

# MioMIP2: Experimental design for Phase 1 & 2 of the Miocene component of the CMIP7/PMP7 Deep-time Model Intercomparison Project (DeepMIP-Miocene)

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**Abstract.** The Miocene Epoch has emerged as an important paleoclimate model intercomparison target due to its dynamic nature, global warmth, and reconstructed atmospheric CO<sub>2</sub> concentrations. Miocene CO<sub>2</sub> estimates range from as low as 300–600 ppmv to as high as 800–1100 ppmv. The formalized Miocene Model Intercomparison Project (MioMIP2) aims to address questions about the mechanisms driving the climatic changes and tipping points seen in the geological record and improve confidence in the ability of climate models to predict future climates and constrain climate sensitivity. Through alignment with PlioMIP and DeepMIP-Eocene efforts, a coordinated MioMIP (DeepMIP-Miocene) effort will allow for cross-time-period studies that sample the full range of Shared Socioeconomic Pathway paleo analogues. Building on the MioMIP1 “MIP of opportunity”, the Miocene community has been developing the experimental design for MioMIP2. The MioMIP1 effort was an ensemble of opportunity that leveraged existing Miocene simulations run with different models, boundary conditions, and reference datasets. MioMIP2 will be the first formal Miocene MIP with common boundary conditions and will target the Miocene Climatic Optimum (MCO; 16.75–14.5 Ma). We present the experimental design for coordinated “CORE” MCO simulations and sensitivity studies. This community effort synthesizes paleogeography, ice sheet, CO<sub>2</sub>, and vegetation boundary conditions based on the latest available literature and will use geological data from companion proxy synthesis efforts to evaluate the simulations.

## 1 Introduction

There is a long history of using climate models to simulate Earth’s paleoclimates (Kutzbach, 1985).

These scientific studies have been motivated by questions on the drivers of deep-time climate, as well as a desire to assess our confidence in the ability of climate models to predict future climates (Bracegirdle et

al, 2012; Tierney et al., 2020; Burls and Sago, 2022; Zhu et al., 2024), and constrain climate sensitivity (Hoffert and Covey, 1992; Rohling et al., 2012; Sherwood et al., 2020). With the realization that climate models can differ significantly in their response to climate forcings, model intercomparison efforts (MIPs) arose (Gates, 1992; Meehl et al., 1997). Early Paleoclimate Modelling Intercomparison Project (PMIP) efforts focused on the Holocene and the Last Glacial Maximum (Bracconnot et al., 2007), which are relatively recent with low atmospheric CO<sub>2</sub> levels. At the same time an effort developed for investigation of the 'modern analog' for CO<sub>2</sub> levels with close examination of the Pliocene (e.g., Chandler et al., 1994).

125 If climate models are to be applied to climates for comparison to future elevated atmospheric CO<sub>2</sub> scenarios (Riahi et al., 2017), paleoclimate MIPs need to target intervals with high atmospheric CO<sub>2</sub> levels and global mean temperatures. To date, coordinated MIPs have been performed for two such deep-time paleoclimate periods: the Pliocene (PRISM then PMIP) (Chandler et al., 2008; Haywood et al., 2009, 2011, 2013, 2016, 2020) and the Eocene (DeepMIP-Eocene) (Lunt et al., 2012, 2017, 2020). Both efforts have supported a broad range of studies investigating both the response of climate models to changes in paleoclimate forcings and the comparison of the climate simulations to available proxy evidence. The Miocene Epoch (23.03–5.33 Ma) falls between the Eocene and Pliocene epochs, in terms of both age and reconstructed global warmth. Miocene temperatures are estimated to have been 8 to 10°C warmer than pre-industrial (PI) during the peak warmth of the Miocene Climatic Optimum (MCO; 16.75–14.5 Ma), cooling to an estimated 4–6°C warmer than pre-industrial during the Late Miocene (11.63–5.33 Ma) (Burls et al., 2021). This long-term Miocene cooling (order of millions of years) is thought to have been primarily driven by declining atmospheric CO<sub>2</sub> concentrations (Foster et al., 2012a), with other factors such as changes in orbital forcing and slow-climate feedbacks (e.g., ice sheets and vegetation) acting in concert with changes in CO<sub>2</sub> and other radiative forcings to drive substantial climate variability on shorter orbital to millennial timescales (Tierney et al., 2022, Steinhilber et al., 2021b).

