recently reached in the Arctic's Barents Sea (Barton et al., 2018; Tesi et al., 2021); sea ice loss in the area is linked with enhanced heat transport via an intensified throughflow, or "Atlantification" (Årthun et al., 2012; Polyakov et al., 2017). In addition, future enhanced precipitation, decline in Arctic sea ice and melting of glaciers and ice sheets could intensify the supply of freshwater to the North Atlantic and Arctic which could lead to the reorganization of the Atlantic circulation and tip the energy distribution between South and North in a similar way as occurred during D-O events (Lenton et al., 2008). If climate models do not reliably simulate past tipping events, it suggests that simulations of the coming century may be giving us a false sense of security.

Coupled Model Intercomparison Project (CMIP) coordinates and designs climate model protocols for the past, present and future climates, and has become an indispensable tool to facilitate our understanding of climate change (IPCC, 2013; Eyring et al., 2016). The Paleoclimate Model Intercomparison Project 4 (PMIP4) is one of the individual Model Intercomparison

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

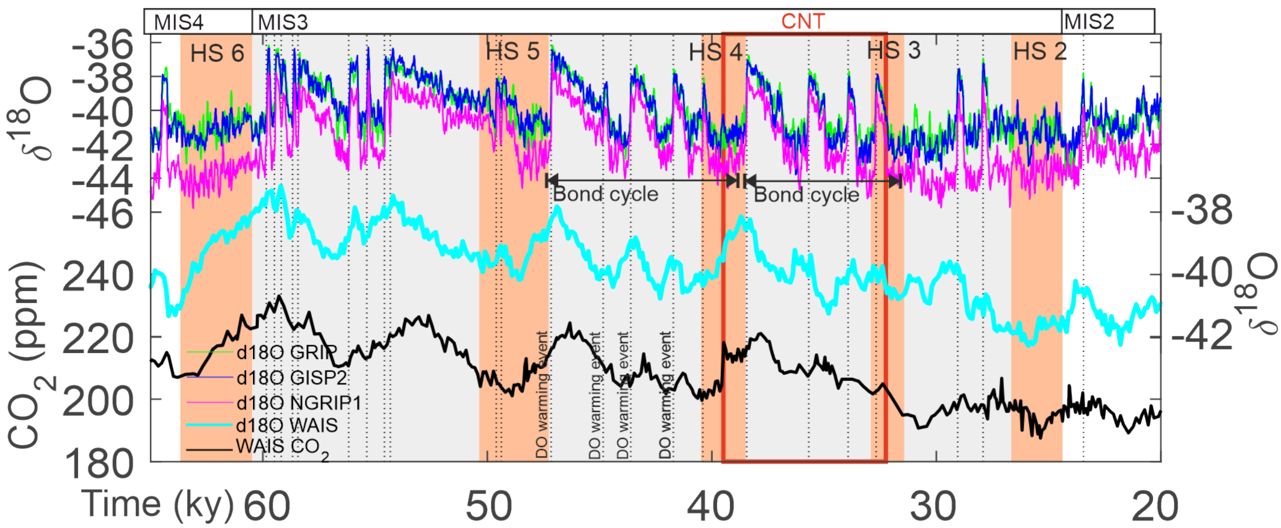

Figure 1. MIS3 ice core records and nomenclature. Stable water isotope and CO_2 measurements from Antarctic and Greenland ice cores (Bauska et al., 2021; NGRIP Project Members, 2004; Kindler et al., 2014). See also Table 1 for D-O nomenclature. The "cnt" red box indicates the 38 to 32 ka period proposed for the MIS3 baseline experiment

Projects which took part in CMIP6 (Kageyama et al., 2018). The design of a common MIS3 experimental protocol would allow the modelling community to address the questions posed above.

This manuscript compiles current information about unforced D-O like oscillations in IPCC-class models and discusses the boundary conditions and mechanisms responsible for these oscillations. Given the nomenclature on D-O events varies throughout the literature. Firstly, Table 1 and Figure 1 provide a framework for a more consistent terminology. Secondly, we review the literature to ascertain whether models reproduce D-O like events under MIS3, or other, climate conditions. We then use this information to develop a protocol for simulations of D-O events. This protocol focuses on Marine Isotope Stage 3 (MIS3) partly because of the excellent records of D-O events and boundary condition during this period (Schulz et al., 1999) but also because, as our synthesis shows, MIS3 conditions are also conducive to promoting D-O like events in some IPCC-class models. In addition to the protocol for a baseline simulation, we also outline a protocol for a Heinrich event (Bond cycle event one type; Table 1) preconditioned variant. These protocols provide a common framework for model experiments to explore cold-period instabilities using commonly specified greenhouse gas (GHG), ice sheet, insolation, and freshwater-related forcings.

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table 1: D-O event nomenclature

Term	Description
Abrupt climate change	A large-scale change in the climate system that takes place over a few decades or less, persists (or is anticipated to persist) for at least a few decades, and causes substantial disruptions in natural systems (Pörtner et al., 2019)
Tipping element	Large-scale components (subcontinental length scale, around 1,000 km) in the Earth system that can go through a tipping point (Lenton et al., 2008). Examples of tipping elements in the Earth system include: Greenland and West Antarctic ice sheet and the Atlantic medidional overturning circulation (AMOC).
Tipping point	A level of change in system properties beyond which a system reorganises, often in a nonlinear manner, and does not return to the initial state even if the drivers of the change are abated. For the climate system, the term refers to a critical threshold when global or regional climate changes from one stable state to another stable state (Pörtner et al., 2019)
Oscillation	The earth's climate undergoes regular cyclical changes. Those related to changes in the orbit of the earth around the sun have a periodicity of tens to hundreds of thousands of years. Those related to the seasons have an annual pattern. Superimposed on these are a number of less regular oscillations. It is not clear that series of D-O events are oscillations in the strict sense.
Stadial-Interstadial	The North Atlantic climate of MIS3 is seperated into warm interstadials and cold stadial periods which generally last several centuries to millennia. The warm and cold stages are described as Greenland Interstadial (GI) and Greenland Stadials (GS).
D-O events	May refer to the abrupt warming or the whole interstadial, sometimes including the transitions back into stadial condition. D-O events should be sub-classed as D-O warming and D-O cooling events depending on whether they mark the GS to GI transition or vice versa.
Heinrich or H events	Large iceberg calving events, marked by Heinrich layers of ice-rafted debris (IRD) across large regions of the North Atlantic (Heinrich, 1988). H events occur during GS between around 40-50°N (Hemming, 2004). They are huge (iceberg) freshwater releases into the North Atlantic and have a role in D-O oscillations (Flückiger et al., 2006). Interestingly the Greenland ice core inferred temperature record shows little or no impact from H-events (Rhodes et al., 2015); and ice core records can exhibit a methane signature at the onset of an H-event before a stadial has begun. Thus, H events do not necessarily cause D-O events (Capron et al., 2021).
Heinrich Stadial	If a GS is punctuated by a Heinrich (H-)event, then it can be referred to as a Heinrich Stadial (HS), where the slowdown in the AMOC happens before ice rafted debris deposition (Henry et al., 2016). A cold phase may be termed a stadial (or GS) until an H-event occurs, then could be classed as a Heinrich Stadial. Indeed, stadials that also contain a H-event were referred to as "Heinrich Stadial" for a few years (Barker et al., 2009; Sanchez Goñi and Harrison, 2010; Stanford et al., 2011). However Andrews and Voelker (2018) suggests this nomenclature should be neglected in favour of the GS and GI terminology. A further issue arises as to whether an Heinrich event has a Laurentide or a Fennoscandian origin (Griem et al., 2019), but the general H-event terminology is currently not sub-classed.
Bond cycle	In the 1990s a connection was made between H-events and D-O events with the concept of the Bond Cycle: a train of D-O events, with a duration of around 10-15 kyr and decreasing in amplitude, following a H-event (Agosta and Compagnucci, 2016). Indeed, D-O events frequently appear grouped in Bond cycles, groups of up to four D-O warming then cooling events with a longer GI event followed by up to three shorter GIs, alternated with GSs (Bond et al., 1993; Lehman, 1993; Bond and Lotti, 1995) Bond cycles finish with a cold event, during which a H event takes place (Hemming, 2004). During MIS3, the individual D-O oscillations are seen in particularly clear Bond cycle clusters following H events H5 and H4 (Bond et al., 1997; Sakai and Peltier, 1996).

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

2 Review of spontaneous D-O type quasi-oscillations in coupled climate models

We compile published evidence of long unforced quasi-oscillations (in the Atlantic Meridional Circulation; AMOC) in IPCC-class models under all climate states in Table A1, alongside glacial boundary condition simulations which do not shown these (Table A3). This permits us to explore the questions of: what proportion of models exhibit D-O like behaviour; which boundary conditions are most conducive to this; and what mechanisms are common to the modelled D-O behaviours. A number of model simulations exhibit long unforced quasi-oscillations in the AMOC (Table. A1) though those that occur under pre-industrial conditions do not appear to be D-O like events. We deal with these first.

Under pre-industrial greenhouse gas (GHG) forcing and present-day ice sheets, spontaneous centennial-length cold events that last around 100-200 years occur in four IPCC-class models (Table A1). EC-Earth and Community Climate System Model version 4 (CCSM4) show high atmospheric blocking over the eastern subpolar gyre that causes a cold event under pre-industrial boundary conditions (Drijfhout et al., 2013; Kleppin et al., 2015, Table A1). ECHAM6-FESOM also produces cooling events under pre-industrial conditions due to sudden reductions of deep water convection and increase of sea ice cover in the Labrador Sea (Sidorenko et al., 2015). Changes in convection also occur in the Kiel Climate Model (KCM; Martin et al., 2015), however here centennial-scale variability of the AMOC is linked to variability in Southern Ocean convection. Unlike the CCSM4 and the EC-Earth models, the KCM and ECHAM6-FESOM studies do not indicate an active role of the atmosphere. Although these four models all show abrupt spontaneous cooling events under pre-industrial boundary conditions, these events do not have the typical saw-tooth characteristics, or longer timescales, of D-O type events.

