

RECONSTRUCTION OF THE FOREBULGE IN THE NORTH SEA REGION DURING THE LAST GLACIAL PERIOD THROUGH GLACIAL ISOSTATIC ADJUSTMENT MODELLING

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1. INTRODUCTION

The process of growth and retreat of Quaternary ice sheets has a major impact on landscape evolution as the large-scale storage of ocean water as continental ice lowers the sea level significantly. According to Clark and Mix (2002), the eustatic sea level during the Last Glacial Maximum (LGM) would have been 120 to 135 m lower than today. However, these continental ice sheets also alter the topography as the lithosphere becomes subject to glacial isostatic adjustment (GIA); the weight of the ice sheets causes deflection of the lithosphere which subsides into the asthenosphere. At the peripheries of these ice sheets, a small region of uplift appears; the so-called peripheral bulge or forebulge (Whitehouse, 2018). This process of GIA is not instantaneous, only after a certain period of time the cryosphere-geosphere system will more or less have reached a state of equilibrium. This is only true assuming that the cryosphere has not evolved during this period of time. Even today the Earth system is still adapting to the continental ice loss since the LGM (Schumacher et al., 2018), meaning that deglaciated areas that contained ice sheets during the LGM still experience lithospheric uplift today. Likewise, areas of peripheral bulges during the previous glaciation, such as the forebulge in the North Sea related to the European ice sheets, are still collapsing nowadays (Kiden et al., 2002). As GIA influences for example shoreline migration (Lambeck, 1995) and induces relative sea-level changes, a comprehensive knowledge of the mechanism is crucial for understanding current glaciological and sedimentological processes. Moreover, understanding present-day GIA resulting from the last deglaciation is crucial to be able to quantify the amount of sea-level rise related to anthropogenically caused global warming.

This thesis focusses on the reconstruction of the forebulge in the southern North Sea resulting from GIA related to the British-Irish ice sheet (BIIS) and the Scandinavian/Fennoscandian ice sheet (FIS). These reconstructions are achieved by GIA modelling. To model the glacial isostatic adjustment resulting from an ice sheet, not only data of the ice sheet extent is needed, but also its thickness. Thickness reconstructions for the Eurasian ice sheets are reliable for the last glacial period as data becomes very scarce when going further back in time. As reconstructing ice sheet thicknesses before the Last Interglacial (LIG) lies beyond the scope of this thesis, the North Sea forebulge reconstructions focus on the last glacial cycle. Different ice sheet reconstructions are used as input into a rather simple, but still quite accurate GIA model considering the earth as a relaxing asthenosphere overlain by an elastic lithosphere. The forebulge reconstructions resulting from these different ice sheet models are compared to evaluate the influence of ice model differences. Moreover, the sensitivity of forebulge reconstructions to the choice of earth parameters such as lithosphere thickness and viscosity of the mantle is assessed as well.

As there is some evidence for an ice-dammed proglacial lake in the North Sea north of the Netherlands during the LGM (Hjelstuen et al., 2018; Patton et al., 2017), reconstruction of the forebulge in this area might be crucial to verify whether this peripheral bulge might have been able to act like a dam bordering this potential proglacial lake. Furthermore, there is evidence for the forebulge having had an influence on valley incisions and rivers deflecting on this forebulge, during the last glacial period (Busschers et al., 2007) but also during the penultimate glaciation (De Clercq et al., 2018), indicating the importance of peripheral bulges on landscape evolution. Finally, understanding the current uplift in previously glaciated regions and the coupled current collapse at locations of former bulges will become easier as evaluations of GIA-related processes in the past become more robust.