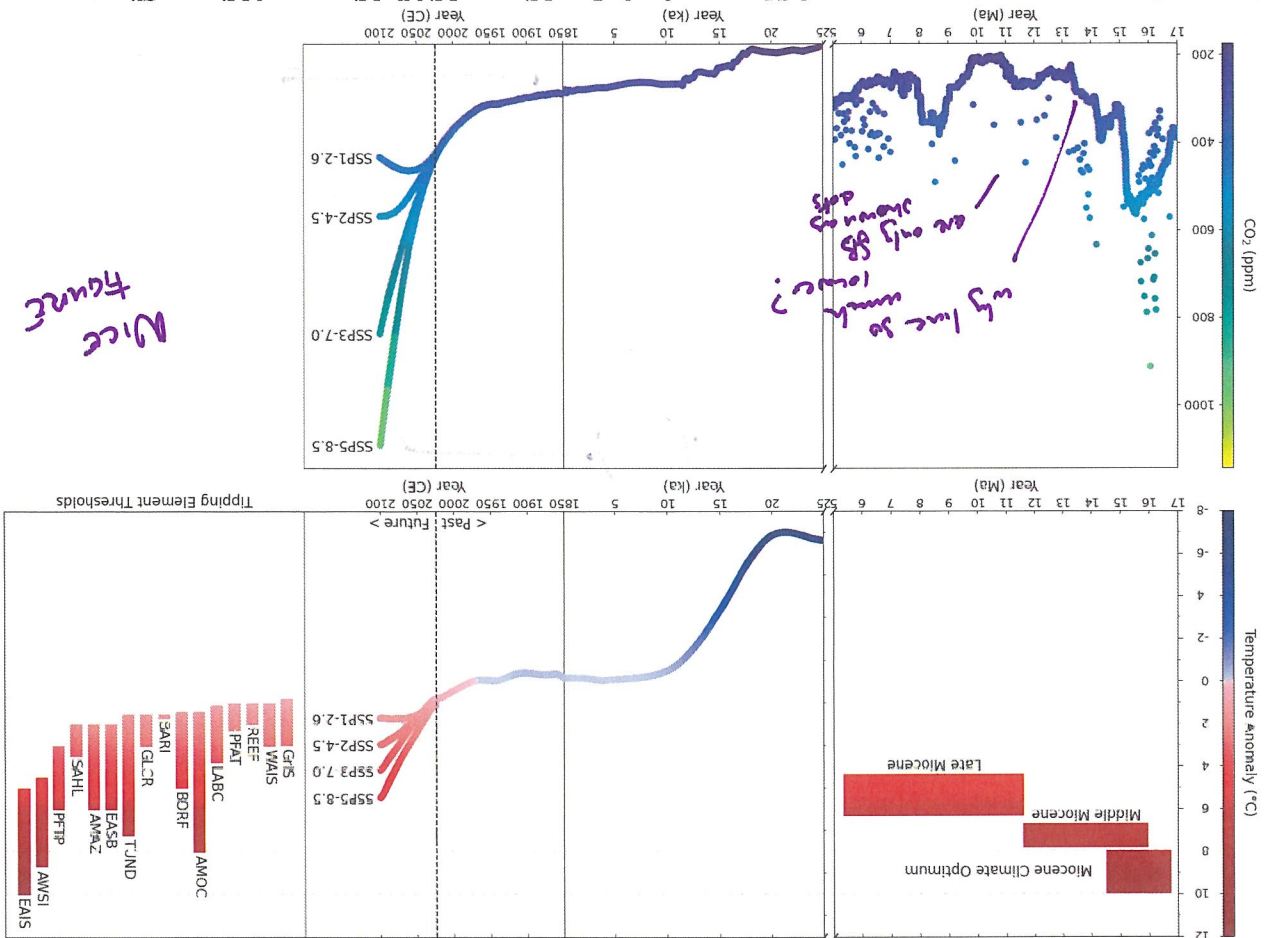


Figure 1: Global mean annual temperature and CO<sub>2</sub> ranges for the Late Miocene, Middle Miocene and Miocene Climate Optimum in relation to the global mean annual temperature and CO<sub>2</sub> timeseries from the Holocene to the end of the century and calculated temperature thresholds for tipping elements. The Miocene global annual mean temperature estimates are taken from Burts et al., (2021). The Last Glacial Maximum to Holocene data are taken from (Osman et al., 2021); the current climate data are taken from HadCRUT5 (Morice et al., 2021), and the future climate projection data are taken from (Nicholls et al., 2021). All temperature timeseries data are the anomaly relative to 1961-1990 and have been smoothed using a rolling mean window of size 100. The Miocene CO<sub>2</sub> estimates are taken from (THE CENOZOIC CO<sub>2</sub> PROXY INTEGRATION PROJECT (CENCO2PIP) CONSORTIUM, 2023) and plotted are the means of the 10,000 year intervals smoothed using a rolling mean window of size 100 (line) and the individual datapoints for the boron proxy data (circles). The Last Glacial Maximum to Holocene CO<sub>2</sub> data are taken from the Antarctic Ice Core 800,000 year record (Berger et al., 2015); the current climate CO<sub>2</sub> data are taken from the Dome ice core record and the NOAA Global Monitoring Laboratory records from Mauna Loa (Dlugokencky and Tans, n.d.; Etheridge, D.M., et al., 2010; Keeling et al., 1976; MacFarling Meure et al., 2006) and the future climate projection CO<sub>2</sub> data are taken from (Riahi et al., 2017). Literature-based temperature threshold estimates for the tipping elements are taken from Armstrong McKay et al., 2022) and the Global Tipping Points Report (2025); GMS=Greenland Ice Sheet (collapse), WAIS=West Antarctic