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Transient climate simulations of the deglaciation 21–9 thousand years before present; PMIP4 Core experiment design and boundary conditions

R. F. Ivanovic[1], L. J. Gregoire[1], M. Kageyama[2], D. M. Roche[2,3], P. J. Valdes[4], A. Burke[5], R. Drummond[6], W. R. Peltier[6], and L. Tarasov[7]

[1]School of Earth & Environment, University of Leeds, Woodhouse Lane, Leeds, LS2 9JT, UK
[2]LSCE/IPSL – Laboratoire des Sciences du Climat et de l'Environnement, UMR8212, CEA-CNRS-UVSQ – CE Saclay, L'Orme des Merisiers, Bat. 701, 91191 Gif-sur-Yvette CEDEX, France
[3]Earth and Climate Cluster, Faculty of Earth and Life Sciences, Vrije Universiteit Amsterdam, Amsterdam, the Netherlands
[4]School of Geographical Sciences, University of Bristol, University Road, Bristol, BS8 1SS, UK
[5]Department of Earth and Environmental Sciences, Irvine Building, University of St. Andrews, St. Andrews, KY16 9AL, UK
[6]Department of Physics, University of Toronto, 60 St George Street, Toronto, Ontario, M5S 1A7, Canada

[7]Department of Physics and Physical Oceanography, Memorial University of Newfoundland and Labrador, St. John's, NL, A1B 3X7, Canada

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Correspondence to: R. F. Ivanovic (r.ivanovic@leeds.ac.uk)

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Abstract

The last deglaciation, which marked the transition between the last glacial and present interglacial periods, was punctuated by a series of rapid (centennial and decadal) climate changes. Numerical climate models are useful for investigating mechanisms that underpin the events, especially now that some of the complex models can be run for multiple millennia. We have set up a Paleoclimate Modelling Intercomparison Project (PMIP) working group to coordinate efforts to run transient simulations of the last deglaciation, and to facilitate the dissemination of expertise between modellers and those engaged with reconstructing the climate of the last 21 thousand years. Here, we present the design of a coordinated Core simulation over the period 21–9 thousand years before present (ka) with time varying orbital forcing, greenhouse gases, ice sheets, and other geographical changes. A choice of two ice sheet reconstructions is given, but no ice sheet or iceberg meltwater should be prescribed in the Core simulation. Additional *focussed* simulations will also be coordinated on an ad-hoc basis by the working group, for example to investigate the effect of ice sheet and iceberg meltwater, and the uncertainty in other forcings. Some of these *focussed* simulations will focus on shorter durations around specific events to allow the more computationally expensive models to take part.

1 Introduction

1.1 Climate evolution over the last deglaciation

The last deglaciation is a period of major climate change, when Earth transitioned from its last full glacial state, to the current interglacial climate. The *Last Glacial Maximum* (LGM) marked the culmination of the last glacial cycle when vast ice sheets covered large regions of the Northern Hemisphere, stretching over North America and Eurasia (e.g. Boulton et al., 2001; Dyke et al., 2002; Peltier et al., 2015; Svendsen et al., 2004;

Tarasov et al., 2012), and the Antarctic Ice Sheet expanded to the edge of the continental shelf (Argus et al., 2014; Briggs et al., 2014; Lambeck et al., 2014 and references therein). Changes in the ice sheets resulted in a total sea level rise of ~ 115–130 m between LGM and the late Holocene (Lambeck et al., 2014; Peltier and Fairbanks, 2006) depending upon the time assumed to correspond to the LGM, and ~ 100 m from 21 to 9 ka (the period of focus for this manuscript).

Historically, EPILOG defined the LGM as having occurred 23–19 ka (21 ka centre point), when climate was generally cool and ice sheets were more or less at their largest, based on ice core and sea level records (Mix et al., 2001). It represents the time of maximum terrestrial ice volume. More recently, the last sea level lowstand has been found to have occurred either around 26 ka (Peltier and Fairbanks, 2006) or 21 ka (Lambeck et al., 2014) with relatively stable (low) sea level between those dates. Nearly all ice sheets were at or close to their maximum extent between 26 and 19 ka (Clark et al., 2009).

During the LGM, global annual mean surface temperatures are estimated to have been around $4.0 \pm 0.8\,^{\circ}$C colder than today (Annan and Hargreaves, 2013). The Earth began warming towards its present state from around 19 ka (Fig. 1h; Buizert et al., 2014; Jouzel et al., 2007), as summer insolation at northern high latitudes and global atmospheric greenhouse gas concentrations gradually increased (Fig. 1c–f; Berger, 1978; Loulergue et al., 2008; Lüthi et al., 2008; Marcott et al., 2014). By 9 ka, although the northern ice sheets had not quite retreated (or disappeared) to their present day configuration, most of the Northern Hemisphere deglaciation had taken place (Clark et al., 2012; Lambeck et al., 2014; Peltier et al., 2015; Tarasov et al., 2012; Figs. 1g and 2), with both surface air temperatures (Fig. 1h–i) and atmospheric greenhouse gases (Fig. 1d–f) approaching present day values. However, much of Antarctica remained heavily glaciated well into the Holocene, with the majority of its ice melting between 12 and 6 ka (Argus et al., 2014; Briggs et al., 2014; Mackintosh et al., 2014). Antarctica's total contribution to post-glacial eustatic sea level is poorly constrained, but recent studies have not supported LGM contributions greater than about 15 m eustatic sea level

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equivalent (Bentley et al., 2014; Briggs et al., 2014; Golledge et al., 2013; Mackintosh et al., 2011; Philippon et al., 2006; Whitehouse et al., 2012), emphasising the dominance of North American and Eurasian Ice Sheet dynamics in the global sea level record during the last deglaciation (Argus et al., 2014; Lambeck et al., 2014; Peltier et al., 2015). It should be noted that there is some controversy over whether deglacial ice sheet reconstructions close the global sea level budget (Clark and Tarasov, 2014), with a potential LGM shortfall of "missing ice".

The last deglaciation is not only an interesting case study for understanding multi-millennial scale processes of deglaciation, but also provides the opportunity to study shorter and more dramatic climate changes. Superimposed over the gradual warming trend (Augustin et al., 2004; Jouzel et al., 2007; Petit et al., 1999; Stenni et al., 2011) are several abrupt climate transitions lasting from a few years to a few centuries (examples of which are given below) and it remains a challenge to reconstruct or understand the chain of events surrounding these instances of rapid cooling and warming.

Heinrich Event 1 (approx. 16.8 ka; Hemming, 2004) occurred during the relatively cool Northern Hemisphere Heinrich Stadial 1 (~ 18–14.7 ka). It was characterised by the release of a vast number of icebergs from the North American and Eurasian ice sheets into the open North Atlantic, where they melted. The existence of these iceberg "armadas" is evidenced by a high proportion of ice rafted debris in North Atlantic sediments between 40 and 55° N, predominantly of Laurentide (Hudson Strait) provenance (Hemming, 2004 and references therein). There are several competing theories for the cause of Heinrich Event 1. There is a substantial body of evidence to suggest that it occurred during or was precursory to a period of Atlantic Meridional Overturning Circulation (AMOC) slow down (e.g. Hall et al., 2006; Hemming, 2004; McManus et al., 2004) and weak North Atlantic Deep Water (NADW) formation (e.g. Keigwin and Boyle, 2008; Roberts et al., 2010) under a relatively cold, Northern Hemisphere surface climate (Shakun et al., 2012). Even though the interpretation of a cause and effect link between Heinrich Event 1 and the diminished strength of the AMOC remains rather compelling (e.g. Kageyama et al., 2013), it is increasingly being suggested that the

melting icebergs might not have caused the recorded AMOC slow down, but may have provided a positive feedback to amplify or prolong AMOC weakening and widespread North Atlantic cooling (e.g. Álvarez-Solas et al., 2011; Barker et al., 2015), whilst also causing mid-latitude Atlantic sea surface warming through northward expansion of the subtropical gyre (Naafs et al., 2013).