Some earlier attempts have been made to reconstruct the location and/or magnitude of the forebulge in the southern North Sea. Fjeldskaar (1994) modelled the decay of the forebulge caused by the Eurasian ice sheets since the LGM and observed no true lateral migration of this bulge. In the region of interest for this thesis, the forebulge would have contributed to an uplift of 10 to 20 m. Based on post-glacial rebound processes, Kiden et al. (2002) predicted the forebulge to be positioned at a strip surrounding the FIS and passing through the northern part of the Netherlands. This reconstruction, however, does not take into account the BIIS. Their result is, however, comparable to reconstructions from Busschers et al. (2007) at 21 ka based on data from the ICE-5G (VM2) model (Peltier, 2004) and on data from Steffen (2006), locating the zone of maximal uplift related to the FIS and BIIS in the northern Netherlands.

It can be expected that the collapse of the forebulge as a result of decaying ice sheets does not result in an extensive lateral migration of this forebulge, as its elevation will simply decrease over time. However, during a period of ice sheet growth the forebulge is likely forced to move away from the expanding ice sheet. Moreover, its geometry is expected to change when besides the Fennoscandian ice sheet also the British-Irish ice sheet starts to expand, and the coalescence of these two ice sheets will most likely alter the geometry of the forebulge again. Based on earlier predictions, the forebulge during the LGM can be expected to be located at the northern part of the Netherlands, but this is not necessarily the case for the complete last glacial period.

2. GEOLOGICAL AND ENVIRONMENTAL SETTING

2.1 North Sea basin

The current North Sea basin is an epicontinental shelf sea almost completely surrounded by the land of seven countries; the UK, France, Belgium, the Netherlands, Germany, Denmark and Norway (Cohen et al., 2017). This shallow sea is connected with the Atlantic Ocean via the English Channel in the south and with the Norwegian Sea in the north, spans 970 km from north to south and 580 km from east to west and comprises an area of 750.000 km² (Fig. 1)(Gibbard and Lewin, 2016). The basin deepens towards the north and north-east. The Southern Bight situated between Belgium and the Netherlands on one side and southeast England on the other side is generally less than 40 m deep, while the depth in the central part of the North Sea increases up to 100 m (Cohen et al., 2017). Between Norway and Scotland the shelf deepens to 200 m. South of Norway a bathymetric depression of depths reaching 700 m is present; the Norwegian Channel (Cohen et al., 2017; Huthnance, 1991). The Oslo fjord, of which the Norwegian Channel is in fact a continuation, represents the pathway of an extensive ice stream during the latest glacial periods (Cohen et al., 2017).