Ice Sheet (collapse), LABC=Labrador-Irminger Seas/Sub-polar gyre (convection (collapse), EASB=East Antarctic Subglacial Basins (collapse), AWSI=Arctic Winter Sea Ice (collapse), AMOC=Atlantic Meridional Overturning Circulation (collapse), AMAZ=Amazon Rainforest (dieback), PFTP=Boreal Permafrost (collapse), EAS=East Antarctic Ice Sheet (collapse), REEF=Low-latitude Coral Reefs (die-off), PFAT=Boreal Permafrost (abrupt thaw), BARI=Barents Sea Ice (abrupt loss), GLCR=Mountain Glaciers (loss), SAHL=Sahel and West African Monsoon (greening), BOREF=Boreal Forest (southern dieback), TUND=Boreal Forest (northern expansion).

165 The Miocene has emerged as an important target for a paleoclimate MIP due to its similarity to near-future projections in terms of global warmth and reconstructed atmospheric CO<sub>2</sub> concentrations. Atmospheric CO<sub>2</sub> estimates for the Miocene range between 300 and 1100 ppmv (Foster et al., 2012b, 2017; Rae et al., 2021a; Soudan et al., 2018; Steinhilber et al., 2021b; Stoll et al., 2019; Tanner et al., 2025; THE CENOZOIC CO<sub>2</sub> PROXY INTEGRATION PROJECT (CENCO2PIP) CONSORTIUM, 2023), covering the full range from the 20th century to the worst case future scenario projected under Shared Socioeconomic Pathways (SSP) 5 (Riahi et al., 2017). Miocene climate modelling studies paired with proxy reconstructions of Miocene climate can provide insights on the sensitivity and tipping points of various elements of Earth's climate system to elevated CO<sub>2</sub> concentrations and global warmth ~~as illustrated in Fig. 1~~. Furthermore, through alignment with PlioMIP and DeepMIP Eocene efforts, a coordinated MIP targeted at the Miocene will allow for cross-time-period studies that sample the full range of paleo analogues to future scenarios in a consistent manner. 175