During the subsequent 14.2–14.7 ka interval, Northern Hemisphere temperatures are seen to have risen by as much as $14.4 \pm 1.9\,^{\circ}$C in just a few decades (Buizert et al., 2014; Goujon et al., 2003; Kindler et al., 2014; Lea et al., 2003; Severinghaus and Brook, 1999), with a dramatic shift in Greenland climate taking place in as little as one to three years (Steffensen et al., 2008). This abrupt event is termed the *Bølling Warming* or *Bølling Transition* (Severinghaus and Brook, 1999). At roughly the same time (~ 14.6 ka), there was a rapid jump in global sea level of 12–22 m in around 350 years or less, known as *Meltwater Pulse 1a* (MWP1a; Deschamps et al., 2012). It is not known exactly which ice mass(es) contributed this 40 mm yr^{-1} (or greater) flux of water to the oceans (e.g. Lambeck et al., 2014; Peltier, 2005). Some older studies have mainly attributed it to a southern source (Bassett et al., 2005, 2007; Carlson, 2009; Clark et al., 1996, 2002; Weaver et al., 2003), whereas more recent work has suggested that at most, less than 4.3 m eustatic sea level equivalent of meltwater could have come from Antarctica (Argus et al., 2014; Bentley et al., 2010, 2014; Briggs et al., 2014; Golledge et al., 2012, 2013, 2014; Licht, 2004; Mackintosh et al., 2011, 2014; Whitehouse et al., 2012) and that Northern Hemisphere ice was the primary contributor (Aharon, 2006; Gregoire et al., 2012; Keigwin et al., 1991; Marshall and Clarke, 1999; Peltier, 2005; Tarasov and Peltier, 2005; Tarasov et al., 2012). Exactly how the Bølling Warming and MWP1a are linked, or what triggered either, remains uncertain.

Ice core records of δD indicate that from around 14.5 to 12.8 ka, the general trend of increasing Southern Hemisphere warming, temporarily stalled (Jouzel et al., 2007; ice core chronology from Veres et al., 2013) for a period known as the *Antarctic Cold Reversal* (Jouzel et al., 1995). Southern Hemisphere cooling is thought to have been relatively widespread, extending from the South Pole to the southern mid-latitudes, with glacial readvance (or stall in glacial retreat) recorded to have peaked 13.0–14.2 ka in Patagonia (García et al., 2012; Kaplan et al., 2011; Strelin et al., 2011) and ∼ 13.0 ka in New Zealand (Putnam et al., 2010; Rother et al., 2014). There are several hypotheses for the cause of the Antarctic Cold Reversal. For example, some have linked it to a change in ocean circulation induced by the delivery of Antarctic ice melt to the Southern Ocean (Menviel et al., 2010, 2011), or possibly as a bipolar response to AMOC recovery and Northern Hemisphere warming during the Bølling Warming (Menviel et al., 2011; Stocker, 1998). Using a CMIP5 level coupled atmosphere–ocean model, Peltier and Vettoretti (2014) and Vettoretti and Peltier (2015) have recently shown that ice core inferred Southern Hemisphere cooling and Northern Hemisphere warming could have been caused by a nonlinear salt oscillator mechanism. Others have argued that a change in Southern Hemisphere winds and ocean circulation is the explanation; for example, a simultaneous northward migration of the southern Subtropical Front and northward expansion of cold water originating in the Southern Ocean (Putnam et al., 2010). The ongoing disagreement over the timing, duration and extent of the Antarctic Cold Reversal means that its cause is difficult to pin down.

The next event of particular interest is the *Younger Dryas cooling*, when Northern Hemisphere temperatures are thought to have dropped by several degrees at 12.8–11.7 ka and most prominently in high latitudes (Buizert et al., 2014; Heiri et al., 2007; Lea et al., 2003; Liu et al., 2012; Simonsen et al., 2011; Steffensen et al., 2008). The event presents a conceptual paradox; the magnitude of the cooling is difficult to reconcile with rising atmospheric CO_2 (approximately +10 ppm compared to the earlier Bølling period ∼ 14.5 ka; Lüthi et al., 2008; Marcott et al., 2014; Veres et al., 2013) and increasing boreal summer insolation (Berger and Loutre, 1991). It is possible that changes in the atmospheric hydrological cycle, such as a shift in source moisture region, could be partly responsible for the $\delta^{18}O$ signal, requiring a smaller temperature anomaly to match the records (Liu et al., 2012). For the climate cooling itself, a rerouting of North American freshwater discharge to the Arctic and/or Atlantic Oceans might have caused a reduction in NADW formation (Broecker et al., 1989; Condron and Win-

sor, 2012; Tarasov and Peltier, 2005). Simulating this period within the context of the preceding climate evolution, could be key to understanding exactly what the surface climate and deep ocean changes were during the Younger Dryas, and how these relate to contemporaneous proxy records.

In this description, we have sought to capture some of the last deglaciation's main climatic events, but there are others that could shape the focus of further study in the working group. For example, early on in the period there is evidence of around 10 m sea level rise taking place in 500–800 years around 20–19 ka (Clark and Mix, 2002; Clark et al., 2004; De Deckker and Yokoyama, 2009; Yokoyama et al., 2001a, b). Whilst the event itself remains somewhat controversial (Cabioch et al., 2003; Hanebuth et al., 2000, 2009; Peltier and Fairbanks, 2006; Shennan and Milne, 2003), it could be the expression of accelerating deglacial ice melt following the Last Glacial Maximum. More recently, the Barbados record of relative sea level history indicates that following the Younger Dryas cooling episode, there may have been another meltwater pulse (Fairbanks, 1989; Peltier and Fairbanks, 2006), referred to as Meltwater Pulse 1b. Significant debate surrounds the magnitude and timing of Meltwater Pulse 1b (Bard et al., 1996; Cabioch et al., 2003; Cutler et al., 2003; Edwards et al., 1993; Shennan, 1999; Stanford et al., 2011) and even its existence, because similar to the 19 ka event, it is not seen in all sea level records spanning the interval (e.g. Bard et al., 1996, 2010; Hanebuth et al., 2000). However, evidence of rapid Antarctic retreat around the time of the event could provide a possible cause for this late deglacial rapid sea level rise (Argus et al., 2014).

1.2 Transient modelling of the last deglaciation

Transient modelling of the last deglaciation is valuable for examining dynamic and threshold behaviours (Braconnot et al., 2012) endemic to the Earth's non-stationary climate system, especially ice–ocean–atmosphere interactions. It is the best tool for reaching a comprehensive understanding of complex and interrelating climate processes with specific regard to chains of events.

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Such simulations are useful for examining the effect of temporally varying climate forcings across the globe and in different environmental systems: what geographical patterns arise and how are they connected, how do these vary through time from seasonal to millennial time scales, and how long does it take before a change in forcing is manifested in a climate response? The spatial coherency of specific events can be investigated to identify processes for simultaneous change as well as lead/lag mechanisms. For example, Roche et al. (2011) investigated patterns of spatial variability in the deglaciation as caused by long-term changes in orbital parameters, atmospheric greenhouse gas concentrations, and ice sheet extent/topography. The results indicated a simultaneous onset of hemispheric warming in the North and South, showing that obliquity forcing was the main driver of the early deglacial warming. In the same investigation, it was found that sea-ice covered regions were the first parts of the world to exhibit significant rises in temperature, implying that a better knowledge of sea-ice evolution could be key to fully understanding the trigger for widespread deglaciation and warming feedbacks. A further example of the insights available into lead–lag relationships provided by long, transient climate simulations under glacial boundary conditions is provided by the previously referenced Dansgaard–Oeschger oscillation-related analyses of Peltier and Vettoretti (2014) and Vettoretti and Peltier (2015), which appear to mimic the Heinrich Stadial 1 to Bølling transition.

Through comparison to geological timeseries data, transient simulations enable the "fingerprinting" of specific climate processes to find out what mechanisms [in the model] can cause recorded climate signals. Comparing complex, global-scale models to combined geological records can provide multiple "fingerprints" in different variables from different archives and in different locations to help narrow down plausible scenarios. For example, Menviel et al. (2011) ran a suite of simulations, varying oceanic meltwater fluxes through the last deglaciation in order to identify which freshwater-forcing scenarios reproduce the Atlantic Ocean circulation state implied by sedimentary records of AMOC strength/depth and ventilation age (Gherardi et al., 2005; McManus et al., 2004 with ages shifted as per Alley, 2000; Thornalley et al., 2011) as well as the Northern

Hemisphere surface climate (Alley, 2000; Bard, 2002; Bard et al., 2000; Heiri et al., 2007; Lea et al., 2003; Martrat et al., 2004, 2007). It was argued that such climate simulations could be used to improve constraints on the timing, duration, magnitude, and location of meltwater inputs to the global ocean.