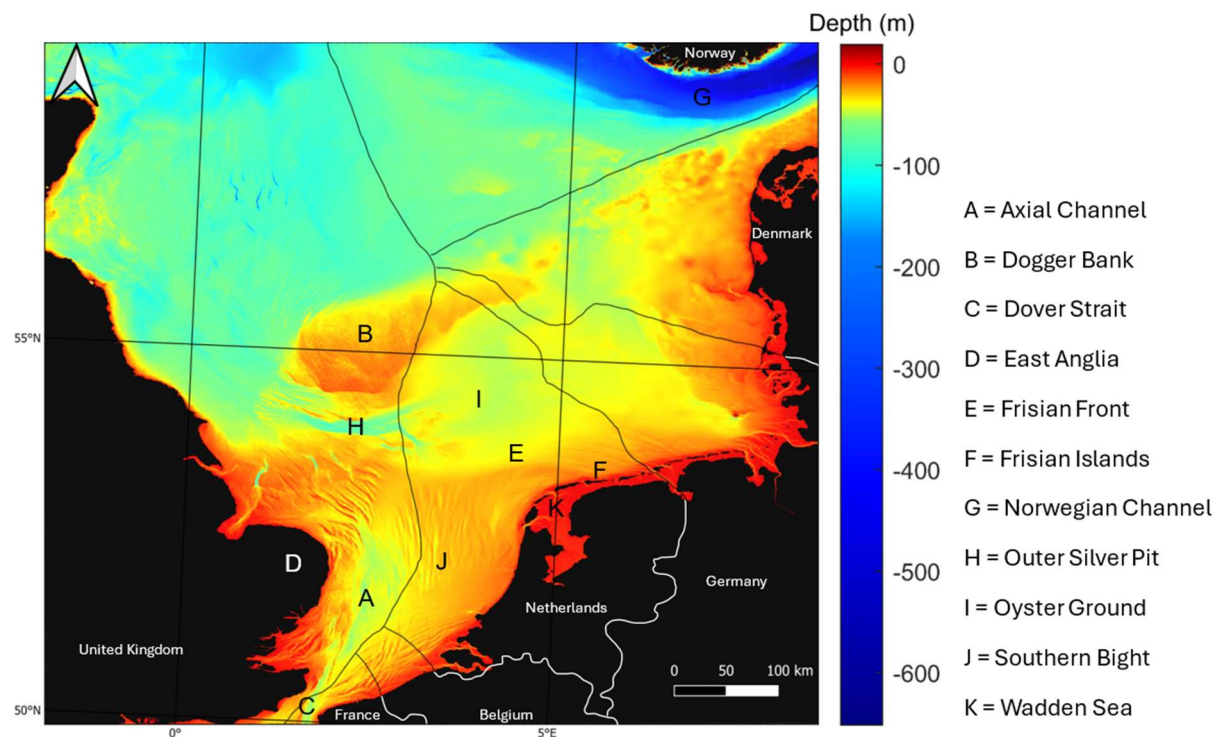


Figure 1: Bathymetry of the North Sea Basin. Data derived from EMODnet Map Viewer: <https://emodnet.ec.europa.eu/geoviewer/>.

The southern North Sea is underlain by a basement consisting of lower Paleozoic sediments that underwent low-grade metamorphism as a result of the Caledonian orogeny. These rocks are locally overlain by alluvial Devonian deposits, but are at least covered by several kilometers of upper Paleozoic, Mesozoic and Cenozoic sediments (Cameron et al., 1993). Several areas that started uplifting at the start of the Carboniferous delimited the North Sea basin until the Late Triassic. These areas include the London-Brabant Massif in the south and the Mid North Sea High and Ringkøbing Fyn High in the north.

Crustal stretching and thermal subsidence provided accommodation space for the deposition of a 3500 m thick sequence of post-Triassic sediments (Cameron et al., 1993). This subsiding rift basin is aligned SE-NW from onshore regions such as the Netherlands towards the offshore direction of southern Norway (Cohen et al., 2017). The rift system was reactivated during the transition between the Eocene and the Oligocene. This reactivation is part of a rather intense basin-inversion pulse attributed to the Pyrenean phase of the Alpine orogeny (de Lugt et al., 2003; Van Wijhe, 1987). After tectonic reactivation of the rift system sedimentation in the basin was largely influenced by delta-systems entering the North Sea from the east (Cameron et al., 1993). One of the most prominent of these deposits is the southern North Sea delta, which is responsible for the predominant part of early to middle Pleistocene deposits in the southern and central parts of the North Sea not originating from glacial activities (Cameron et al., 1992; Graham et al., 2011).

The same tectonic situation was dominant during the Quaternary, allowing for several hundred meters of sediments to be deposited in the depocenter, while the peripheries of the basin, for example in the southernmost parts, experienced slight uplift (Cameron et al., 1993; Cohen et al., 2017). Rivers such as the Rhine and Thames, but also the Eridanos were the main source of Quaternary sediment supply. The latter was a river system located within the present-day Baltic Sea with its catchment located in the Baltic area and Fennoscandia. This river supplied 80 % of the sediments reaching the North Sea, until the system collapsed with the formation of larger Fennoscandian ice sheets around 1 Ma (Cameron et al., 1992; Cohen et al., 2017; Overeem et al., 2001). Concurrently, the Rhine started to develop an extensive delta within the central and southern North Sea (Gibbard and Lewin, 2016).