180 The Buris et al. (2021) MioMIP1 study provided a comprehensive quantitative assessment of published Miocene climate simulations and the range of model-data agreement, highlighting mechanisms common to all Miocene simulations and regions where differences across experiments result in a large spread in warming responses. The MIP of opportunity approach offers certain advantages in that it samples the full range of sensitivity, not only across different climate models but to differing boundary conditions chosen to be representative of the Miocene. Yet this approach comes with its limitations, specifically when the question at hand requires an assessment of the robustness of a result across climate models. Buris et al. (2021) recommends a formalized Miocene MIP (MioMIP2), bringing the community 185 together under the auspices of DeepMIP-Miocene to write this experimental design paper.

This paper is the result of a coordinated community effort in which ~~both~~ <sup>both</sup> experts from the Miocene paleoclimate modelling and data communities have worked to design the outlined boundary conditions and experimental setups. DeepMIP-Miocene's "MioMIP2" effort will target the MIP and consist of 2 phases. Phase 1 uses the Middle Miocene paleogeography, vegetation, and land-ice boundary conditions published and described in Buris et al. (2021), hereafter referred to as "updated Herold". Phase 2,

described herein, uses the state-of-the-art paleogeography, vegetation, and land-ice boundary compiled by consulting experts, and the literature, to incorporate the latest advances in our understanding of conditions during the MCO. In accordance with the previous deep-time PlioMIP and DeepMIP-Eocene efforts, we outline “CORE” experiments that each modelling group participating in MioMIP2 should complete, as well as suggested sensitivity experiments that will benefit from a multi-model approach to testing process uncertainty. Additionally, we hope that the boundary conditions published with this paper will also support a broader range of Miocene sensitivity experiments with individual models that target scientific questions suited to a single model approach. The Miocene community envisions that the simulations conducted as part of the MioMIP2 effort outlined in this paper (as well as future MioMIP efforts targeting other Miocene time intervals) will help address a range of scientific questions.

## 2 The Chosen Interval

Comprehensive investigations of all recognized climatic events throughout the Miocene are envisaged. In the present study, as an initial step, the focus is placed on one of the major climatic events of the Miocene, namely the Miocene Climatic Optimum (MCO; 16.75–14.5 Ma).

## 3 Experimental Design

The DeepMIP-Miocene, “MioMIP2” experimental protocol has 4 CORE experiments and several optional sensitivity experiments (Table 1). The CORE experiments consist of a pre-industrial control simulation (Section 3.1), an abrupt  $4\times\text{CO}_2$  simulation (Section 3.1), a Phase 1 MCO simulation (Section 3.2), and a Phase 2 MCO simulation (Section 3.3). Some modelling groups may choose not to do Phase 1 experiments, in which case they can omit the Phase 1 CORE MCO simulation. Optional sensitivity experiments are described in Section 3.4 and summarized in Table 1 (if the possible sensitivity experiments proposed, modelling groups are asked to prioritize the  $\text{CO}_2$  sensitivity experiments, specifically the  $1\times(1)$  experiment, because this sensitivity study allows the non- $(1)$  forcings to be distinguished from the  $\text{CO}_2$  forcing.

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### 3.1 Pre-industrial Control and Abrupt 4xCO<sub>2</sub> Simulations

As in DeepMIP-Eocene (Lunt et al., 2017, 2026), the pre-industrial simulation should be as close as possible to the picontrol run with 1850 forcings that is associated with the Coupled Model

Intercomparison Project (CMIP) version of the model being used to conduct the MioMIP experiments e.g., Eyring et al., (2015) if a CMIP6 model is being used and Dunne et al., (2025) if a CMIP7 model. As far as is possible, model physics and parameterization schemes should be consistent between the pre-industrial simulation and Miocene experiments unless the Miocene experiment is a sensitivity experiment designed to test the implications of changing a parameterized process such as tidal mixing (see Section 3.3.5).