Liu et al. (2009) used climate "fingerprinting" to identify possible mechanisms for the abrupt Bølling Warming Event, finding that in their model, a forced cessation of freshwater inputs to the North Atlantic (representing ice sheet melt) superimposed on a steady increase in atmospheric CO_2 caused an abrupt resumption in the strength of the AMOC (almost matching a record produced by McManus et al., 2004). This in turn induced a rapid warming in Northern Hemisphere surface climate (close to records from Bard et al., 2000; Cuffey and Clow, 1997; and Waelbroeck et al., 1998) and an increase in tropical rainfall over the Cariaco Basin (comparable to Lea et al., 2003), whilst Antarctic surface temperatures remained relatively stable (similar to Jouzel et al., 2007). Using a suite of simulations from the same model, Otto-Bliesner et al. (2014) went on to suggest that a combination of rapid strengthening of NADW seen by Liu et al. (2009) and rising greenhouse gas concentrations was responsible for increased African humidity around 14.7 ka, matching the model output to a range of regional climate proxies (including deMenocal et al., 2000; Tierney et al., 2008; Tjallingii et al., 2008; Verschuren et al., 2009; Weijers et al., 2007).

Thus, climate proxy fingerprinting can be useful for understanding the spatial coherency of climatic changes and their underlying mechanisms. However, correlation between model and geological data does not guarantee that the correct processes have been simulated; there is always the problem of *equifinality*, whereby the same end state can be reached by multiple means. In a process sense, this may be particularly uncertain when a model does not reproduce the full chain of events that led to a distinguishable climatic signal. For example, mechanisms for many of the major changes in oceanic freshwater inputs proposed by Liu et al. (2009) and Menviel et al. (2011) have not yet been directly simulated (e.g. by dynamic ice sheet models). In both studies, they are imposed as model boundary conditions. Further simulations with different forcing scenarios and from a range of models would help to address such uncertainties.

Transient simulations of the last deglaciation also provide necessary boundary conditions for modelling a variety of Earth System components that may not be interactively coupled to the climate model being used. For example, Gregoire et al. (2015) drove a dynamic ice sheet model with climate data produced by a similar set of simulations to Roche et al. (2011). Using a low resolution GCM, individual climate forcings – including orbit, greenhouse gases, and meltwater fluxes – were isolated so that their relative contribution to melting the modelled North American ice sheets could be examined. The work concluded that the last deglaciation was primarily driven by changes in Northern Hemisphere insolation, causing around 60 % of the North American Ice Sheet melt, whilst increasing CO_2 levels were responsible for most of the remaining changes (Gregoire et al., 2015). The sufficiency of these two forcings for North American glaciation/deglaciation had previously also been identified with fully coupled glaciological and energy balance climate models (Tarasov and Peltier, 1997). Gregoire et al. (2012) were also able to highlight a possible "saddle-collapse" mechanism, whereby gradual warming trends could result in abrupt ice sheet melting events, such as MWP1a and the *8.2 kyr Event*, when a threshold in ice mass balance was crossed. The opening of the ice-free corridor between the Cordilleran and Laurentide ice sheets has long been built into the ICE-NG, Tarasov and Peltier (2004) and Tarasov et al. (2012) sequence of models as geological inferences (Dyke, 2004) indicate that it occurred around the same time as MWP1a.

A further example is given by Liu et al. (2012), who carried out an asynchronous (or "offline") coupling between simulated sea surface temperatures and an isotope-enabled atmospheric model to investigate the Younger Dryas cooling event (∼ 12 ka). The results revised the presupposed Greenland temperatures at this time by 5 °C, demonstrating that changes in moisture source must be an important consideration for the robust interpretation of Greenland ice core $\delta^{18}O$ records and our understanding of high-latitude climate sensitivity. More recently, the same methodology was applied

to understanding Chinese cave records of the East Asian Summer Monsoon 21–0 ka (Liu et al., 2014), not only to better interpret what the speleothem $\delta^{18}O$ tells us about regional hydroclimate variability, but also to understand the wider teleconnections controlling those patterns.

In addition, there are now transient simulations of the last deglaciation from climate models that have been interactively coupled with dynamic ice sheet models (Bonelli et al., 2009; Heinemann et al., 2014) and isotope systems (Caley et al., 2014). Furthermore, a fast Earth System Model of Intermediate Complexity (EMIC) that includes an interactive ice sheet model has been used to look at Earth System dynamics (the role of orbital cycles, aeolian dust, subglacial regolith properties, the carbon cycle, and atmospheric trace gases) on much longer, glacial–interglacial timescales > 120 ka and encompassing the last deglaciation (Bauer and Ganopolski, 2014; Brovkin et al., 2012; Ganopolski and Calov, 2011; Ganopolski et al., 2010). However, the older, uncoupled climate-ice sheet model approach discussed above remains useful because it enables a wider suite of models to be employed than would otherwise be feasible due to limited computational efficiency (e.g. of state-of-the-art, high resolution/complexity models) or software engineering capability. It may also allow for the same Earth System component model (e.g. of ice sheets or $\delta^{18}O$) to be driven by multiple climate models, in order to examine the range of responses and assess [climate] model performance.

With sufficient computational power to make long simulations of the last deglaciation a feasible undertaking, it is timely to coordinate new efforts to ensure that a framework exists to (i) utilise the cutting edge science in climate modelling and palaeoclimate reconstruction, and (ii) robustly intercompare simulations run with different models by different groups and palaeoclimatic data.

1.3 Establishing a new PMIP working group

For twenty years, the Paleoclimate Modeling Intercomparison Project (PMIP) has been internationally coordinating multi-model simulations with complex climate models in order to evaluate model performance and better understand [past] climate changes (Braconnot et al., 2007, 2012; PMIP website, 2007). Currently entering its fourth phase, PMIP is a growing organisation that continues to contribute towards other coordinated efforts to understand present day climate change; including the Coupled Model Intercomparison Project (CMIP; e.g. Taylor et al., 2011a, b) and the Intergovernmental Panel on Climate Change's (IPCC) Assessment Reports (e.g. the Fifth Assessment Report; Flato et al., 2013; Masson-Delmotte et al., 2013). It encompasses a broad range of models, from very fast, lower resolution EMICS, through a range of coupled GCMs to the latest generation of higher resolution and complexity Earth System Models. Thus, the main challenges for the fourth Phase of PMIP include: designing experiments that are suitable for all of its participants; addressing sufficiently fundamental questions to be of interest to the EMIC community; defining adequately focused scope for the feasible participation of the latest generation of ESMs; and prescribing flexible model setups that can be implemented in this range of models, whilst maintaining the ability to robustly compare results.

One of the most recent working groups to be established in PMIP is the Last Deglaciation Working Group. With the aim of coordinating transient simulations of the last deglaciation, the challenge of including the full range of PMIP models is at the forefront of our experiment design. The experiment will be partitioned into three phases (Fig. 1b and Sect. 4), which will form milestones for managing its long duration (12 thousand years) as well as for scheduling any shorter, alternative simulations to the Core.

The aim of this paper is to outline the model setup for the transient Core simulation of the last deglaciation, specifically for the sub-period of 21–9 ka. Prescribed boundary conditions include orbital parameters, atmospheric trace gases and ice sheets. In association with the ice sheet reconstructions, we also provide bathymetric, orographic and land–sea mask evolution.