Regular cycles of D-O type quasi-oscillations are found in UofT CCSM4 under LGM boundary conditions (Peltier and Vettoretti, 2014). The initiation of the abrupt D-O type warming events is associated with the opening of a large polynya over the Irminger Sea (Vettoretti and Peltier, 2016) (Table A1). During the first thousand years of the simulation as the model is spun up and the ocean cools to reach a state consistent with glacial boundary conditions, there are two thermal thresholds during which the strength of the AMOC rapidly reduces (see Figure 2 in Peltier et al., 2020). These abrupt transitions in the AMOC coincide with abrupt reductions in surface temperatures in the North Atlantic and abrupt expansions of sea ice coverage. During the second of these events the AMOC is reduced to approximately 12 Sv, about half its strength in the pre-industrial control (Peltier et al., 2020). This event may resemble the impact of a Heinrich event-like "kick" to the AMOC though no freshwater perturbation was imposed (Peltier et al., 2020). After this, the AMOC spontaneously exhibits D-O like quasi-oscillations (Peltier et al., 2020). The Peltier et al. (2020) salt oscillator is maintained by the salinity gradient between the subtropical gyre and the Northern North Atlantic, similar to that identified by Brown and Galbraith (2016) in a simulation with LGM CO_2 but pre-industrial ice sheets. Although UofT CCSM4 is the only model to show long unforced quasi-oscillations in the AMOC under full glacial conditions, most of the other PMIP4 LGM simulations (Kageyama et al., 2021a) have not been run long enough to be sure that such oscillations would not arise if they were run for longer (see Table B1). Having said that, ideally models should not show oscillatory D-O type behaviour when configured under a full glacial climate state, given that in reality D-O events do not occur under full glacial conditions (Huber et al., 2006; Galaasen et al., 2014; Kindler et al., 2014; Tzedakis et al., 2018).

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Under late glacial conditions, at 30 ka, a quasi-oscillating AMOC is produced by the HadCM3 model (Armstrong et al., 2021) and results from a North Atlantic salt oscillator mechanism similar to that in UofT CCSM4 (Peltier and Vettoretti, 2014; Vettoretti and Peltier, 2016; Peltier et al., 2020). Under intermediate glacial conditions (MIS3: 40-32 ka), the COSMOS model shows spontaneous millennial-scale climate oscillations triggered solely by orbitally driven insolation changes (Zhang et al., 2021). Variations in either obliquity or eccentricity-modulated precession lead to climate variations over the tropical and subpolar North Atlantic which exert opposite effects on AMOC strength, and hence result in an oscillatory climate regime (Zhang et al., 2021). The CM2Mc model also produces somewhat smoothed quasi-oscillating AMOC under intermediate MIS3-like boundary conditions, with a present-day ice sheet distribution in combination with a CO_2 concentration of 180 ppm and low obliquity (22°) (Brown and Galbraith, 2016; Galbraith and de Lavergne, 2019) (Table A1). The MPI-ESM model exhibits more abrupt D-O like quasi-oscillations with a present-day ice sheet distribution in combination with CO_2 concentrations ranging between 190-217 ppm (Table A1; Klockmann et al., 2018, 2020).

In contrast to the above, neither NorESM nor CCSM3 produce D-O type events or quasi-oscillations under MIS3 conditions (38 ka) (Table A3; Guo et al. (2019); Zhang and Prange (2020)). The NorESM MIS3 simulation is in a stable regime with strong convection in the Norwegian and Labrador seas and the model state appears to be far from a possible threshold (Guo et al., 2019). Zhang and Prange (2020) use the LGM ICE-5G ice sheet configuration (Peltier, 2004), with a high Laurentide Ice Sheet (at just over 4000 m) which may have contributed to a strong AMOC in the CCSM3 simulation, alongside its particular background climate.

In summary, IPCC-class models set up with pre-industrial or present-day conditions do not exhibit D-O type warming events, but can feature shorter centennial length cooling and warming events. This model behaviour is consistent with observations, since millennial timescale D-O events do not occur under interglacial conditions but periods of centennial-scale AMOC variability are present throughout several interglacials (Galaasen et al., 2014; Tzedakis et al., 2018; Galaasen et al., 2020). Some models which are set up with more MIS3 like conditions exhibit D-O type warming events, but some do not. Under full LGM conditions only one model (UoT-CCSM4) out of ten (PMIP4 LGM simulations: Kageyama et al. (2021a)) show spontaneous D-O type oscillations (Tables A1 and B1; Kageyama et al., 2021a; Peltier and Vettoretti, 2014).

Since it can take some time for D-O type oscillations to evolve, it is unclear if some models would develop such oscillations if they were run for longer (at least for 2000 model years). Of the thirty-eight LGM/MIS3-like simulations (Table B1; Kageyama et al., 2021a; Armstrong et al., 2021; Klockmann et al., 2018), sixteen simulations have been run for less than 2000 years (Table B1), which makes it difficult to tell whether any of these simulations are capable of, or likely to, exhibit D-O like behaviour under specific boundary conditions. In addition the duration of LGM/MIS3 simulations is currently inadequate, we note that the majority of CMIP6 models appear not to have performed any form of glacial period simulation. Thus, it is difficult to ascertain what proportion, or indeed which, models are capable of capturing D-O like behaviour, under any form of glacial period state (Table C1).

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

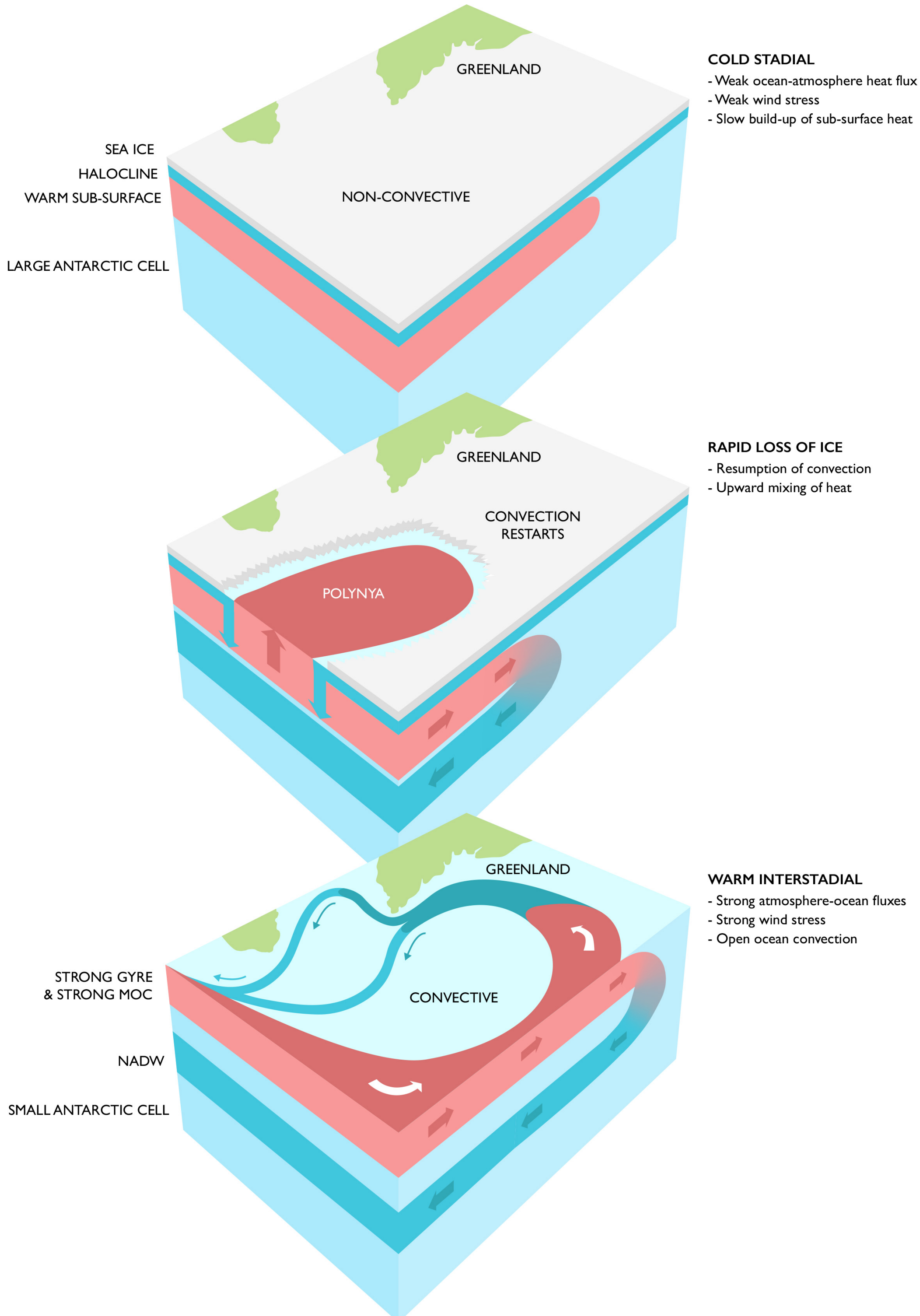

Figure 2. Schematic depicting the transition from GS to GI conditions *i.e.* a D-O warming event.