The picontrol simulation should be initialized from observations or a similar control simulation. The length of simulation and output saved should be the same as the Miocene control simulation, as outlined in Sections 3.3.7 and 3.3.8.

To evaluate each model's equilibrium climate sensitivity under modern boundary conditions, each modelling group is asked to perform the standard CMIP abrupt 4xCO<sub>2</sub> simulation run for at least 150 years, ideally 300 years as this is now part of the new CMIP7 protocol (Dunne et al., 2025).

### 3.2 Phase 1 Middle Miocene Climate Optimum Simulation

MioMIP2 Phase 1 uses the "updated Herold" Middle Miocene paleogeography (Figure 2b), vegetation, and land-ice boundary conditions described in Burts et al. (2021). The starting point for this

paleogeography is the Miocene dataset described in Herold et al. (2008) with the minor modifications of Herold et al. (2011). For a full description of the "updated Herold" dataset see the supporting

information of Burts et al. (2021). The "updated Herold" boundary condition files published in Burts et al. (2021) are available on Zenodo, as outlined in Table 2. Modelling groups participating in MioMIP2 should download these files for their CUBE Phase 1 experiment (deepmip-mio-mco-p2-5xCO<sub>2</sub>), and any of the three Phase 1 sensitivity experiments (deepmip-mio-mco-p1-XxCO<sub>2</sub>) they choose to do

(Table 1). For guidance on how to address the applicable experimental setup choices when running their



deepmip-mio-mco-  
 p2-OpenTethys  
 deepmip-mio-mco-  
 p2-HotSpot

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Corresponding 2-degree version of Phase 2 boundary condition file is available with filename ending in “\_2deg.nc”  
<sup>5</sup> Simulations run with dynamical vegetation should initialize from specified vegetation file  
 # It is recommended that modelling groups use the 2 degree version (filename ending in “\_2deg.nc”) to help ensure consistency in river routes across different model resolutions

### 3.3 Phase 2 Middle Miocene Climate Optimum Simulation

255 Several studies advancing our understanding of the MCO have been published since the “updated Herold” Middle Miocene boundary conditions were created (Burlis et al., 2021). We therefore review the latest literature and compile state-of-the-art paleogeography, vegetation, and land-ice boundary conditions as described in the following two subsections (Sections 3.3.1 & 3.3.2). We refer to these boundary conditions as the “MioMIP2” Phase 2 paleogeography, vegetation, and land-ice boundary conditions (Table 2) which are used to carry out the Phase 2 experiments as summarized in Table 1.

#### 260 3.3.1 Paleogeography and land-sea mask

Since the “updated Herold” paleogeography in Burlis et al. (2021) was compiled, new paleogeographic information and reconstructions for several important regions have been published. Using the “updated Herold” paleogeography (Figure 2b) as our starting point, topographic/bathymetric updates were applied to produce the MioMIP2 Phase 2 boundary conditions (Table 2 & Figure 2c) as described below. Note, we do not make any modifications to large parts of eastern North America, Siberia, Northern Asia, Australia, and New Zealand, and so these regions retain the “updated Herold” reconstruction; see Figure S1 for a comparison of the Phase 1 “updated Herold” and Phase 2 paleogeography. We create our final MioMIP2 Phase 2 paleogeography at a  $0.25^\circ \times 0.25^\circ$  resolution and additionally make a  $2^\circ \times 2^\circ$  version available using conservative regridding (Figure S2).

We update the plate tectonic reconstruction using Zahirovic et al. (2022), but notably depart from the default mantle-based reference frame of this model and use the paleomagnetic reference frame available as part of this model (from Meredith et al. (2021); anchored plate ‘701701’) as the basis for our new paleogeography. This change reflects both an update in the overall plate kinematic reconstruction

While there will always be some difficulties in reconstructing paleobathymetry, especially for regions where oceanic crust has been lost due to subduction and therefore requires synthetic reconstruction, there have been several recent advancements in the reconstruction of regional paleobathymetry in the Cenozoic. In particular, Hochmuth et al. (2020b) provide a detailed reconstruction of Southern Ocean paleobathymetry at key time intervals since 34 Ma (including the Middle Miocene, 14 Ma), determined using backstripping combined with drill core data and seismic stratigraphy. This reconstruction incorporates the Southern Ocean gateways, i.e., the Drake Passage and Tasman Gateway, which are

to this approach and its components have since been explored in Wright et al. (2020).