1.4 Approach

One of the roles of PMIP has been to systematically study the ability of climate models to retrodict different past climates for which there are "observational" data from geological archives (e.g. Braconnot et al., 2000, 2007, 2012; Haywood et al., 2010; Joussaume et al., 1999; Kageyama et al., 2006; Kohfeld and Harrison, 2000; Masson-Delmotte et al., 2006; Otto-Bliesner et al., 2009; Weber et al., 2007). In this vein, many palaeoclimate model intercomparison projects have been designed to facilitate the robust comparison of results from the same "experiment" (i.e. simulation set) across a range of different models, usually taking a prescriptive approach to model setup to ensure that any differences observed in the results are attributable to differences in model structure and not to differences in chosen "boundary conditions" and climate forcings. However, as Schmidt et al. (2011) point out, the choice of one particular configuration from a range of plausible boundary conditions and forcings is often arbitrary and does not account for uncertainties in the data used for developing the forcings/boundary conditions. Moreover, in designing the PMIP last deglaciation experiment, we have attempted to strike a balance between establishing a framework within which to assess model differences and performance, and taking the opportunity to utilise the full range of PMIP climate models (Earth System, General Circulation and Intermediate Complexity) to examine uncertainties in deglacial forcings, trigger-mechanisms and dynamic feedbacks. Consequently, forcings/boundary conditions that are relatively well established (atmospheric trace gases and orbital parameters) are tightly constrained in the Core experiment design. Others are given with multiple precisely described possibilities to choose from (ice sheet reconstructions) and the remainder (e.g. aerosols and vegetation) are left to the discretion of individual participants, although we recommend the use of preindustrial values when they are not model prognostics. Further to this, it will be left to the expert user to decide how often to make manual updates to those boundary conditions that cannot evolve automatically in the model, such as bathymetry, orography and land sea mask.

In addition to the Core, we will also coordinate additional experiments that are designed to:

i. explore uncertainties in the boundary conditions and climate forcings

ii. test specific hypotheses for mechanisms of climate change

iii. focus on shorter time periods (for example, abrupt events) and thus include computationally expensive models for which a twelve thousand year simulation is unfeasible.

These optional simulations will be referred to as *focussed* experiments, and participants are encouraged to contribute towards the design and coordination of these simulations within the working group (https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:wg:degla:index).

The start date for the experiment has been chosen to be in line with PMIP's historical definition of the LGM; 21 ka (e.g. Braconnot et al., 2000; Kohfeld and Harrison, 2000; Abe-Ouchi et al., 2015). However, we are aware that some groups may prefer to begin their simulations from the earlier date of 26 ka (around the last sea level lowstand; Clark et al., 2009; Lambeck et al., 2014; Peltier and Fairbanks, 2006) and both orbital and atmospheric trace gas parameters will be provided from this earlier date. Although the working group's focus will at least initially be 21–9 ka, boundary conditions for the Core simulation will be provided from 21 ka to the preindustrial (26 ka to the preindustrial for orbital insolation and trace gases).

The following is not meant to be an exhaustive review of climate forcing reconstructions through the last deglaciation. Instead, our intention is to consolidate the current knowledge in a practical experiment design for a range of climate models. Within this coordinated context, the aim is to explore the forcings and underlying feedback mechanisms for the rapid climate events that punctuated the gradual warming and deglaciation of the Earth.

The paper is structured so that Sect. 2 outlines the model boundary conditions and climate forcings for the Core simulation. Section 3 presents how we will ensure the

feasible participation of a range of climate models with different complexity and computational efficiency, as well as the plan to run additional, targeted, hypothesis- and sensitivity-led simulations. Section 4 discusses the three phases of the long Core experiment.

2 Core simulation (21 to 9 ka)

The Core simulation for the last deglaciation will focus on the period from 21 to 9 ka, although there will also be the option to spin up the simulation with time-evolving orbital and trace gas parameters from 26 ka and all boundary conditions will be available from 21 ka to the preindustrial. Recommendations for the initialisation state at 21 ka are summarised in Table 1 and described below (Sect. 2.1). Prescribed boundary conditions include insolation via the Earth's astronomical parameters (Sect. 2.2), atmospheric trace gases (Sect. 2.3), ice sheets (Sect. 2.4), meltwater fluxes (Sect. 2.5), and orography/bathymetry (Sect. 2.6), as summarised in Table 2. Boundary condition data for the Core simulation are provided on the PMIP wiki; https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:wg:degla:bc:core (PMIP Last Deglaciation Working Group, 2015).

2.1 Last Glacial Maximum spinup

There is a choice of two possibilities for starting the last deglaciation Core simulation. Either the simulation should be initialised from the end of a spun-up, PMIP-compliant LGM (21 ka) simulation, or a simulation with transient orbital and trace gas forcing should be run from an earlier time period (orbital and trace gas parameters will be provided from 26 ka onwards). Whichever method is applied, we require that it is comprehensively documented along with information on the model's state of spinup at 21 ka (e.g. timeseries of surface climates, maximum strength of the North Atlantic Meridional

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Overturning Circulation stream function, net radiation at the top of the atmosphere etc.).

2.1.1 Equilibrium-type spinup (21 ka)

For setting up an equilibrium-type spinup, please make sure to use the following constraints, which may differ from other PMIP 21 ka simulation protocols:

- Insolation should be set so that eccentricity is 0.018994, obliquity is 22.949°, perihelion–180° is 114.42°, the date of the vernal equinox is 21 March at noon, and the solar constant is the same as for the preindustrial (e.g. 1365 $W m^{-2}$, as in the PMIP3-CMIP5 preindustrial experiment). These are consistent with previous PMIP LGM boundary conditions (PMIP LGM Working Group, 2010).
- Prescribed atmospheric trace gases should be as follows: CO_2 at 188 ppm, CH_4 at 375 ppb, N_2O at 200 ppb (Fig. 3), with CFCs at 0 and O_3 at the PMIP3-CMIP5 preindustrial value (e.g. 10 DU). This is to be compatible with the time-evolving boundary conditions for the Core simulation (Sect. 2.3). Note that the LGM atmospheric CO_2 and CH_4 concentrations have changed slightly from earlier LGM experiments (e.g. PMIP3, which used 185 ppm and 350 ppb, respectively; PMIP LGM Working Group, 2010). However, N_2O remains at 200 ppb, which is more representative of the longer glacial period than the 187 ppb concentration recorded at 21 ka (Fig. 3c). These updates are in line with the latest ice core age model, (AICC2012; Veres et al., 2013), which is also used for the transient forcings described below (Sect. 2.3).
- Prescribed ice sheets should use either the GLAC-1D or ICE-6G_C reconstruction at 21 ka (see Sect. 2.4). The associated topography and coastlines should be used as per the chosen ice sheet reconstruction; it is optional whether or not to implement the associated bathymetry. These data will be provided with the ice

sheet reconstructions. Whichever ice sheet reconstruction is chosen for the LGM spinup should be carried through to the Core transient simulation.
- Global ocean salinity should be +1 psu, compared to preindustrial, to account for the increased terrestrial ice mass at the LGM (PMIP LGM Working Group, 2015).
- Any other boundary conditions should be set to be consistent with the Core transient simulation to follow (Sects. 2.2–2.7).

On the freshwater budget, PMIP advises groups to "carefully check the fresh water budget in their LGM experiments in order to avoid unnecessary drifts of the ocean salinity. It can be necessary to route the snow which has fallen in excess on the ice sheets to the ocean. Given the change in coastlines, it is also sometimes necessary to relocate the large river estuaries on the coast" (PMIP LGM Working Group, 2015). Tarasov and Peltier (2006) provides a glaciological example of the possible re-routings for North America.

The integration time required for spinning up the LGM climate state should be decided on a case-by-case basis by the user. Groups may choose to initialise their equilibrium-type simulation from other PMIP LGM runs. However, please be careful. Some of the boundary conditions for the PMIP4-CMIP6 (not finalised at the time of writing) and previous PMIP LGM simulations are different to the setup outlined here, specifically in terms of ice sheets and trace gases concentrations, and therefore need to be adapted to match these requirements. Please also provide timeseries data for the diagnosis of model [dis]equilibrium at 21 ka (introduction to Sect. 2.1).

2.1.2 Transient orbital and trace gas parameters (26–21 ka)

If this is the preferred option to initialise the Core, it is recommended that the simulation is setup as per Sect. 2.1.1, but with time-evolving orbital and trace gas parameters instead of fixed ones. Specifically for orbit, the eccentricity, obliquity, perihelion–180° and date of the vernal equinox values listed above should be replaced with their transient equivalents, as per Berger (1978). For the atmospheric trace gases, carbon dioxide, methane and nitrous oxide values should be replaced with the transient equivalents provided on the PMIP Wiki (https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:wg:degla:bc:core) and according to Lüthi et al. (2008), Loulergue et al. (2008) and Schilt et al. (2010), respectively, on the AICC2012 chronology (Veres et al., 2013); Fig. 3.

In this case, all other boundary conditions should remain fixed in line with the LGM equilibrium-type experiment design until 21 ka, when the fully transient Core simulation begins. This transient spin-up can be initialised from a spun-up previous LGM, cold ocean, preindustrial, or observed present day ocean simulation.