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

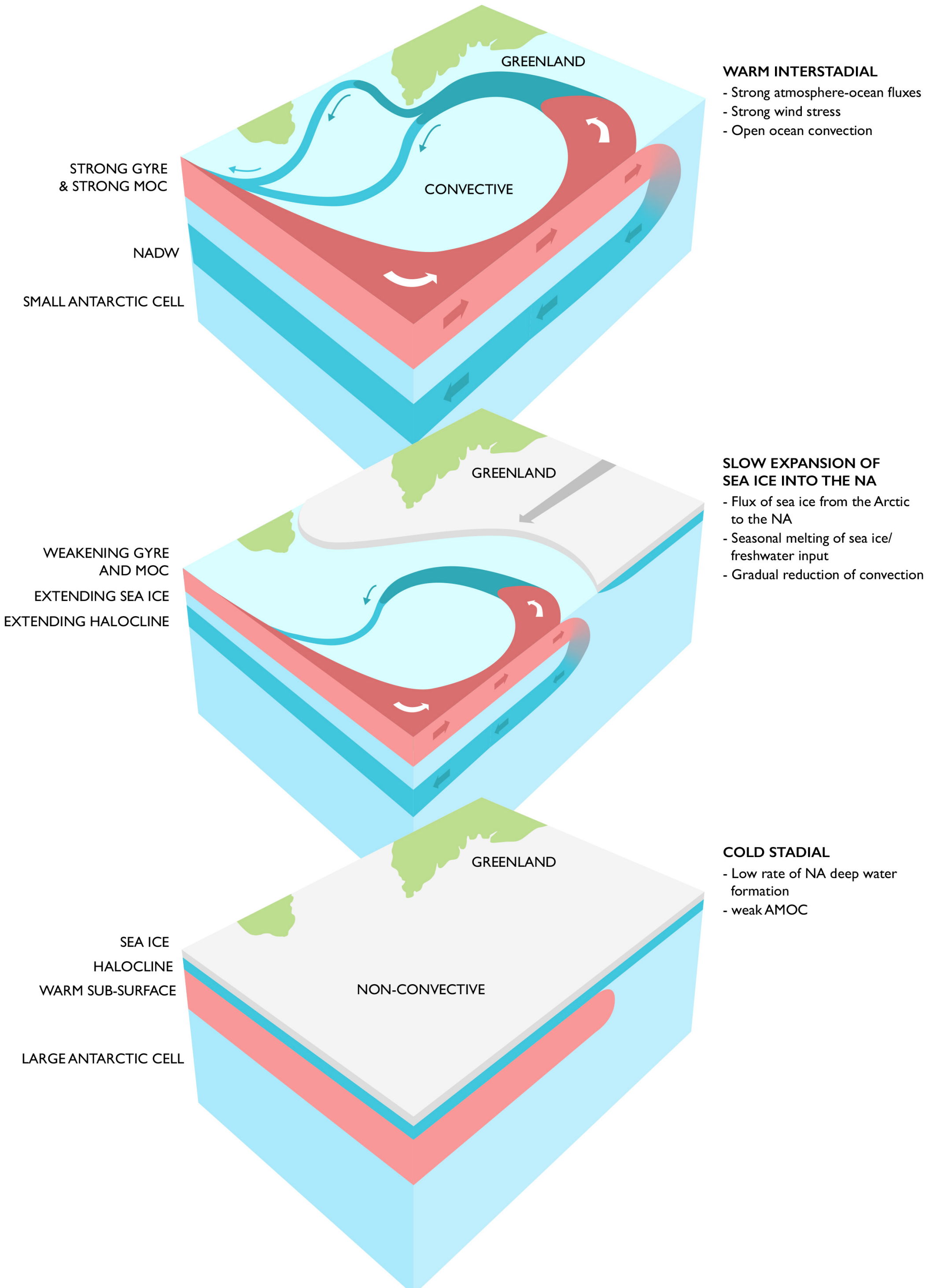

Figure 3. Schematic depicting the transition from GI to GS conditions *i.e.* a D-O cooling event.

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

2.1 The role of ocean and sea-ice feedbacks

Changes in the AMOC are crucial to the correct simulation of D-O events (Broecker and Peteet, 1985). The AMOC features stabilising positive feedbacks: a strong AMOC transports warm and salty water into the subpolar North Atlantic, thus weakening the stratification and also keeping the sea ice cover reduced (*e.g.* Rahmstorf, 2002; Clark et al., 2002). As a consequence, there is a large transport of heat northward across the hemispheres (*e.g.* Feulner et al., 2013; Buckley and Marshall, 2016), strong heat loss in the North Atlantic and Arctic, and active deep convection that sustains the strong AMOC. A weak AMOC, on the other hand, is associated with a weaker northward transport of salt and heat. This increases the stratification in the subpolar North Atlantic and thus favors the expansion of sea ice. The weak northward heat transport and the insulating effect of the sea ice keep the density gain due to heat loss small and the AMOC in a weak state (*e.g.* Klockmann et al., 2018). This weak AMOC state is stable when Antarctic Bottom Water becomes dense and salty enough to replace North Atlantic Deep Water (NADW) in the deep North Atlantic.

Sea ice can act as both a slow and fast positive feedback on AMOC-induced changes in climate. Extensive stadial sea ice cover during a weak AMOC state cools Greenland and suppresses atmosphere-ocean exchange of heat and oceanic convection in the North Atlantic (Li et al., 2005, 2010). This also leads to a slow build up of heat in the North Atlantic subsurface. Foraminifera from marine sediment cores offer evidence to back-up that this sub-surface warming occurred before the onset of fast D-O warming events (Rasmussen and Thomsen, 2004; Singh et al., 2014; Dokken et al., 2013). This heat build-up sets up the conditions for subsequent fast losses of GS sea ice.

Wind-driven, AMOC, and sea-ice linked salinity changes also play a crucial role in D-O positive and negative feedbacks. Indeed the net freshwater transport in the Atlantic basin by the AMOC can be used to assess the stability regime of the AMOC (Rahmstorf, 1995; Huisman et al., 2010). The interaction of subpolar and tropical salinity anomalies at the surface and in the subsurface (Jackson and Vellinga, 2013), and possible roles of the intertropical convergence zone and freshwater export through the Fram Strait, are also important in D-O related salinity feedbacks. Klockmann et al. (2018) note that if the subtropical gyre shifts northward and the sub polar gyre contracts: an inflow of salty subtropical water extends over the entire Atlantic basin east of the Mid-Atlantic Ridge. This inflow can supply salty water to the deep-convection sites in the Iceland Basin and Irminger Sea, and help maintain continuous deep convection and a strong AMOC even at low CO_2 concentrations (Brown and Galbraith, 2016; Klockmann et al., 2018; Guo et al., 2019; Muglia and Schmittner, 2015; Sherriff-Tadano et al., 2018), thus preventing the initiation of GS-like conditions. Where the AMOC does enter a weak state for a prolonged period, and the climate enters a GS, a build-up of heat in subsurface waters and salt in the tropical Atlantic can enable the very rapid resumption of the AMOC (Lynch-Stieglitz, 2017), with the upward mixing of heat from the subsurface and importation of salt from the tropic Atlantic via gyre mechanisms (Peltier and Vettoretti, 2014).

The importance of vertical (diapycnal) mixing in the ocean for these long timescale, D-O type, instabilities has long been recognised (Welander, 1982). However, we note that the different ocean- and climate-models (Table A1) parameterise diapycnal mixing in very different ways (*e.g.* Nilsson et al., 2003; de Lavergne et al., 2019). The lack of a single consistent paramaterisation and differences in the strength of diapycnal mixing across climate models means it is to be expected that

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

some models will produce D-O like oscillations “out of the box” under MIS3 boundary conditions, but others may require changes, or tuning, to their diapycnal mixing parameters. Even within the same model, a large range of diapycnal diffusivities may yield steady states that satisfy common plausibility constraints such as AMOC transport and sea ice distribution (see e.g. Holden et al., 2010). This is partly because wind-driven Southern Ocean upwelling plays a complementary role to diapycnal mixing in setting the steady state overturning (Samelson, 2004), and partly because surface buoyancy forcing controls the relative strength of the upper (Atlantic) and lower (Antarctic) overturning cells (Oliver and Edwards, 2008). A realistic AMOC transport may be obtained due to compensating biases in these processes, which has serious implications for whether AMOC feedbacks (necessary for capturing D-O behaviour) are represented in an adequate manner within these models.

Figure 2 and 3 show some of the key states, processes, and ocean sea-ice feedbacks that enable D-O events. Following Lohmann and Ditlevsen (2019), D-O events can be broken down into four periods: (1) cold stadial state (Fig. 2a), (2) rapid warming phase governed by very fast-time-scale mechanisms (Fig. 2b), (3) warm interstadial state (Fig. 2c and Fig. 3a) and, (4) gradual cooling phase (Fig. 3b) followed by a faster abrupt transition into a cold stadial phase (Fig. 3c). For some of the D-O events, the magnitude of the warming transitions are on the order of ten degrees in a decade, while the slow cooling in the interstadials is on the order of a few degrees in a millennium (the sawtooth shape) (Lohmann and Ditlevsen, 2019). This picture of rapid retreat of North Atlantic sea ice (Spolaor et al., 2016; Dokken et al., 2013) associated with the resumption of convection and the AMOC, alongside an upwards mixing of salt and heat, followed by a slower cooling phase back into stadial conditions matches accumulation, temperature, and water isotopes retrieved from Greenland ice core records of D-O warming events (Li et al., 2005, 2010; Sime et al., 2019).

2.2 The role of Northern Hemisphere Ice Sheets

Section 2 and Table A1 suggest that large Northern Hemisphere Ice Sheets and the wind regime associated with these can contribute to a strong AMOC which stabilises the North Atlantic and prevents D-O events. Thus ice sheets have a critical role to play in setting up the conditions for D-O events (Zhang et al., 2014; Klockmann et al., 2018; Brown and Galbraith, 2016; Muglia and Schmittner, 2015; Sherriff-Tadano et al., 2018). Figure 4 and 5 show some of the key mechanisms and feedbacks that are behind a state of reduced likelihood for D-O events and a potentially D-O type oscillating state, respectively.

The Northern Hemisphere Eurasian ice sheet was most probably limited to mountainous areas during mid-MIS3 (Helmens, 2014; Hughes et al., 2016), and its impact on D-O dynamics was probably relatively small. However, the size and presence (or absence) of the Laurentide ice sheet (LIS), which has elevations reaching a maximum of approximately 3000 m (Abe-Ouchi et al., 2015) at the LGM, does appear to cause important and robust (across multiple models) changes to Northern Hemisphere atmospheric circulation and resultant wind forcing of the ocean. LIS-dependent wind changes influence the subpolar gyre and the stability of the atmosphere-ice-ocean coupled system (Li and Born, 2019; Zhang et al., 2014).