(2008b), which forms the paleobathymetry in Herold et al. (2008) and its iterations, though refinements et al. (2008b) and Wright et al. (2020). This approach was used in the paleobathymetry of Müller et al. (Sykes, 1996), and a model of oceanic plateau and/or large igneous province emplacement, see Müller e.g., Müller et al., (2008b); Dutkiewicz et al., (2017); Straume et al., (2020), its isostatic correction McKenzie, (2009); Richards et al. (2018), with the addition of a model of predicted sediment thickness, Global paleobathymetry has traditionally been reconstructed based on the oceanic crustal age ('age grids') generated using the underlying plate tectonic model, applying an oceanic thermal subsidence relationship to determine basement depths, e.g., Stein and Stein, (1992) ['GDH1'; Crosby and

**Paleobathymetry**

(Wessel et al., 2019a) or GPately (Mather et al., 2024).

model, using software such as pyGPlates (Mather et al., 2024) and Generic Mapping Tools (GMT) reconstructions are rotated into the same absolute reference frame and/or plate tectonic reconstruction as such, we ensure prior to any paleogeography modifications and incorporations that all into our paleogeography reconstruction were originally published in alternative plate tectonic models; many of the following paleogeography advancements described below and subsequently incorporated tectonic model and reference frame used in DeepMIP-Eocene-Phase 2 (Lunt et al., 2026). We note that Herold et al. (2008) and its derivations, including Burtis et al. (2021), and is consistent with the plate (Müller et al., 2008a) and its mantle-based (hot spot) reference frame that was used as the basis within

thought to have allowed for deep-water exchange during this time (2020b). Hochmuth et al. (2020a) merge their reconstructed paleobathymetry with the recent Antarctic bedrock by Paxman et al. (2019).

I cannot see this difference in Fig 2. In fact Drake Passage looks shallower than today

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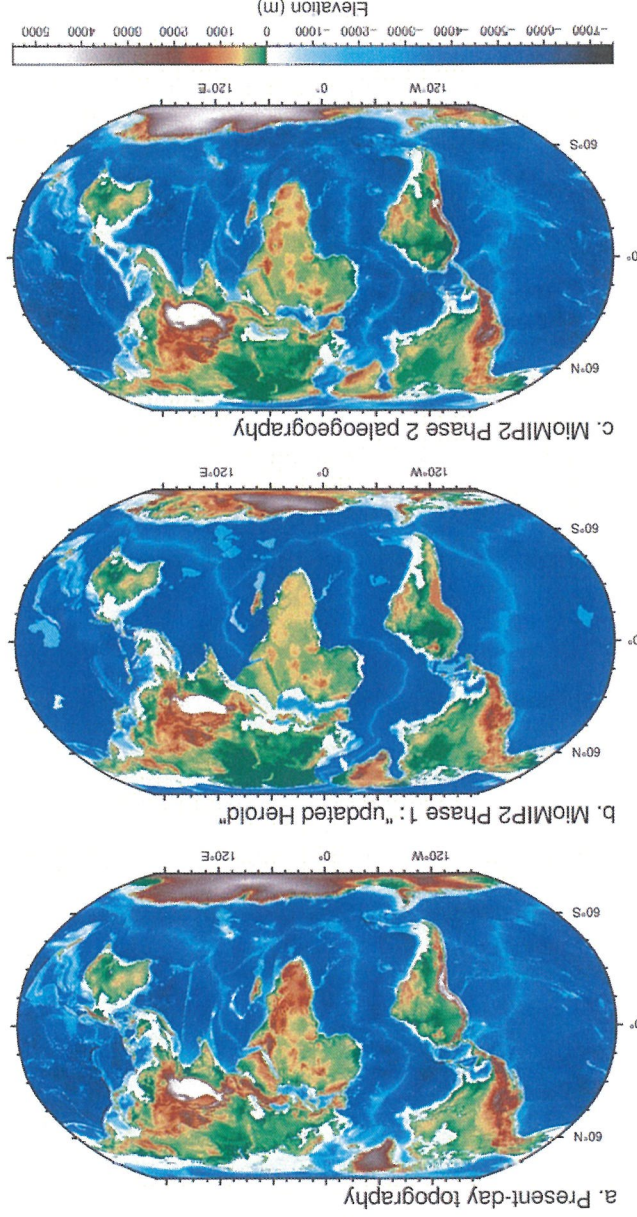