2.2 Insolation (21–9 ka)

As per Sect. 2.1, the solar constant should be fixed to the established preindustrial conditions (e.g. 1365 $W m^{-2}$) throughout the run, which is the PMIP preindustrial experiment setup (PMIP LGM Working Group, 2015). However, the orbital parameters should be time-evolving through the deglaciation to follow Berger (1978); e.g. Fig. 1c.

2.3 Atmospheric trace gases (21–9 ka)

For the deglaciation, CFCs should be fixed at 0, and O_3 should be set to PMIP3-CMIP5 preindustrial values (e.g. 10 DU), as used for the LGM. When a model is not running with dynamic atmospheric chemistry, the remaining trace gases should be time-evolving, with CO_2 following Lüthi et al. (2008), CH_4 following Loulergue et al. (2008) and N_2O following Schilt et al. (2010), all adjusted to the AICC2012 chronology (Veres et al., 2013); Fig. 1d–f.

Temporally higher resolution CO_2 data from the West Antarctic Ice Sheet Divide has been provided by Marcott et al. (2014), spanning 23–9 ka ("WDC" on Fig. 3a). However, the newer data are consistently offset from other Antarctic ice core data by ~ 4 ppm and the cause for this remains unresolved. Furthermore, although the data encompasses the last deglaciation (and the period we are focussing on; 21–9 ka), it

would not be easily spliced into a longer record (e.g. for groups wishing to run their simulations through to the present day). This is why the higher resolution data (Marcott et al., 2014) will not be used for the Core, reverting to the older record from Lüthi et al. (2008). However, it may form the basis of a coordinated additional simulation, which will be optional for participant groups. Other sensitivity-type simulations could also be coordinated to assess the influence of timing in the CO_2 records on climate and ice sheet evolution, addressing age model uncertainty. The details of the setup for such *focussed* simulations will be discussed and determined at a later date.

It is noted that the N_2O value from Schilt et al. (2010) and Veres et al. (2013) does not match the previously defined LGM N_2O concentration (Sect. 2.1.1); 187 ppb compared to 200 ppb (Fig. 3c). This is because the N_2O record is highly variable during the last glacial lowstand (26–21 ka), with a range of ~ 33 ppb (183–216 ppb) and a mean of 201 ppb. Thus 200 ppb seems a reasonably representative N_2O concentration for the spinup phase of the simulation, although the Core simulation will start with the more chronologically accurate value of 187 ppb.

2.4 Ice sheet reconstructions (21–9 ka)

For the Core experiment, ice sheet extent and topography should be prescribed from one of two possible reconstructions: ICE-6G_C (Figs. 2a and 4a) and GLAC-1D (Figs. 2b and 4b).

The ICE-6G_C reconstruction is fully published (Argus et al., 2014; Peltier et al., 2015), and the reader is directed to this literature for further information. The GLAC-1D reconstruction is combined from different sources (Briggs et al., 2014; Tarasov and Peltier, 2002; L. Tarasov, personal communication, 2014; Tarasov et al., 2012) and whilst it is mostly published, there are some new components; therefore, a short description follows. The Eurasian and North American components are from Bayesian calibrations of a glaciological model (Tarasov et al., 2012; L. Tarasov, personal communication, 2014), the Antarctic component is from a scored ensemble of 3344 glaciological model runs (Briggs et al., 2014) and the Greenland component is the handtuned glaciological model of Tarasov and Peltier (2002). All four of the GLAC-1D ice sheet components employ dynamical ice sheet models that have been constrained with relative sea level data. Where available, they have also been constrained by geologically-inferred deglacial ice margin chronologies, pro-glacial lake levels, ice core temperature profiles, present-day vertical velocities, past ice thickness, and present day ice configuration. Details of exactly how these constraints were derived and applied are given in the relevant references above. The four components (North American, Eurasia, Antarctica and Greenland) were combined under Glacial Isostatic Adjustment (GIA) post-processing for a near-gravitationally self-consistent solution (Tarasov and Peltier, 2004), which was tested against complete Glacial Isostatic Adjustment solutions (L. Tarasov, personal communication, 2014). The topography in the global combined solution was adjusted in Patagonia and Iceland following ICE-5G (Peltier, 2004), but the changes in these ice caps are not reflected in the ice mask.

Both datasets include ice extent and topography at intervals of 1000 years or less through the deglaciation. Ice extent is provided as a fractional ice mask for ICE-6G_C and a binary ice mask in GLAC-1D.

The two reconstructions incorporate similar constraints for North American ice sheet extent (i.e. Dyke, 2004). For Eurasia, ICE-6G_C follows the ice extent provided by Gyllencreutz et al. (2007), whereas GLAC-1D uses data from Hughes et al. (2015). The reconstructions only differ slightly in their ice extent evolution (Figs. 2 and 4), for example the Barents Sea deglaciates earlier in GLAC-1D than in ICE-6G_C (Fig. 2). The main differences between the reconstructions are in the shape and volume of individual ice sheets. In particular, the North American Ice Sheet reaches an elevation of 4000 m in ICE-6G_C, but is only 3500 m high in GLAC-1D. Similarly, the shape and thickness of the Barents Sea Ice Sheet are not the same in the two reconstructions. The ICE-6G_C dataset is been provided at 1° and 10 min horizontal resolution, GLAC-1D is provided at 1° horizontal resolution.

Ice surface elevation (topography) should be implemented as an anomaly from present day topography and added to the model's present day topography after re-

gridding onto the model resolution, following the LGM experimental protocol (PMIP LGM Working Group, 2010, 2015). Land surface properties will need to be adjusted for changes in ice extent. Where ice retreats, land surface should be initialised as bare soil if a dynamic vegetation model is used, otherwise use prescribed vegetation (see Sect. 2.7) with appropriate consideration of soil characteristics. Where ice is replaced by ocean, it is advised to follow the procedure for changing coastlines described in Sect. 2.7. Inland lakes can be prescribed based on the ice sheet and topography reconstructions, but this is not compulsory. It is also optional whether to include changes in river routing basins and outlets, which can be calculated from the provided topography and land–sea mask data (see Sect. 2.6).

Groups are free to choose how often to update ice extent and elevation. This could be done at regular intervals (e.g. the 1000 year time slices provided) or at specific times during the deglaciation, as was done in the TraCE-21 ka experiment (Liu et al., 2009). Changes in ice extent can have a large impact on climate through ice albedo changes and feedbacks. We thus recommend that when possible, ice sheets are not updated at times of abrupt regional or global climate change, particularly the events that the working group will focus on, as this could artificially introduce stepped shifts in climate. Groups are also advised to consider that ice sheet boundary conditions may need to be updated more often at times of rapid ice retreat. The timing and way in which land ice changes are implemented must be documented.

Alternative ice sheet reconstructions or simulations can be used to test the sensitivity of climate to this boundary condition. Simulations with coupled ice sheet-climate models are also welcomed. Although these will not form part of the Core, for which ICE-6G_C or GLAC-1D should be used, they will be coordinated as important supplementary *focussed* simulations.

2.5 Ice meltwater

The Core simulation will not include any prescribed ice melt (i.e. freshwater fluxes) to the ocean. This may seem controversial given the levels of terrestrial ice sheet melt and sea level rise known to have taken place during this period (e.g. Lambeck et al., 2014) and the historical importance attached to the influence of [de]glacial freshwater fluxes on climate (e.g. Broecker et al., 1989; Condron and Winsor, 2012; Ganopolski and Rahmstorf, 2001; Liu et al., 2009; Rahmstorf, 1995, 1996; Teller et al., 2002; Thornalley et al., 2010; Weaver et al., 2003). However, considering the current uncertainty on exactly when and where ice melt entered the ocean during the last deglaciation (e.g. discussion of MWP1a in Sect. 1.1), this is the best way to ensure that the Core experiment is based on robust geological data. Furthermore, there is an ongoing debate over the role of catastrophic freshwater fluxes in bringing about abrupt deglacial climate change and several alternative or complementary mechanisms have been proposed (e.g. Adkins et al., 2005; Álvarez-Solas et al., 2011; Barker et al., 2010, 2015; Broecker, 2003; Hall et al., 2006; Knorr and Lohmann, 2003, 2007; Roche et al., 2007; Rogerson et al., 2010; Thiagarajan et al., 2014). In light of this, and because we are keen to see what the climate response to non-freshwater-forced scenarios will be in the PMIP models, the decision has been made to have no prescribed freshwater fluxes in the Core simulation. This experiment is thus designed to constitute a reference for experiments in which fresh water fluxes will be introduced.