A larger LIS (especially its height) causes stronger Northern Hemisphere winds (Li and Battisti, 2008; Pausata et al., 2011; Hofer et al., 2012; Ullman et al., 2014; Löfverström et al., 2014; Merz et al., 2015); an amplified stationary wave over North America (Manabe and Broccoli, 1985; Cook and Held, 1988); the North Atlantic glacial jet to be more stable due to differences

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

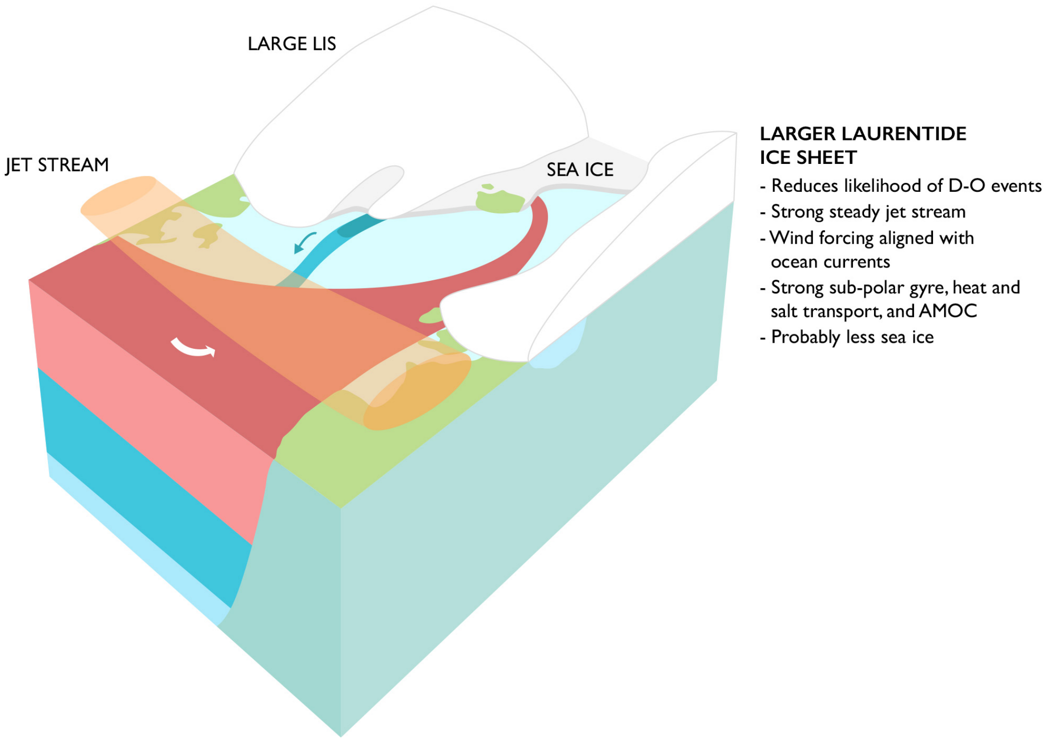

Figure 4. Schematic showing a state of reduced likelihood for D-O events.

in wave-mean flow feedbacks (Riviere et al., 2010); and alters variability of the large-scale atmospheric circulation, especially in the North Atlantic (Justino and Peltier, 2005; Pausata et al., 2009; Riviere et al., 2010). In addition, LIS height could control the sea-ice coverage and gyre circulation by shifting the westerlies over the North Atlantic region Zhang et al. (2014).

LIS altered winds that have wide implications for D-O relevant tipping elements (Seager and Battisti, 2007; Wunsch, 2006). Li and Born (2019) note that, first, the presence of a large LGM-type LIS is linked to a strong, more zonal and equatorward-shifted North Atlantic jet which weakens atmospheric heat transport into the North Atlantic (van der Schrier et al., 2010) and favours episodes of Greenland blocking (Madonna et al., 2017). Both could trigger the atmosphere-ice-ocean feedbacks that cause abrupt climate change in this area. Second, a steadier and stronger North Atlantic jet strengthens the wind-driven component of the subpolar gyre (Li and Born, 2019). Given that at latitudes north of about 45N, the subpolar gyre, which is essentially wind-driven, plays a crucial role in the northward transport of heat and salt, and is strongly linked to the AMOC (*e.g.* Jungclaus et al., 2013), wind-driven changes in this gyre have a strong impact on the density gain in the North Atlantic.

In many simulations with a large LIS (LGM-like ice sheets), the subtropical gyre can shift northward and cause an inflow of salty subtropical water over deep-convection sites, contributing to continuous deep convection and a strong AMOC even at low CO_2 concentrations (Brown and Galbraith, 2016; Klockmann et al., 2018; Guo et al., 2019; Muglia and Schmittner, 2015; Sherriff-Tadano et al., 2018; Zhang et al., 2014). Similarly, Zhang et al. (2014) note that a higher LIS can promote less South

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

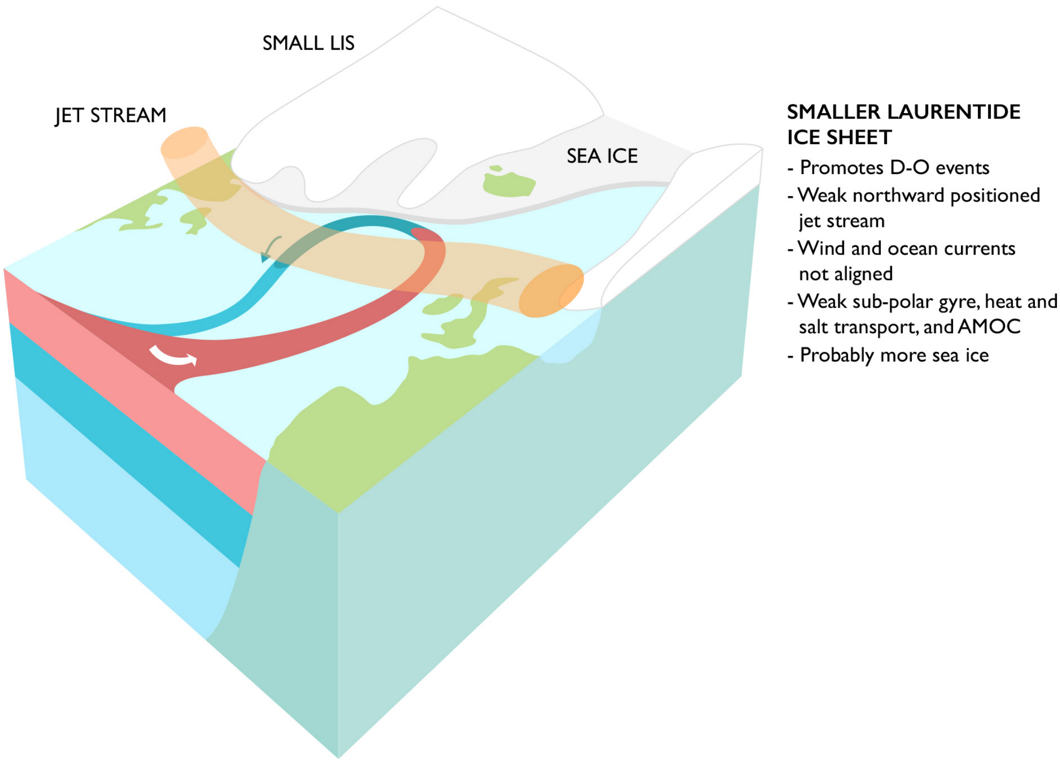

Figure 5. Schematic of a potentially D-O type oscillating state.

Labrador Sea sea ice export to northeastern North Atlantic (which reduces sea ice concentration) to permit deep convection and shift the core of westerlies northwards, strengthening subtropical gyre for heat and salt transport (Zhang et al., 2014).

For these reasons, large LGM-type ice sheets, particular a large LIS, tend to lead to a density gain over the North Atlantic and the northward salt transport is enhanced with respect to the PI ice sheet case. For many, but not all models, this tends to lead to more active convection in the North Atlantic and a strong AMOC (across a wide range of CO_2 concentrations). That said, the AMOC in many LGM simulations is likely too strong (Klockmann et al., 2018; Kageyama et al., 2021b). Thus the AMOC is far away from a tipping point with LGM-size ice sheets for many models (Zhang et al., 2014; Klockmann et al., 2018; Guo et al., 2019).

In some simulations with reduced ice sheets, the jet stream shifts northwards, leading to regional cooling and a rise in seasonal sea ice concentration over the subpolar gyre region (Armstrong et al., 2021). This freshens the area and lowers deep-water formation, which weakens the subpolar gyre and as a result the simulations are more prone to enter a weak convection, weak AMOC mode which is conducive to D-O type oscillations (Klockmann et al., 2018; Armstrong et al., 2021). Thus, with intermediate MIS3 LIS, *i.e.* reduced in its height compared to the LGM, multiple AMOC states are more likely (Zhang et al., 2014; Kawamura et al., 2017; Zhang and Prange, 2020; Armstrong et al., 2021; Klockmann et al., 2018).

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

3 Contours of a baseline MIS3 experiment protocol

Although the choice of a time within MIS3 for a D-O baseline experiment should be unimportant, given that in reality D-O events occurred during the whole of the MIS3, our analysis of existing simulations, boundary conditions and mechanisms above suggests that there are periods which may be particularly conducive to D-O events occurring in models. Furthermore, if this baseline experiment is to serve as a starting point for further sensitivity simulations, the boundary conditions in terms of CO_2, orbital forcing, and ice sheet forcing should be from a central value for MIS3. Oscillatory D-O type behaviour appears to be more likely, but not guaranteed (Guo et al., 2019; Zhang and Prange, 2020), when run with intermediate or low MIS3 CO_2 values and ice-sheets, *i.e.* reduced in size compared to the LGM (Brown and Galbraith, 2016; Kawamura et al., 2017; Klockmann et al., 2018; Zhang and Prange, 2020; Galbraith and de Lavergne, 2019; Zhang et al., 2014; Vettoretti et al., 2022), and particularly without a high LIS. The impact of orbital parameters has been investigated in less detail than the role of GHGs and ice sheets. Only a small number of studies have explored the potential importance of orbital configuration changes on triggering millennial-scale climate variability under intermediate glacial conditions (Rial and Yang, 2007; Mitsui and Crucifix, 2017; Zhang et al., 2021). Whilst Zhang et al. (2021) demonstrate that abrupt transitions from interstadial to statial states can be sensitive to obliquity-driven reduction in high-latitude mean annual insolation and/or precession-driven rise in low-latitude boreal summer insolation, the ubiquity of D-O events throughout MIS3 suggest that insolation forcing should not be a primary driver of D-O events.