Figure 2: MioMIP2 Paleogeography. a, modern topography; b, MioMIP2 Phase 1 paleogeography ("updated Herold" at 0.5° x 0.5°; c, MioMIP2 Phase 2 paleogeography (this study), at 0.25° x 0.25°. See Supplementary Figure S1 for a comparison of MioMIP2 Phase 2 in its high (0.25°x0.25°) and low (2°x2°) resolution.

In addition, Straume et al. (2020) reconstruct global paleogeography for the Cenozoic, with a specific focus on the evolution of oceanic gateways. While Straume et al. (2020) use a similar method to Herold et al. (2008) globally, they reconstruct and modify key regions using additional geological and geophysical data. Key regions and oceanic gateways include the Fram Strait, Greenland-Scotland Ridge, the Central American Seaway and the Tethys Seaway, with further refinements to the Tethys Seaway, Arabia, and Eurasia made in Straume et al. (2024, 2025). Furthermore, a new method to reconstruct paleobathymetry has been made available using 'pybacktrack v1.4' software (Müller et al., 2018) which allows for both the reconstruction of presently preserved oceanic crust and continental crust, where the former is based on an oceanic subsidence relationship, while the latter is based on its rifting and deformation history, and both methods incorporate the progressive decompaction of sediments (and other user-optional parameters). As this method can only reconstruct presently preserved regions, to derive a global paleobathymetry it still requires the integration of paleobathymetry reconstructed in the more traditional manner for synthetically reconstructed oceanic crust.

There are physiographic uncertainties for several oceanic gateways during the Middle Miocene, including the Fram Strait, Tethys Seaway, Central American Seaway, and Indonesian Seaway. We review these below and conclude this section with a summary of how we combine these recent advances to construct the paleobathymetry and oceanic gateways in MioMIP2 Phase 2 (Figure 2c).

The Fram Strait is a key gateway allowing for exchange between the northern Atlantic and Arctic Oceans and is thought to have had an influence on North Atlantic Deep Water (NADW) and Atlantic Meridional Overturning Circulation (AMOC). In the present climate, it is the only deep-water connection between the Arctic and Atlantic, impacting overturning in the Nordic Seas (Arrhūn et al., 2023). While there is some uncertainty in the specific timing and the depth evolution of the Fram Strait, it is thought to have transitioned from a shallow connection to a deep connection around 17.5 Ma (Jakobsson et al., 2007). Within the "updated Herold" paleogeography, this gateway was a narrow deepwater (~3500 m) connection. Straume et al. (2020, 2024) however, reconstruct the Fram Strait as



340 somewhat shallower (~2500 m) but a wider connection, encompassing more of the surrounding submerged continental shelves. Additionally, the Faroe-Shetland Channel is much less deep in Straume et al. (2024) (~500 m), whereas it is ~4500 m in “updated Herold” (Figure S3).