Moreover, a thorough investigation of the extent to which non-freshwater-forced climate evolution matches the geological records has merit in its own right; can abrupt deglacial changes be simulated without ice-meltwater, as has been proposed (e.g. discussion above)? To what extent can "observed" patterns be attributed to better constrained forcings, such as atmospheric CO_2 and Earth's orbit? To complete the investigation, freshwater-flux scenarios will be targeted by opt-in *focussed* simulations that test specific ice-melt hypotheses as well as instances where/when the Core falls short of the "observed" patterns. For example, routing of ice melt computed from GLAC-1D (Sect. 2.4) will be provided as a possible transient boundary condition.

2.6 Topography, bathymetry, coastlines and rivers

Changes in the ice sheets and their glacial eustatic and isostatic influence affected continental topography and ocean bathymetry, which in turn shifted the coordinates of river mouths and the coastal outline throughout the deglaciation. Hence time-varying topographic, bathymetric and land–sea mask fields that match the chosen ice sheet from Sect. 2.4 (i.e. ICE-6G_C or GLAC-1D) should be used.

Topography should be updated at the same time as the model's ice sheet is updated; this is mainly implicit to implementing the ice sheet reconstruction because the major orographic changes through the deglaciation relate directly to ice sheet evolution. This said, due to glacial isostatic adjustment components in the ice sheet reconstructions, there is evolution in continental topography that is not directly the lowering/heightening of the ice surface, and it is up to individuals whether they incorporate this or mask only the changes in ice sheet orography.

Ocean bathymetry will be provided, but is an optional boundary condition to vary through time. Coastlines, on the other hand, will need to be varied according to changes in global sea level (and each model's horizontal grid resolution). It will be left to the discretion of participants to decide how often to update either boundary condition, and when deciding on their frequency it is recommended that groups consider the implications for opening/closing seaways and their effect on ocean circulation and climate. Furthermore, the frequency need not be regular and may instead focus on key "events" in the marine [gateway] realm. However, whenever possible and foreseeable, groups are encouraged to avoid making stepwise changes to model boundary conditions that would interfere with signals of abrupt climate change; particularly those events that the working group aims to focus on (e.g. Heinrich Event 1, the Bølling Warming, MWP1a, the Younger Dryas etc.) unless the forcing (e.g. opening of a gateway) is assumed to be linked with the event.

If groups wish, model river networks can be remapped to be consistent with this and updated on the same timestep as the ice sheet reconstruction, either manually or by the

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model. However, it is appreciated that the technical challenges associated with such a methodology would be impractical for many. Therefore, following the recommendation of the PMIP3 LGM Working Group (2010), "river pathways and basins should be at least adjusted so that fresh water is conserved at the Earth's surface and care should be taken that rivers reach the ocean" at every timestep that the bathymetry is adjusted; for example, when sea levels were lower, some river mouths may need to be displaced towards the [new] coastline to make sure they reach the ocean.

2.7 Vegetation, land surface and other forcings

In this section, recommendations are made for last deglaciation vegetation, land surface and aerosol (dust) parameters in the model.

There are three recommended options for setting up the Core simulation's vegetation and land surface parameters, they can either be: (i) computed using a dynamical vegetation model (e.g. coupled to the atmospheric component of the model), (ii) prescribed to match the CMIP5 preindustrial setup (Taylor et al., 2011a, b) with fixed vegetation types and fixed plant physiology (including leaf area index), or (iii) prescribed to match the CMIP5 preindustrial setup (Taylor et al., 2011a, b) with fixed vegetation types and interactive plant physiology if running with an enabled carbon cycle. If prescribing vegetation and land surface, i.e. using option (ii) and (iii), groups should be aware that coastal land will be emerged compared to preindustrial because of the increased terrestrial ice volume and associated lower eustatic sea level (with the maximum during the early stages of the Core). Therefore, vegetation/land surface will need to be interpolated onto the emerged land from preindustrial grid cells, for example using nearest neighbour methods.

For models with prognostic aerosols, the parameters for dust [forcing] can be computed dynamically. Alternatively, it is recommended that Core simulations fix the associated parameters according to the CMIP5 preindustrial simulation (Taylor et al., 2011a, b), with no temporal variation.

It has already been described that for the LGM (i.e. the very start of the Core simulation), groups are recommended to adjust the global freshwater budget by +1 psu to account for the increased [terrestrial] ice volume (Sect. 2.1.1). If salinity is reset at any subsequent point (e.g. to correct for model drifts or to account for ice volume changes), this must be documented.

There is no last deglaciation protocol for setting up other forcings, transient or fixed in time. For all simulations, groups are required to fully document their methods, including experiment design and especially when different or with additional components to the setup described here.

3 Coordinating further simulations

As already alluded to, we are faced with the challenge of designing an experiment that is suitable to be run with a wide range of models, from the more computationally efficient class of intermediate complexity models, to state-of-the-art Earth System Models. One particular difficulty is enabling the most complex and highest resolution climate models to participate in this 12 thousand year long experiment when for some, even the integration to reach the LGM spinup state demands a huge amount of computational resource. There is no easy solution and our approach will be to augment the Core simulation with shorter *focussed* simulations that target specific questions, mechanisms and time periods. Whilst the most computationally expensive models (e.g. the latest generation of Earth System Models) may not feasibly be able to participate in the Core, they will be included in the shorter subset of *focussed* simulations. Similarly, alternative full-deglaciation simulations can be coordinated for the less computationally expensive models in the working group (e.g. low resolution General Circulation Models, and Earth System Models of Intermediate Complexity).

One line of investigation relating to meltwater inputs from ice sheets and icebergs is to carry out a suite of sensitivity simulations examining different injection sites. These simulations would help to address some of the uncertainty that led to the exclusion

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of freshwater fluxes from the Core. For example, geochemical evidence suggests that smaller and more localised discharges of freshwater than have traditionally been considered in climate models may have an important influence on ocean circulation (e.g. Hall et al., 2006), implying that precise freshwater fluxes are needed in the models to examine their effect. Certainly, others have shown that the location of injection is a controlling factor on the impact of freshwater delivery to the ocean, not just laterally (e.g. Condron and Winsor, 2012; Smith and Gregory, 2009), but also in terms of depth (e.g. Roche et al., 2007).

A set of coordinated simulations exploring a range of uncertainty in the freshwater forcing (location, depth, duration, magnitude, and physical characteristics such as temperature and density) would be well suited for the *focussed* experiments, thus building on the meltwater-free Core. However, freshwater is not the only issue and other *focussed* experiments could include the influence of timing in greenhouse gas records (e.g. Lüthi et al., 2008; compared to Marcott et al., 2014), differences in ice sheet reconstructions (e.g. the PMIP3 merged ice sheet from Abe-Ouchi et al., 2015, ICE-6G_C, GLAC-1D) or simulations with [coupled] ice-sheet models, the relative importance of different forcings (e.g. insolation vs. trace gases vs. ice sheet evolution), event-specific hypothesis testing, and shorter-term variability within the climate system. Based on ongoing discussions, it is likely that the first set of *focussed* simulations will be hypothesis-driven, investigating possible mechanisms for preconditioning the glacial ocean for the relatively cool Heinrich Stadial 1 and ensuing catastrophic iceberg discharge (Barker et al., 2015).

We have described the plans for *focussed* simulations to highlight the depth of the working group's aims and to properly contextualise the Core simulation, but the purpose of this manuscript is to outline the model setup for the Core simulation. The experiment design for subsequent *focussed* simulations will be described at a later date on the PMIP Last Deglaciation Working Group Wiki (https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:wg:degla:index) and we welcome contributions to the discussion of what further simulations to coordinate there.

4 Working group phases

The experiment will be split into three phases that are designed to run seamlessly into each other (Fig. 1a). Phase one begins at the LGM (21 ka) and will finish at the abrupt Bølling Warming event, which is where Phase 2 picks up, encompassing the Bølling Warming. Phase 3 begins at the start of the Younger Dryas cooling and is currently planned to continue through to the end of the Core simulation at 9 ka.