These considerations suggest that the interval starting at 38 ka to 32 ka is a good choice for the proposed baseline experiment: it is characterised by (1) a rather regular sequence of D-O events (Fig. 1), (2) no evident changes in ice volume and atmospheric CO_2, and (3) has the ideal central-to-low GHG conditions and intermediate MIS3 ice-sheet configuration conducive to generating D-O-type quasi-oscillations (Section 2). In addition, Zhang et al. (2021) report, for the interval around 36-32 ka, unforced AMOC oscillations for a transient simulation performed with only varying orbital parameters from 40 to 32 ka.

A baseline simulation needs to be run for a sufficient duration to allow the strong positive feedbacks, together with long time-scale negative feedbacks, that enable D-O type oscillations. The analysis of existing simulations (Section 2) suggests this should be a minimum of 5000 years (Peltier and Vettoretti, 2014; Kleppin et al., 2015; Sidorenko et al., 2015; Brown and Galbraith, 2016; Klockmann et al., 2018, 2020). However, given computational constraints, a minimum duration of around 2000 years, with a spin-up period of 1000 years, may be a more practical minimum requirement for most modeling groups. It would, however, be important to examine and document key metrics for model drift (such as top-of-atmosphere radiation imbalance, deep ocean or global mean ocean temperature) during the initial spin-up. The exact length of spin-up is thus subject to discretion of each modelling group based on these key metrics.

There are two obvious possibilities for spinning up the MIS3 control experiment (MIS3-cnt). The baseline experiment could be initialised from either the end of a well spun-up LGM or PI experiment. Other possibilities could be to spin up from a linear combination of LGM and PI states (as done in Klockmann et al., 2016, 2018) or spinning up from present day's observations (as done in Guo et al., 2019). Modelling groups are encouraged to choose whichever option is more feasible/convenient for

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

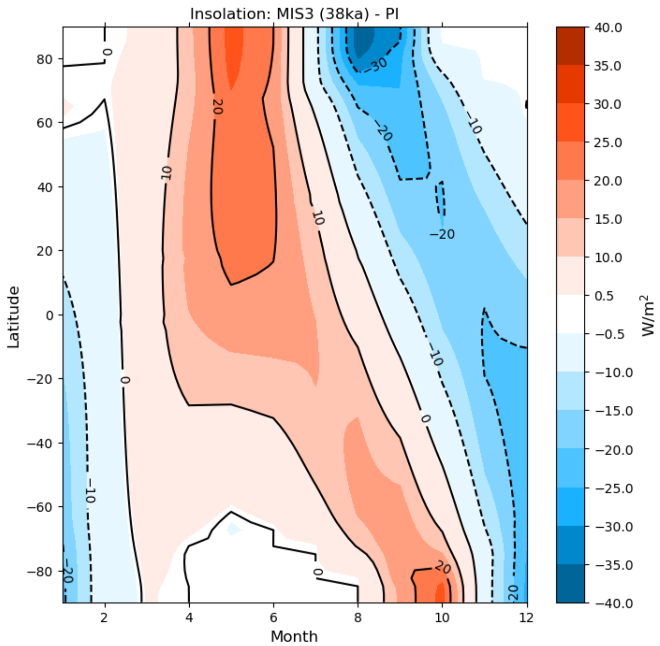

Figure 6. Monthly zonal-mean MIS3 (38ka) - PI anomalies of the top-of-atmosphere short-wave incoming radiation (W m^{-2}).

them. In case that several spin-up options are available, short spin-ups with diagnosed top-of-atmosphere (TOA) imbalance or global mean ocean temperature could help distinguish the faster spin-up option.

We suggest to perform a MIS3-cnt experiment using boundary conditions following Guo et al. (2019), i.e. GHG and orbital conditions for 38 ky (Fig. 6); and ice sheet configurations as outlined below.

3.1 Atmospheric trace gases

MIS3 atmospheric CO_2 values varied between a maximum of $\sim$ 233 ppm to a minimum of $\sim$ 187.5 ppm (Table 2; Figure 1; Bauska et al., 2021). Interestingly, increases of around 5 ppm happened during the abrupt warming of most D-O events and increases of up to 10 ppm happened within some Heinrich stadials (Bauska et al., 2021). GHG forcing is critical to model stability. Low (LGM-like) to intermediate (MIS3) CO_2 concentrations tend to be associated with abrupt D-O type AMOC transitions in models (Section 2 and Klockmann et al., 2018; Zhang et al., 2017, 2014; Brown and Galbraith, 2016; Vettoretti et al., 2022). We thus suggest to perform the MIS3-cnt experiment using the GHGs values specified in Table 2 and keep these values fixed for the whole duration of the simulation including the spin-up.

3.2 Northern Hemisphere Ice Sheets

Constraining MIS3 ice-sheet boundary conditions is a challenge. Scarcity and fragmentation of evidence (Kleman et al., 2010; Batchelor et al., 2019) is an issue. In particular, it is difficult to determine the size and shape of the ice sheets during MIS3

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

because subsequent larger LGM configurations have overridden and destroyed evidence of the position of the margins of these smaller ice sheets.

Global sea level fluctuations during the mid-MIS3 were driven nearly exclusively by the LIS (Gowan et al., 2021). Global average sea level remained above -55 m for the period between 30-55 ka. From glacial isostatic modelling and geological constraints, a global mean sea level between -30 m and -50 m is inferred (Dalton et al., 2022). For much of MIS3, since the Eurasian ice sheets and the Cordilleran Ice Sheet were likely restricted to mountain-based caps (Helmens, 2014; Hughes et al., 2016), the primary control on ice volume is assumed to be from the LIS. Recent work in the area of the Hudson Bay (Dalton et al., 2016, 2019; McMartin et al., 2019; Dalton et al., 2022) suggests ice-free conditions may have occurred during mid-MIS3. This implies climatic conditions in this region similar to present (Dalton et al., 2017), and a LIS margin removed from the southern Hudson Bay. Similarly Tarasov et al. (2012) show a considerably lower and less extensive LIS compared to ICE-5G and ICE-6G (Peltier, 2004; Peltier et al., 2015) LGM ice sheet reconstructions. Pico et al. (2017) sea-level curves are consistent with the estimated MIS3 ice-sheet volumes from Batchelor et al. (2019). Using Glacio Isostatic Adjustment modeling, Pico et al. (2017) also show that a small LIS can explain high MIS3 sea-level estimates alongside the eastern coast of the United States. Batchelor et al. (2019)'s synthesis of numerical modelling results and empirical data provides additional support for a considerable reduction in the MIS3 LIS extent and very minimal European ice sheet.

The recent MIS3 ice sheet reconstruction, PaleoMIST 1.0 (Paleo Margins, Ice Sheets, and Topography), was developed independently of far-field sea-level records and indirect proxy records by Gowan et al. (2021). This reconstruction is based on trying to fit the evolution of ice flow indicators, as well as chronological constraints of ice-free conditions.

Gowan et al. (2021) provide a maximum and minimal MIS3 reconstruction. However the maximum scenario is the more consistent with recently discovered eastward oriented, pre-LGM ice flow direction indicators found in southeastern Manitoba (Gauthier and Hodder, 2020). The 37.5 ka time slice is representative of conditions prior to Heinrich Event 4 (Andrews and Voelker, 2018), and therefore the ice thickness in Hudson Bay may be somewhat larger than ideal for the post H4 D-O events. The ice margin elsewhere for the Laurentide Ice Sheet is based on chronological constraints, most that are documented in the compilation by Dalton et al. (2019). The Cordillera Ice Sheet extent is based on evidence of relatively restricted ice cover during MIS 3 (Clague and Ward, 2011). The Greenland Ice Sheet margin is set to be intermediate of the LGM and present day extent. The Eurasian ice cover is taken to be intermediate of the DATED-1 minimum and maximum extent margins for their 35-38 ka time slice (Hughes et al., 2016). For East Antarctica, the margin is set to be the same as present, while West Antarctica, the margin is between present day and LGM extent.

Given its strong evidence basis, we thus suggest the use of the maximum 37.5 ka Gowan et al. (2021) PaleoMIST ice sheet configuration. We note the LIS is considerably reduced in size, compared to the ICE-6G LGM reconstruction in the southeastern margin (Fig. 7a,d); the EIS is also significantly smaller (Fig. 7a,d).