Long standing debates have persisted around the configuration of the Central American Seaway during the Middle Miocene (Bacon et al., 2015; Jaramillo et al., 2017, 2018; Molnar, 2017; Montes et al., 2012, 2015; O’Dea et al., 2016; Sepulchre et al., 2014). Herold et al. (2008) and “updated Herold” depict the Central American Seaway as a wide and deep (~4500 m) seaway (Figure S4). However, most recent studies suggest that intermediate to deep (~> 2000 m) water connections persisted through the Middle Miocene in some part of the seaway (O’Dea et al., 2016; Vallejo-Hincapié et al., 2024), but that its width might have been restricted by the presence of volcanic arcs (Vallejo-Hincapié et al., 2024) explaining diverging local paleobathymetry evidence (Montes et al., 2012, 2015; Vallejo-Hincapié et al., 2024). Shoaling is suggested to have occurred during the Late Miocene, around 11–9 Ma (Kirillova et al., 2019; Osborne et al., 2014).

The Tethys Seaway was shown as a deep connection (~4200 m) in Herold et al. (2008) and modified to be a shallow seaway (~200 m) in the “updated Herold” (Figure S5). Since then, Straume et al. (2024) has explored the detailed evolution of the Tethys Seaway based on reconstructions from the Darius Programme (Barrier et al., 2018), various geologic and biogeographic indicators for seaway morphology (Bialik et al., 2019; Okay et al., 2010; Segev et al., 2017), and a modified version of the regional kinematic model of van Hinsbergen et al. (2020). In their reconstruction, the Tethys Seaway closed by ~19–20 Ma with the drying of the Mesopotamian Trough and emergence of the *Gophotherium* Landbridge (Bialik et al., 2019; Harzhauser et al., 2007; Segev et al., 2017; Straume et al., 2024). Recent studies on faunal assemblages suggest that any substantial oceanic exchange between the Indian Ocean and Atlantic Ocean ceased by the Early Miocene. At 15 Ma, the Tethys Seaway was in a transitional stage, likely shallow and narrow or completely closed, where evidence suggests there may have been a temporary reflooding of the seaway in the Early and Middle Miocene before its final closure at ~14 Ma (Bialik et al., 2019; Straume et al., 2025). Consequently, a shallower connection

between the proto-Mediterranean and Indo-Pacific may have been operational at 15 Ma, however, no deep connections ( $> 300$  m) were likely at this time (Straume et al., 2024, 2025).

370 The Indonesian Seaway allows water masses from the tropical Pacific Ocean to be transferred into the tropical Indian Ocean (Jordon, 2005) and is thought to have had major impact on Neogene climate evolution (Gallagher et al., 2024; Kuhnt et al., 2004). Paleogeographic evolution of the region during the Miocene is essentially forced by the northward migration of the Australian continent and its collision with Papua New Guinea and South-East Asia (Hall et al., 2012). A recent synthesis by Gallagher et al. (2024) suggests that the seaway was only partially restricted during the Middle Miocene, with constriction occurring later. We note that reconstructing the paleogeography of such a complex seaway is far from straightforward, high uncertainty persists due to active tectonic deformation, the presence of many small islands and narrow straits and a lack of direct quantitative paleobathymetric evidence. Frigola et al. (2018) made some modifications to the Herold et al. (2008) Middle Miocene reconstruction, including an alternative reconstruction for this region, based on the more recent work from Hall (2012). However, as Hall (2012) does not provide any absolute elevation values, Frigola et al. (2018) assigned average elevation and bathymetry values to Hall's (2012) geographic feature categories and modified the depth of the gateway so that intermediate depth passages exist at depths between 600 and 1000 m.

385 Here, we combine many of these recent advances in paleobathymetry and oceanic gateways configurations. We create our paleobathymetry using the plate tectonic model of Zahirovic et al. (2022), based on the combination of pybacktrack 1.4, for presently preserved regions and submerged continental crust, and we use the workflow described in Wright et al. (2020) for regions requiring synthetically reconstructed oceanic crust. We use the Stein and Stein oceanic subsidence relationship within both these workflows. We incorporate the Toclum et al. (2020a) reconstruction for the Southern Ocean, and Straume et al. (2020) for the Iram Strait (Figure S3), Central American Seaway (Figure S4), and Tethys Seaway and surrounds (including the Paratethys; Figure S5). We additionally adopt the reconstruction from Frigola et al. (2018) for the Maritime Continent region and Indonesian