Perhaps most importantly, this affords near-future milestones for managing the ultimate completion of the long full deglacial simulation across all participant groups. It will provide a timetabled framework for beginning and continuing the longer simulations; for scheduling shorter, event- or challenge-specific transient simulations by more computationally expensive models (see discussion in Sect. 3); and for the analysis and publication of results as the milestones are reached. Another motivation is to ensure that the experiment design for later periods of the last deglaciation is updated according to knowledge gained from simulations of the preceding time period; for example, changes in ocean and climate states, which have previously been shown to have a strong influence on climate trajectories (e.g. Kageyama et al., 2010; Timm and Timmermann, 2007). Splitting the period into phases also provides the opportunity to update model boundary conditions and climate forcing data with cutting edge palaeoclimate reconstructions, as they emerge during the lifespan of the multi-model experiment. However, care will be taken to ensure that these are physically consistent between phases.

Each phase will encompass at least one distinguishable climate event; Heinrich Stadial 1 and Heinrich Event 1 in Phase 1 following on from the LGM; Meltwater Pulse 1a, the Bølling Warming and the Antarctic Cold Reversal in Phase 2; and the Younger Dryas cooling in Phase 3 (Fig. 1b). As outlined in Sect. 3, simulations of these shorter events can be coordinated in the *focussed* simulations. This is to engage the higher complexity/resolution models, which are unable to run longer simulations, but can use the wider framework of the working group to provide valuable knowledge on rapid climate changes known to have taken place in the last 21 ka.

5 Summary

The last deglaciation presents a host of exciting opportunities to study the Earth System and in particular, to try to understand a range of abrupt climate changes that occurred over just a few years to centuries within the context of more gradual trends. Numerical climate models provide useful tools to investigate the mechanisms that underpin the events of this well-studied time period, especially now that technological and scientific advances make it possible to run multi-millennium simulations with some of the most complex models. Several recent modelling studies have begun this task, but many questions and untested hypotheses remain. Therefore, under the auspices of the Paleoclimate Modelling Intercomparison Project (PMIP), we have set up an initiative to coordinate efforts to run transient simulations of the last deglaciation, and to facilitate the dissemination of expertise between modellers and those engaged with reconstructing the climate of the last 21 thousand years.

The first step has been to design a single, Core simulation suitable for a range of PMIP models; from relatively fast and coarse resolution Earth System Models of Intermediate Complexity, to new generations of the more complex and higher resolution General Circulation and Earth System Models. The setup for this Core simulation, is based on an approach that tries to combine a traditional Model Intercomparison Project method of strictly prescribing boundary conditions across all models, and the philosophy of utilising the breadth of participants to address outstanding uncertainty in the climate forcings, model structure and palaeoclimate reconstructions. Accordingly, we have made recommendations for the initialisation conditions for the simulation and have stated our minimum requirements for the transient experiment design, as summarised in Tables 1 and 2, respectively.

However, there are some uncertainties that the Core is not designed to deal with directly; two examples discussed in this manuscript being the influence of ice melt on the oceans and climate, and the effect of timing in the trace gas records. We know that the Core simulation will not tackle all of our questions, and is likely to give rise to

others. Therefore, additional *focussed* simulations will also be coordinated on an ad-hoc basis by the working group. Many of these will build on and be centred around the Core; often taking shorter snapshots in time, thus including the most computationally expensive models in the experiment, or presenting twelve-thousand year alternatives to the Core for faster models to contribute. Not all simulations will be suitable for all models, but the aim is that taken as a whole, the experiment can utilise the wide range of PMIP model strengths and hence minimise individual weaknesses.

Essentially, the Core simulation has been designed to be inclusive, taking into account the best compromise between uncertainties in the geological data and model limitations. The hypothesis-driven *focussed* experiments will go further than the Core to target the questions that remain. It is hoped that this exciting initiative will improve our individual efforts, providing new opportunities to drive the science forwards towards understanding this fascinating time period, specific mechanisms of rapid climate warming, cooling and sea level change, and Earth's climate system more broadly.

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Table 1. Summary of recommended model boundary conditions to spin up the last deglaciation Core simulation (pre 21 ka); see text for details. Participants are not required to follow the recommendation for these boundary conditions, but must document the method used, including information on the simulation's state of spinup at the point when the Core is started. Data are available from PMIP Last Deglaciation Working Group Wiki: https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:wg:degla:index. Boundary condition group headings are in bold.

Spinup type	Boundary condition	Description
Last Glacial Maximum (LGM; 21 ka)	**Insolation**	
	Solar constant	Preindustrial (e.g. 1365 $W m^{-2}$)
	Eccentricity	0.018994
	Obliquity	22.949°
	Perihelion–180°	114.42°
	Vernal equinox	Noon, 21 Mar
	Trace gases	
	Carbon dioxide (CO_2)	188 ppm
	Methane (CH_4)	375 ppb
	Nitrous oxide (N_2O)	200 ppb
	Chlorofluorocarbon (CFC)	0
	Ozone (O_3)	Preindustrial (e.g. 10 DU)
	Ice sheets, orography and coastlines	21 ka data from either: – ICE-6G_C (references in text) – GLAC-1D (references in text)
	Bathymetry	Keep consistent with the coastlines, using either: – Data associated with the ice sheet – Preindustrial bathymetry
	Global ocean salinity	+1 psu, relative to preindustrial
Transient orbit and trace gases (26–21 ka)	**Orbital parameters**	All orbital parameters should be transient, as per Berger (1978) 26–21 ka
	Trace gases	Adjusted to the AICC2012 (Veres et al., 2013)
	Carbon dioxide (CO_2)	Transient, as per Lüthi et al. (2008)
	Methane (CH_4)	Transient, as per Loulergue et al. (2008)
	Nitrous oxide (N_2O)	Transient, as per Schilt et al. (2010)
	All others	As per LGM (21 ka) spinup type.

Table 2. Summary of required model boundary conditions for the last deglaciation Core simulation 21–9 ka; optional boundary conditions are labelled as such. Data are available from PMIP Last Deglaciation Working Group Wiki: https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:wg:degla:index. See text for details. Boundary condition group headings are in bold.

Boundary condition	Description
Initial conditions (pre 21 ka)	Recommended (optional) to use either: – Last Glacial Maximum (LGM; 21 ka) equilibrium simulation, including +1 psu global ocean salinity – Transient orbit and trace gases (26–21 ka) and all other boundary conditions fixed as per equilibrium LGM See Table 1 for details. The method must be documented, including information on the state of spinup
Insolation	
Solar constant	Preindustrial (e.g. 1365 $W\,m^{-2}$)
Orbital parameters	Transient, as per Berger (1978)
Trace gases	Adjusted to the AICC2012 age model (Veres et al., 2013):
Carbon dioxide (CO_2)	Transient, as per Lüthi et al. (2008)
Methane (CH_4)	Transient, as per Loulergue et al. (2008)
Nitrous oxide (N_2O)	Transient, as per Schilt et al. (2010)
Chlorofluorocarbon (CFC)	0
Ozone (O_3)	Preindustrial (e.g. 10 DU)
Ice sheet	Transient, with a choice of either: – ICE-6G_C (references in text) – GLAC-1D (references in text) How often to update the ice sheet is optional
Orography and coastlines	Transient. To be consistent with the choice of ice sheet. Orography is updated on the same timestep as the ice sheet. It is optional how often the land–sea mask is updated.
Bathymetry	Keep consistent with the coastlines and use either: – Transient data associated with the chosen ice sheet; it is optional how often the bathymetry is updated. – Preindustrial bathymetry
River routing	Ensure that rivers reach the coastline It is recommended (optional) to use one of the following: – Preindustrial configuration for the model – Transient routing provided with the GLAC-1D ice sheet – Manual/model calculation of river network to match topography
Freshwater fluxes	No land ice or iceberg meltwater fluxes to the ocean
Other (optional)	
Vegetation and land cover	Prescribed preindustrial cover or dynamic vegetation model
Aerosols (dust)	Prescribed preindustrial distribution or prognostic aerosols