An alternative data-constrained ice sheet model for 38 ka is shown in Fig. 7c,f. This consists of North American (Tarasov et al., 2012), Eurasian (Lev Tarasov, personal communication), Greenland (Tarasov and Richard Peltier, 2002), and Antarctic (Briggs et al., 2014) ice sheets. This particular ice sheet configuration has been previously used by Guo et al. (2019). This reconstruction is also smaller than ICE-6G (Fig. 7a,d), with a southeastern LIS margin further north. Similarly it has a smaller

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

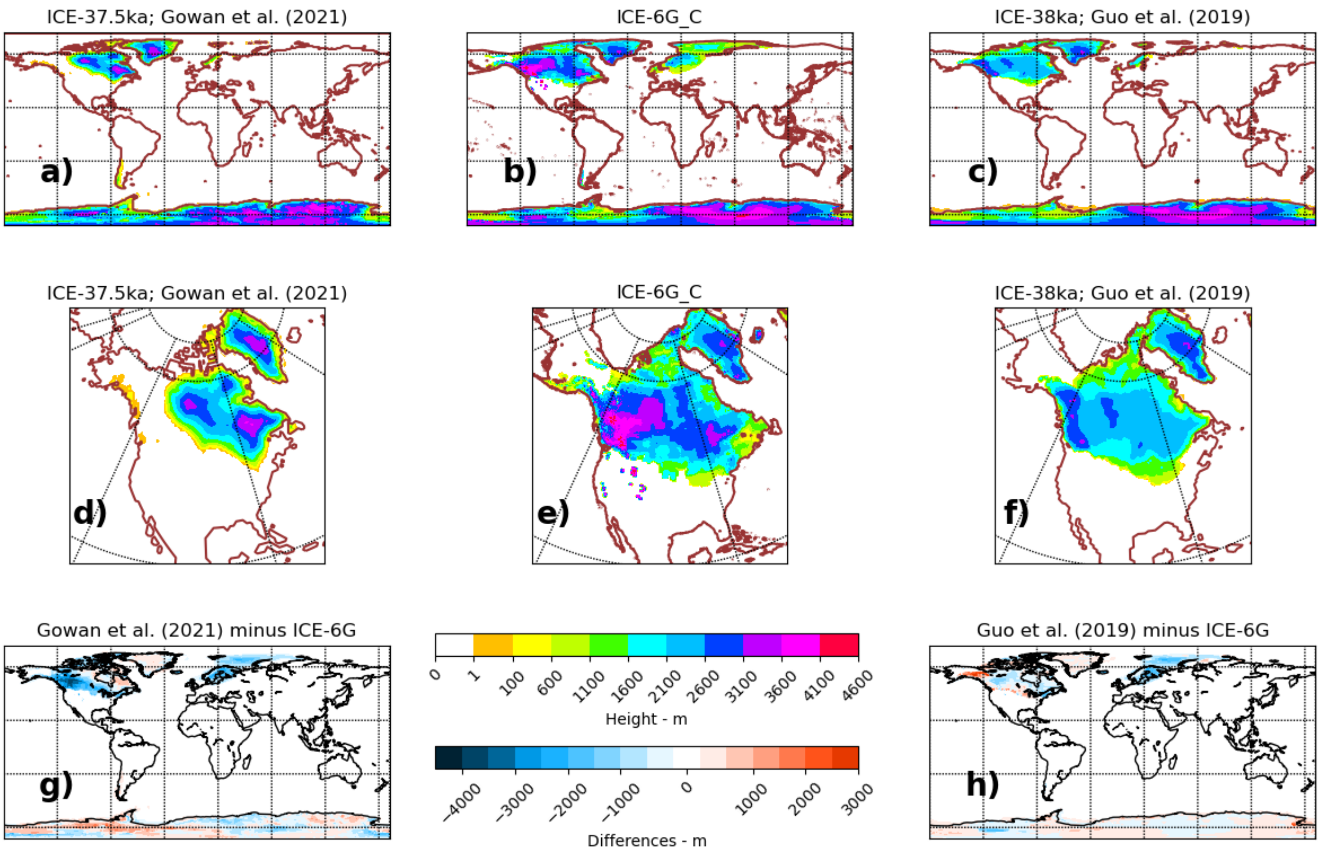

Figure 7. MIS3 ice sheet reconstruction from (a,d) Gowan et al. (2021) and (c,f) Guo et al. (2019). Also shown for comparison is the (b,e) LGM ICE-6G ice sheet reconstruction from Peltier et al. (2015). Third row shows the differences between the MIS3 ice sheet reconstruction and the LGM ICE-6G reconstruction.

EIS, although Fennoscandia is covered by land ice (Fig. 7c,f). The Cordilleran Ice Sheet is merged with the LIS and the Barents Sea is kept free of land ice. There is a significant amount of land ice over the Canadian Archipelago, blocking the transport of water between Baffin Bay and the Arctic (Fig. 7c). In Antarctica, grounded ice cover the Weddell and Ross rather than floating ice shelves as present-day (Fig. 7c).

Whilst the implementation of the ice sheets will differ between models, the steps of Kageyama et al. (2017) describe how to implement a glacial state ice sheet in the IPSL climate model. For consistency, we likewise recommend the same steps should be followed as far as possible. Since a reduced sea-level can modify river courses, Kageyama et al. (2017) recommend that as a minimum, rivers should reach the oceans. Also, the ocean should be initialized with a salinity 0.6 psu higher than the PI experiment, to account for the sea-level difference between MIS3 and PI experiment (freshwater stored as ice on land) (Guo et al., 2019).

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

3.3 An additional kicked Heinrich Event preconditioned option

The MIS3 D-O events are mostly grouped in Bond cycles (Bond et al., 1993; Lehman, 1993; Bond and Lotti, 1995), where cycles begin and finish with a H-event. The freshwater delivered during Heinrich event iceberg discharge extends GS duration and suppresses the AMOC, leading to accumulation of heat in the Southern Hemisphere, and in the North Atlantic subsurface waters (*e.g.* Stocker and Johnsen, 2003). Estimates of the meltwater input into the North Atlantic during Heinrich events range between 2 m and 15 m of sea level equivalent ice volume (Hemming, 2004; Chappell, 2002; Rohling et al., 2004; Roche et al., 2004; Roberts et al., 2014b; Siddall et al., 2008; Grant et al., 2014). It is logical to presume that these freshwater events are important in preconditioning the climate system with respect to D-O behaviour and indeed Peltier et al. (2020) have suggested that some models require an Heinrich event as a precursor to the D-O type quasi-oscillatory behaviour. Freshwater perturbations can trigger changes between AMOC states (*e.g.* Ganopolski and Rahmstorf, 2001; Timmermann et al., 2003; Stouffer et al., 2006; Hu et al., 2008; Kageyama et al., 2012; Roberts et al., 2014a; Sime et al., 2019) and a relatively small freshwater flux applied over convection areas can lead to a shutdown of the AMOC (*e.g.* Roche et al., 2010). Studies have shown similarities between observed global features of abrupt D-O changes and the behaviour seen in freshwater forcing experiments (Liu et al., 2009; Menviel et al., 2014). However, the sensitivity of the AMOC to a wide range of freshwater inputs varies according to model, where the meltwater is added, and the background climate state (Ganopolski and Rahmstorf, 2001; Timmermann et al., 2003; Stouffer et al., 2006; Hu et al., 2008; Kageyama et al., 2012; Roberts et al., 2014a; Zhang et al., 2014). Given these various uncertainties, we suggest that it would be useful to run an additional experiment to investigate how preconditioning through a kicked H-event impacts the simulation of D-O like oscillations under MIS3 boundary conditions.

The range of H-event volumes calculated using ice sheet models varies from 24.2 to 125 x 10^4km^3 (MacAyeal, 1993; Dowdeswell et al., 1995; Marshall and Clarke, 1997; Hulbe, 1997), whilst isotope based estimates and a precipitation balance approach have yielded 86, 649, and 946 x 10^4km^3 of ice volume (Hemming, 2004; Roche et al., 2004; Levine and Bigg, 2008). Roberts et al. (2014b), and using a sediment modelling approach estimated a discharge of 30 to 120 x 10^4km^3 of ice volume. Some of the spread in these estimates could be because the relationship between the oxygen isotope record, sea level, and meltwater volume is not constant when ice is lost from marine basins, such that the use of oxygen isotopes for calculating H-event volumes may produce unrealistically high values (Gasson et al., 2016; Hemming, 2004; Roberts et al., 2014b). There is also some uncertainty about the duration of the H-events, with some previous studies suggesting they could be as short as 250 years (Hemming, 2004) and others suggesting a duration of 500 yr is more typical (Roberts et al., 2014b). These considerations suggest that it is possible to justify the use of anywhere between 0.02 - 0.6 Sv freshwater flux over 500 years; or 0.04 - 1.2 Sv over 250 years. More recent estimates of H-event magnitudes tend to favour the lower end of this range. If all forcings are set to MIS3-cnt values and the H-event freshwater flux is distributed across the North Atlantic this could yield a range of stadial climates (Sime et al., 2019; Zhang and Prange, 2020). A subsequent switch-off of the freshwater forcing after this 250-500 year H-kick would be a useful addition to the MIS3-cnt.

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table 2. Summary of the boundary conditions (BC) and forcings for the MIS3-cnt experiment.

BC/Forcing	Suggested value MIS3-cnt
Atmospheric trace gases	CO_2: 219 ppm (Bauska et al., 2021) CH_4: 526 ppb (Loulergue et al., 2008) N_2O: 250 ppb (Schilt et al., 2010)
Insolation	Eccentricity: 0.013676 Berger (1978) Obliquity: 23.268° Berger (1978) Perihelion - 180°: 205.94° Berger (1978)
Solar constant	Same as PI control
Ice sheets	38ka ice sheet reconstruction (Guo et al., 2019) 37.5ka ice sheet reconstruction (Gowan et al., 2021); mean global salinity increased by 0.6PSU to account for ice volume
Global freshwater budget	Closed to avoid drifts; Snow should not accumulate over ice sheets and rivers should flow into the ocean. Models need to consider lakes when closing the global freshwater budget
Vegetation	Dynamic or fixed as in PI. If fixed vegetation: tundra in land new points
Dust	As in PI control
H-kicked variant	initial 0.04 - 1 Sv over 250-500 years followed by standard MIS3-cnt simulation

4 Conclusions

D-O events are abrupt, large climate changes that punctuated the last glacial period. There is uncertainty whether current IPCC-class models can effectively represent the processes that cause D-O events. We have shown that reduced ice sheets relative to LGM and low-to-medium MIS3 CO_2 values are more likely to lead to unforced quasi-oscillatory D-O type behaviour. However, the simulations need to be run long enough to allow the strong positive AMOC feedbacks, along with negative feedbacks on long time-scales, which can then lead to D-O type oscillations. Around 42% of the simulations set-up with full LGM or more MIS3-like conditions, have a run length of less than 2000 model years, which makes it difficult to tell whether any of these simulations are capable of, or likely to, exhibit D-O like behaviour. In addition, the vast majority of PMIP4/CMIP6 models have not run LGM or MIS3-like simulations long enough to be sure which models have the capability to oscillate.

We have provided boundary conditions for a baseline MIS3-cnt simulation, and a H-event-like preconditioned variant (freshwater forced experiment). The MIS3-cnt experiment covers the interval from 38 to 32 ka because: (1) it features a rather regular sequence of D-O events, (2), it is characterised by no evident changes in ice volume and atmospheric CO_2, and (3) it yields the

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

ideal combination of intermediate ice sheets (smaller in size compared to LGM) and medium-to-low GHG values conducive to oscillatory D-O type behaviour in models. Ideally, the MIS3 baseline experiment should be run for 5000 years, however, given computational constraints a minimum duration of 2000 years together with a spin-up of at least 1000 years is a more practical minimum requirement. This baseline MIS3-cnt protocol provides a common framework to explore cold-period instabilities using particular GHG-, insolation-, freshwater-, and NH ice sheet-related forcings, together with diapycnal mixing. These simulations will allow us to answer questions such as: Is there a difference between how different classes of climate models represent D-O like behaviour? How important are atmospheric dynamics, or are ocean-sea ice interactions dominant? What controls the time-scales and amplitudes of the oscillations? And finally, are climate models too stable?

Team list. Ayako Abe-Ouchi; Andreas Born; Nathaelle Bouttes; Peter Ditlevsen; Michael P. Erb; Georg Feulner; Evan J. Gowan; Lauren Gregoire; Chuncheng Guo; Sandy P. Harrison; Heather Andres; Masa Kageyama; Marlene Klockmann; Fabrice Lambert; Allegra N. LeGrande; Ute Merkel; Larissa S. Nazarenko; Kerim H. Nisancioglu; Kevin Oliver; Bette Otto-Bliesner; William R. Peltier; Matthias Prange; Kira Rehfeld; Alexander Robinson; Lev Tarasov, Paul. J Valdes; Guido Vettoretti; Nils Weitzel; Qiong Zhang; Xu Zhang

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Competing interests. No competing interests are present

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Preprint. Discussion started: 6 September 2022

Appendix A

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A1: Models that exhibit spontaneous oscillations.

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings	Mechanisms for D-O type cooling event	Mechanisms for D-O type warming event	Similar Mechanism to schematic in Fig.2 or 3
Drijfhout et al. (2013)	EC-Earth model	PI	PI	PI	PI	None	1125 yrs	Spontaneous cold event that last around 100 years	An anomalous high atmospheric blocking over the eastern subpolar gyre causes the cold event. Ocean currents transport sea ice southwards and there is a shut-down of deep water convection in the Labrador Sea.	No warming event reported.	No. Atmospheric blocking–sea-ice–ocean feedback identified as a main cause behind the cold event.
Kleppin et al. (2015)	CCSM4	PI	PI	PI	PI	None	1000 yrs	D-O type oscillations in Greenland temperature are shown.	The cooling event has a duration of 200 years and is linked to a weak-ened state of the SPG and deep wa-ter convection in the Labrador Sea.	The warming event is triggered by a stronger Icelandic low and therefore deep water convec-tion recovers and SPG circulation resumes.	No. Stochastic atmo-spheric forcing identified as a potential cause for sea ice variations.
Sidorenko et al. (2015)	ECHAM6-FESOM	Present day (PD)	PD - 1990	PD	PD - 1990	None	350 yrs	Events of sudden re-duction of deep water convection and increase of sea ice cover in the Labrador Sea.	An anomalous in-flow of warm and saline water into the deep Labrador Sea causes a weaken-ing of the subpolar gyre and modifies the upper freshwa-ter budget over the Labrador Sea.	No warming event reported.	No. The strong surface winds over the subtropi-cal NA alter the Gibraltar Strait outflow path into the NA and are identi-fied as the main cause be-hind the saline and warm anomalies in the deep NA.

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https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A1 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings	Mechanisms for D-O type cooling event	Mechanisms for D-O type warming event	Similar Mechanism to schematic in Fig. 2 or 3
Martin et al. (2015)	KCM	PD	PD	PD	PD	None	1300 yrs	Centennial-scale variability of the AMOC as well as variations in the NA heat content and subpolar gyre strength.	As the NA Current accelerates, deep convection in the Weddell Sea enables a positive heat content anomaly to propagate northwards in the upper Atlantic Ocean. Eventually, the heat anomaly reaches the northern NA resulting in a reduction in deep water formation there.	The retreat of Antarctic Bottom Water (AABW) leads to enhanced meridional density gradient that results in an increased North Atlantic Deep Water (NADW) cell.	No. Interhemispheric teleconnection; Variability in the Southern Ocean deep water convection identified as the main cause behind AMOC oscillations.
Peltier and Vettoretti (2014); Vettoretti and Peltier (2016, 2018)	UofT CCSM4	LGM	LGM – 21ka	LGM - ICE-6G (VM5a)	LGM – 21ka	None	5000 yrs	Regular cycles of D-O type oscillations	The continuous flux of sea ice from the Arctic basin into the NA subpolar gyre area across the East Greenland Current, favours melting of sea ice as it moves over the warm ocean surface. This freshwater input restratifies the high-latitude NA and results in a considerable decrease in the rate of NA Deep Water formation.	The initiation of the abrupt warming events is associated with the opening of a large polynya over the Irminger Sea. The stability of the water column is key and depends on transport of salt to the subpolar gyre along the Irminger Current and Denmark Strait in the decades preceding the warming event.	Yes

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https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A1 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings	Mechanisms for D-O type cooling event	Mechanisms for D-O type warming event	Similar Mechanism to schematic in Fig. 2 or 3
Vettoretti et al. (2022)	CCSM4	Glacial simulations	CO_2 levels from 190 to 225 ppm	LGM - ICE-6G (VM5a)	LGM	None	8000-10000 yrs	Regular cycles of D-O type oscillations within a window of CO_2 levels from 190 to 225 ppm	Old sea ice from the Arctic is exported to the NA, sea-ice growth is favoured through ice-albedo feedback, high-latitude convection is reduced through sea-ice melt, and consequently Antarctica and Greenland cool. The interstadial-to-stadial transition happens with fast NA sea-ice expansion and NADW production collapse.	During a NA stadial, sea ice thins in the Southern Ocean and Antarctica warms. There are increases in salt convergence in the NA, NADW fluctuations are amplified via salt advection feedback, and the volume of NADW increases, allowed by late-stadial decreases in AABW formation. Late-stadial NA vertical stratification experiences thermohaline instability and the Nordic and Irminger Seas are destabilised, triggering rapid sea-ice loss in the NA and the transition from stadial to interstadial states.	Yes

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https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A1 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings	Mechanisms for D-O type cooling event	Mechanisms for D-O type warming event	Similar Mechanism to schematic in Fig. 2 or 3
Armstrong et al. (2021)	HadCM3B	MIS3 - 30ka	30ka	30ka	30ka	None	6000 yrs	D-O type stochastic oscillations on a 1500-year timescale	Ocean forcing initiates the stadial phase; The collapse of the salinity gradient between the Northern NA and STG leads to a reduced advection in the Nordic Seas and decreased deep-water formation	The initiation of the interstadial phase is associated with a wind-driven atmospheric forcing in the Nordic Seas due to increased regional temperatures, reduced sea ice cover and increased sea level pressure, which enhances wind stress and convection	Yes, partially. The D-O type oscillations reflect a salt oscillator mechanism in the NA.

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https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A1 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings	Mechanisms for D-O type cooling event	Mechanisms for D-O type warming event	Similar Mechanism to schematic in Fig. 2 or 3
Zhang et al. (2021)	COSMOS	MIS3: 40-32ka	40ka	40ka	One transient simulation of 40-32ka and one 40ka snapshot simulation (with 34 ka orbital parameters)	None	+5000 yrs	D-O type oscillations on a 1200-year timescale; Orbitally induced AMOC changes	Transitions from warm interstadial to cold stadial are linked to (1) a precession-controlled rise in low-latitude boreal summer insolation by modifying the NA low-latitude hydroclimate and/or (2) an obliquity-controlled reduction in high-latitude annual insolation by altering high-latitude sea ice–ocean–atmosphere interactions.	While the AMOC is in its weak phase, a gradual increase in subsurface temperature in the subpolar ocean together with enhanced northward transport of salt in the NA, drive the AMOC back to its strong phase.	Some similarities: unforced AMOC oscillations are triggered by either the tropical salt impact (linked to preccesion-controlled summer insolation) and/or the subpolar thermal impact (linked to obliquity-controlled mean annual insolation).
Brown and Galbraith (2016)	CM2Mc	Mixed forcing	CO_2 = 180 ppm	PI	Obliquity: 22°; Precession: 90°	None	more than 8000 yrs	Unforced AMOC oscillations	During a weak AMOC phase, NA deep convection is largely reduced and there is an expansion of sea ice in the northeast Atlantic. Heat accumulates at depth in the NA linked to the weak advection of warm waters from the tropics.	During a strong AMOC phase, NA deep convection is intense and there is a retreat of sea ice in the northeast Atlantic	Yes, partially. Salt advection is a key driver of the oscillations, specifically the salt exchange between subpolar and subtropical NA.

Continued on next page

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A1 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings	Mechanisms for D-O type cooling event	Mechanisms for D-O type warming event	Similar Mechanism to schematic in Fig. 2 or 3
Klockmann et al. (2018, 2020)	MPI-ESM	Mixed forcing	CO_2 = 206 ppm; CH_4 = 444 ppb; N_2O = 218 ppb	PI	LGM – 21 ka	None	12000 yrs	Spontaneous millennial-scale AMOC oscillations	Stadial phases correspond to weak AMOC and strong SPG phases. The extensive SPG results in low northward salt transport and deep convection only occurs sporadically in the Iceland basin. The Nordic Seas are entirely ice covered which results in a weak Icelandic Low and therefore in a weak wind stress curl. Subsurface waters in the Nordic Seas are around 3 K warmer than during interstadial phases.	During interstadial phases, the AMOC is strong and the SPG is contracted and weak. There is a broad inflow of salty subtropical water to the subpolar NA. Changes in the SPG are driven by variations in the cross-gyre density difference. The eastern NA is fully ice-free. Deep convection occurs continuously in the Iceland basin, Irminger Sea and the Nordic Seas.	Yes, partially. The proposed mechanism behind the spontaneous AMOC oscillations compromises three components: (1) oscillations in salinity comparable to Peltier and Vettoretti (2014), (2) a density-driven feedback loop comparable to Montoya et al. (2011), and (3) a wind-driven feedback loop comparable to Drijfhout et al. (2013) and Kleppin et al. (2015).
			CO_2 = 195 ppm; CH_4 = 396 ppb; N_2O = 209 ppb				8000 yrs				

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

A1

Table A2: List of simulations run under MIS3/mid-glacial conditions.