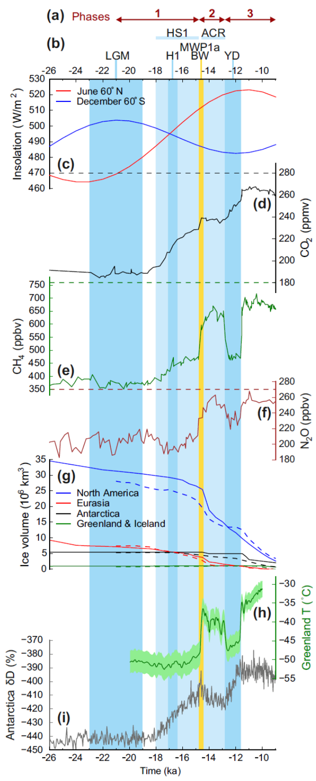

Figure 1. The last deglaciation; forcings and events. **(a)** The three phases of the Core simulation (Sect. 4). **(b)** Climate events/periods discussed in the text; Last Glacial Maximum (LGM; 23–19 ka as according to the EPILOG definition; Mix et al., 2001), Heinrich Stadial 1 (HS1), Heinrich Event 1 (H1), Bølling Warming (BW) and Meltwater Pulse 1a (MWP1a), Antarctic Cold Reversal (ACR) and the Younger Dryas cooling (YD). **(c)** June insolation at 60° N and December insolation at 60° S (Berger, 1978). **(d)** Atmospheric carbon dioxide concentration (composite of EPICA Dome C, Vostok and Taylor Dome records, Antarctica; Lüthi et al., 2008); black dashed line shows preindustrial concentration. **(e)** Atmospheric methane concentration (EPICA Dome C, Antarctica; Loulergue et al., 2008); green dashed line shows preindustrial concentration. **(f)** Atmospheric nitrous oxide concentration (Talos Dome, Antarctica; Schilt et al., 2010); brown dashed line shows preindustrial concentration. **(g)** Volume of the ice sheets according to the ICE-6G_C reconstruction (solid lines; Argus et al., 2014; Peltier et al., 2015) and the GLAC-1D reconstruction (dashed lines; Briggs et al., 2014; Tarasov and Peltier, 2002; Tarasov et al., 2012). **(h)** Greenland temperature reconstruction with $\pm 1\sigma$ shaded (averaged GISP2, NEEM and NGRIP records; Buizert et al., 2014). **(i)** Antarctic δD (EPICA Dome C; Jouzel et al., 2007). **(d–f)** and **(h–i)** are given on the AICC2012 timescale (Veres et al., 2013).

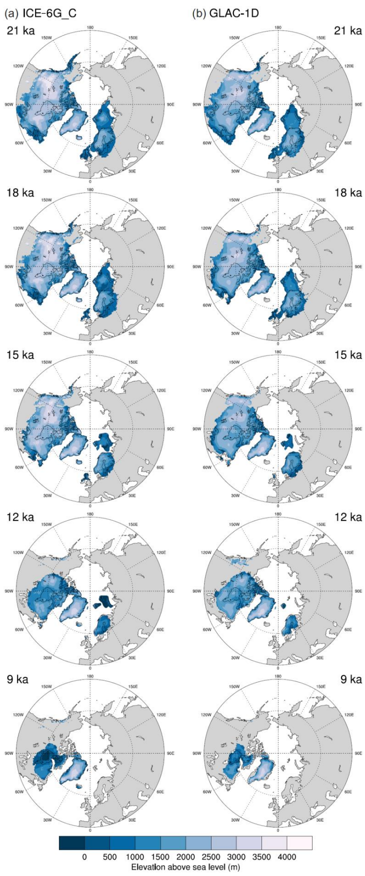

Figure 2. Northern Hemisphere ice sheet elevation at 21, 18, 15, 12 and 9 ka; **(a)** ICE-6G_C reconstruction at 10 arcminute horizontal resolution, elevation is plotted where the fractional ice mask is more than 0.5 (Peltier et al., 2015); **(b)** GLAC-1D reconstruction at 1° horizontal resolution, elevation is plotted where the binary ice mask is one (Briggs et al., 2014; Tarasov and Peltier, 2002; Tarasov et al., 2012).

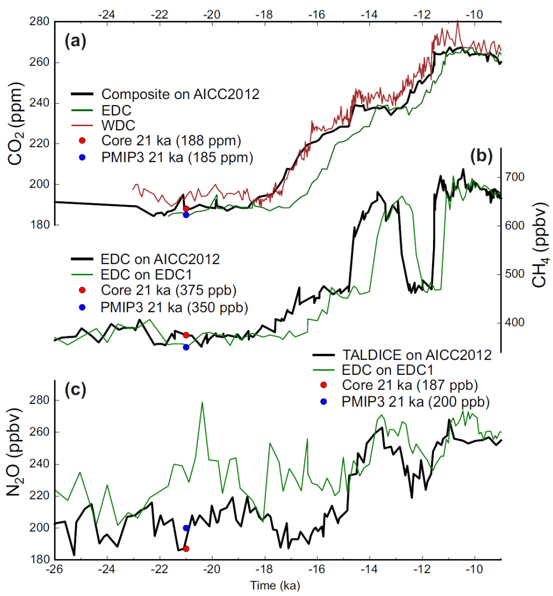

Figure 3. Atmospheric trace gases through the last deglaciation from Antarctic ice cores. **(a)** Carbon dioxide according to a composite record from EPICA Dome C (EDC), Vostok and Taylor Dome (thick black line; Lüthi et al., 2008), adjusted to the AICC2012 chronology (Veres et al., 2013). The original EDC CO_2 record (green line; Monnin et al., 2004) and more recent, higher resolution West Antarctic Ice Sheet Divide (WDC) CO_2 record (dark red line: Marcott et al., 2014) are shown for comparison. **(b)** Methane according to the EPICA Dome C (EDC) record (Loulergue et al., 2008), shown both on the original EDC1 chronology (green line; Spahni et al., 2005) and adjusted to the more recent AICC2012 chronology (thick black line; Veres et al., 2013). **(c)** Nitrous oxide according to the Talos Dome (TALDICE) record (Schilt et al., 2010), adjusted to the AICC2012 chronology (thick black line; Veres et al., 2013). For comparison, the earlier EPICA Dome C (EDC) record on the EDC1 chronology is also shown (green line; Spahni et al., 2005). The nearest measured N_2O concentration to 21 ka is from 21.089 ka; hence the small offset between the slightly earlier concentration (187 ppb) used for the Core and the interpolated value plotted at 21 ka. For **(a–c)** 21 ka concentrations according to the AICC2012 age model (red dots) are shown in contrast to previous PMIP3 LGM concentrations (blue dots; PMIP LGM Working Group, 2010). If using an equilibrium-type spinup for the start of the transient Core simulation at 21 ka (Sect. 2.1.1), use 188 ppm CO_2, 375 ppb CH_4 and 200 ppb N_2O.

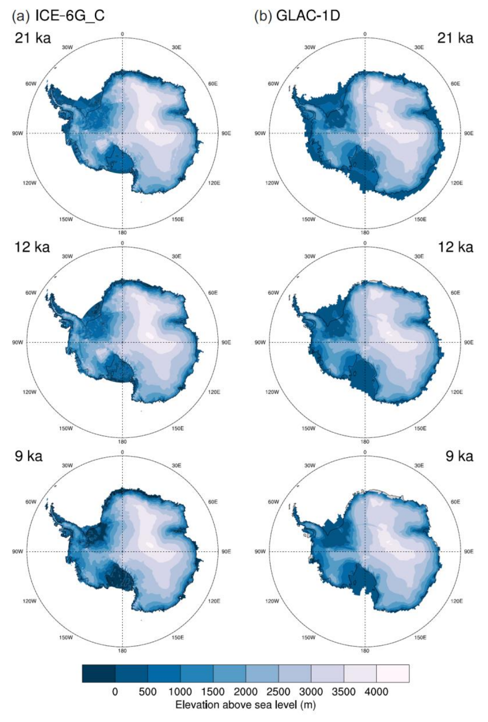

Figure 4. Southern Hemisphere ice sheet elevation at 21, 12 and 9 ka; **(a)** ICE-6G_C reconstruction at 10 arcminute horizontal resolution, ice elevation is plotted where the fractional ice mask is more than 0.5 (Argus et al., 2014; Peltier et al., 2015); **(b)** GLAC-1D reconstruction at 1° horizontal resolution, elevation is plotted where the binary ice mask is 1 (Briggs et al., 2014; Tarasov and Peltier, 2002; Tarasov et al., 2012).