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings
Guo et al. (2019)	NorESM1-F	MIS3 – 38 ka	CO_2 = 215 ppm; CH_4 = 550 ppb; N_2O = 260 ppb	Data-constrained 38 ka	38 ka	None	2500 yrs (recently extended to 6000 model years)	The equilibrium MIS3 simulation does not show spontaneous D-O type oscillations. Attempts at perturbing the system into a cold stadial state, by modifying the height of the LIS and atmospheric CO2 levels, show that the modelled MIS3 interstadial state is rather stable, and thus questioning the occurrence of spontaneous D-O type oscillations in the lack of interactive ice sheet-meltwater dynamics.
Zhang and Prange (2020)	CCSM3	MIS3 – 38 ka	CO_2 = 215 ppm; CH_4 = 501 ppb; N_2O = 234 ppb	ICE-5G ice sheet configuration (Peltier, 2004).	38 ka	12 hosing/extraction experiments with freshwater fluxes from ±0.005 Sv to ± 0.2 Sv, injected in the Nordic Seas for 500 years.	2170 yrs	AMOC is more sensitive to meltwater fluxes under MIS3 conditions than under LGM conditions. The lower AMOC stability under MIS3 conditions proposes that D-O type oscillations could have been triggered by small perturbations in the ocean surface meltwater forcing e.g. linked to ice-sheet processes.

Continued on next page

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A2 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings
Kawamura et al. (2017)	MIROC 4m	Mid-glacial state	CO_2 = 215 ppm; CH_4 = 350 ppb; N_2O = 200 ppb	Intermediate-size ice-sheet configuration (at 15 ka)	15 ka	Hosing experiments with freshwater fluxes of 0.05 Sv and 0.1 Sv, injected in the North Atlantic Ocean (50°N to 70°N) for 500 years.	More than 2000 yrs	The climate response to freshwater perturbations is much lower under LGM conditions than under MIS3 conditions. The unperturbed LGM AMOC is unusually weak (around 6 Sv) and thus could barely be further lessened, such that meltwater hosing does not largely affect the large-scale climate.
Vettoretti et al. (2022)	CCSM4	Glacial run	CO_2 = 210 ppm	LGM - ICE-6G (VM5a)	LGM	Hosing experiment (H event-like pulse) with freshwater fluxes of 0.05 Sv for 500 years in two separate stadial periods in a glacial simualtion run with CO_2 of 210 ppm. The freshwater flux is injected in the North Atlantic (50°N to 70°N).	8000 yrs	The Heinrich simulation has a large Northern Hemisphere temperature and AMOC overshoot after the Heinrich stadial ends. Nevertheless, this fast AMOC rise above regular interstadial levels is in agreement with observations only for a few H-stadial periods (H4 and H5).

Continued on next page

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A2 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings
Zhang et al. (2014, 2017)	COSMOS (ECHAM5-JSBACH-MPIOM)	Mixed forcing	LGM	Sensitivity experiments applying different heights of the Northern Hemisphere ice sheets	LGM	None	Snapshot and transient simulations (no longer than 250 yrs)	An AMOC bi-stability regime is found under intermediate CO_2 and ice sheet conditions roughly resembling that of the MIS3 climate. In the bi-stable MIS3 regime, transitions from weak to strong AMOC state and vice versa could be initiated by not only gradual variations in LIS height and atmospheric CO_2 but also freshwater perturbations.
Zhang et al. (2014, 2017)	COSMOS (ECHAM5-JSBACH-MPIOM)	Mixed forcing	LGM	Moderate ice volume, 0.25-0.45 times the LGM	LGM	Hosing experiments with freshwater fluxes of ±0.02 Sv, injected in the North Atlantic Ocean (50–65° N, 5–30° W) for 100-300 years	no longer than 700 yrs	Changes in the LIS height can initiate a positive atmosphere-ocean-sea ice feedback leading to D-O type climate shifts. A gradual increase in the ice sheet height results in a northward shift of the winds, and favours a more saline Labrador Sea both by reducing the sea ice/freshwater import from the Arctic and increasing the advection of salt into the area.

Continued on next page

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A2 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings
Zhang et al. (2014, 2017)	COSMOS (ECHAM5-JSBACH-MPIOM)	Mixed forcing	CO_2 levels increased from 185 to 205 ppm during 500 years.	40% of LGM ice-sheet configuration	LGM	None	500 yrs	For a moderate ice volume, 0.25-0.45 times the LGM, two stable AMOC modes are identified. A variation of 15 ppm in atmospheric CO_2 concentration – equivalent to changes during D-O cycles containing HE – is enough to trigger oscillations between a weak stadial state to a strong interstadial circulation state.
Zhang et al. (2014, 2017)	COSMOS (ECHAM5-JSBACH-MPIOM)	Mixed forcing	CO_2 levels increased from 185 to 239 ppm at a rate of 0.02 ppm per year.	Intermediate ice sheet configuration - ice volume equivalent to approximately a sea-level drop of 40 m	LGM	None	600 yrs	
Zhang et al. (2014, 2017)	COSMOS (ECHAM5-JSBACH-MPIOM)	Mixed forcing	CO_2 levels increased from 185 ppm and 245 ppm at a rate of 0.05 ppm per year.	LGM ice volume	LGM	Persistent freshwater flux of 0.15 Sv	600 yrs	

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

A2

Table A3: Possible oscillatory behaviour for other time periods

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings
Brugger et al. (2019)	CLIMBER 3α	415 Ma and 380 Ma	CO_2 = 1500 ppm. Few runs with 500 ppm, 800 ppm, 2000 ppm	None	S0=1315 W/m^2 (415 Ma), 1319 W/m^2 (380 Ma), various obliquity values, eccentricities, precession angles	None	5000 yrs	Decadal to centennial temperature fluctuations at high northern latitudes.
Rind et al. (2018)	GISS E2-R (TCADI)	Future warming simulation (experiment "abrupt 4x CO2" in Rind et al. (2018)	4 x PI CO_2	PD	PD	None	2500 yrs	Multi-centennial cessation with restoration and rapid overshooting in NADW formation

Continued on next page

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

Table A3 – continued from previous page

Study	Model	Period	GHG	Ice sheet	Insolation	FWF	Run Length	Main findings
	GISS E2-R (TCADI)	Future warming simulation (experiment "1 pct CO2" in Rind et al. (2018)	CO_2 rises at 1% per year until 4 x CO_2 (after 140 years) and then held constant.				2500 yrs	
	GISS E2-R (TCADI)	Future warming simulation (experiment "rcp85" in Rind et al. (2018)	rcp85 emissions				4300 yrs	

https://doi.org/10.5194/egusphere-2022-707
Preprint. Discussion started: 6 September 2022

790 **Appendix B**

Table B1. Summary of LGM/MIS3-like simulations discussed in the text

Study	Model	Period	N° of simulations	Run length
Peltier and Vettoretti (2014)	UofT CCSM4	PMIP4 LGM	1	5000
Lohmann et al. (2020)	AWI-ESM1-1-LR	PMIP4 LGM	1	1300
Sidorenko et al. (2019)	AWI-ESM-2-1-LR	PMIP4 LGM	1	600
Tierney et al. (2020)	CESM1.2	PMIP4 LGM	1	1800
Valdes et al. (2017)	HadCM3B-M2.1aD	PMIP4 LGM	3	400-2900[a]
Lhardy et al. (2021)	iLOVECLIM1.1.4	PMIP4 LGM	2	5000
Volodin et al. (2018)	INM-CM4-8	PMIP4 LGM	1	50
Sepulchre et al. (2020)	IPSLCM5A2	PMIP4 LGM	1	1200
Ohgaito et al. (2021)	MIROC-ES2L	PMIP4 LGM	1	8960
Mauritsen et al. (2019)	MPI-ESM1.2	PMIP4 LGM	1	3850
Armstrong et al. (2021)	HadCM3B-M2.1aD	MIS3 (30 ka)	1	6000
Zhang et al. (2021)	COSMOS	MIS3 (40-32 ka)	2	5000
Guo et al. (2019)	NorESM	MIS3 (38 ka)	1	+6000
Zhang and Prange (2020)	CCSM3	MIS3 (38 ka)	1	2170
Kawamura et al. (2017)	MIROC 4m	Mid-glacial conditions	1	+2000
Vettoretti et al. (2022)	CCSM4	Glacial conditions	4[b]	8000
Brown and Galbraith (2016)	CM2Mc	Mixed forcing	1	+8000
Klockmann et al. (2018)	MPI-ESM	Mixed forcing	3	+8000
Zhang et al. (2014)	COSMOS	Mixed forcing	11[c]	300-4000 [d]

[a] Only one simulation run longer than 2000 model years

[b] Four simulations run with CO_2 levels: 200, 210, 220, 225 ppm

[c] We do not consider FWF runs nor transient simulations forced with varying CO_2 and/or NH ice sheet height

[d] Only two simulations with a duration of +2000 model years

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Appendix C

Table C1. Contributing members to PMIP4/CMIP6 that have run simulations under LGM or MIS3 conditions.

Model	Period	Run length
ACCESS-ESM1-5	-	-
AWI-ESM-1-1-LR	LGM	1300
CESM2	-	-
CNRM-CM6-1	-	-
EC-Earth3-LR	-	-
FGOALS-f3-L	-	-
FGOALS-g3	-	-
GISS-E2-1-G	-	-
HadGEM3-GC31-LL	-	-
INM-CM4-8	LGM	50
IPSL-CM6A-LR	-	-
MIROC-ES2L	LGM	+8000
MPI-ESM1-2	LGM	3850
MRI-ESM2-0	-	-
NESM3	-	-
NorESM1-F	MIS3	6000
NorESM2-LM	-	-