The bathymetry of the current North Sea is relatively flat, except for some plateau-like features such as the Dogger Bank. The principal type of seabed sediment is sand. Along the SE coast of England large estuaries are the witnesses of ancient valleys that became drowned due to the postglacial sea level rise (Cohen et al., 2017). Similar flooded valleys can be found at the other side of the Southern Bight, such as the Rhine-Meuse valley (Busschers et al., 2007). Modern estuaries of the Scheldt, Meuse and Rhine rivers interrupt the current coastline of the Netherlands, while more to the north the Wadden Sea lagoon of the Netherlands, Germany and southern Denmark is protected by the barrier islands referred to as the Frisian Islands (Cohen et al., 2017).

2.2 Glaciations

2.2.1 Quaternary glaciations

The Quaternary represents the final phase of a long-term Cenozoic cooling trend starting around 49 million years ago during the Eocene (de Boer and van de Wal, 2021). The first Cenozoic ice sheet on Antarctica was established at the Eocene-Oligocene transition around 34 Ma and approximated its present-day shape around 10 Ma. The onset of Northern Hemisphere glaciations around 2.8 Ma and the subsequent start of glacial-interglacial cycles ushered in the Quaternary at 2.6 Ma (de Boer and van de Wal, 2021). During interglacials, such as the current Holocene, only Greenland and Antarctica contain ice sheets, while these expand during glacial periods and also cover large parts of North America and Eurasia. This regularity can be linked to Milankovitch cyclicities caused by changes in Earth's orbit around the sun. Variations in the Earth's precession, obliquity and eccentricity, with respective periodicities of 19 to 23 kyr, 41 kyr and ~100 kyr cause variations in the amount of solar energy that reaches the Earth's surface, thus influencing the global climate (Imbrie et al., 1992). During the Middle Pleistocene Transition (MPT) the duration of glacial-interglacial cycles switched from a lower-amplitude, higher-frequency variability characterised by a 41 kyr cyclicity to a higher-amplitude, lower-frequency variability with a cyclicity of more or less 100 kyr (Clark et al., 2024; Pisias and Moore, 1981). This switch happened between 1.2 and 0.8 Ma (Ehlers et al., 2018) and was accompanied by an intensification of the long-term cooling (Clark et al., 2024). Before this intensification, eustatic sea level lowstands only were up to 70 m below current sea levels, while they lowered up to 150 m during some glaciations after the MPT (Cohen et al., 2014). The increase in amplitude of the more recent glaciations is associated with an increase in river valley incision during periods dominated by a cold climate, and with sediment volumes transported by the Rhine and Meuse rivers becoming more important (Gibbard and Lewin, 2016).

Within the Quaternary, Marine Isotope Stages (MIS) are often used when referring to glacial periods and interglacials, as they are not locally defined and can thus be used globally. Odd stages refer to interglacials or interstadials (relatively warm periods within a glacial), while even stages refer to glacials or stadials (cold periods within a glacial). MIS 1 refers to the Holocene, while MIS 5e refers to the LIG or Eemian. MIS 2 represents the LGM while MIS 6 equals the Penultimate Glacial Period. MIS 2-5d are often referred to as the Last Glacial Period, which is called Weichselian in northern Europe or Devensian in Great-Britain (Graham et al., 2011).

There is evidence for at least three major glaciations with ice sheets expanding into the North Sea during the last 0.5 Myr (during MIS 12, MIS 6 and MIS 2), but quite recently evidence of imprints of older glaciations within the North Sea sediments has been discovered as well (Graham et al., 2011). During MIS 12, the North Sea basin was most likely almost completely covered by ice (Fig. 2), while post-Elsterian bedform geomorphology and sediments suggest more than only three major episodes of glacial advance and retreat (Graham et al., 2011). Up until around 0.5 Ma, Britain was connected to mainland Europe even during sea level highstands via the Weald-Artois anticline (Fig. 2). Because of this ridge, rivers draining areas in southern England and northern France - such as the paleo-Solent, Somme and Seine - entered the Atlantic Ocean between Cornwall and Brittany via the Channel River. Contrarily, more northerly located rivers such as Thames, Scheldt, Meuse and Rhine continued in a northward direction and discharged into the North Sea (Gibbard and Cohen, 2015).

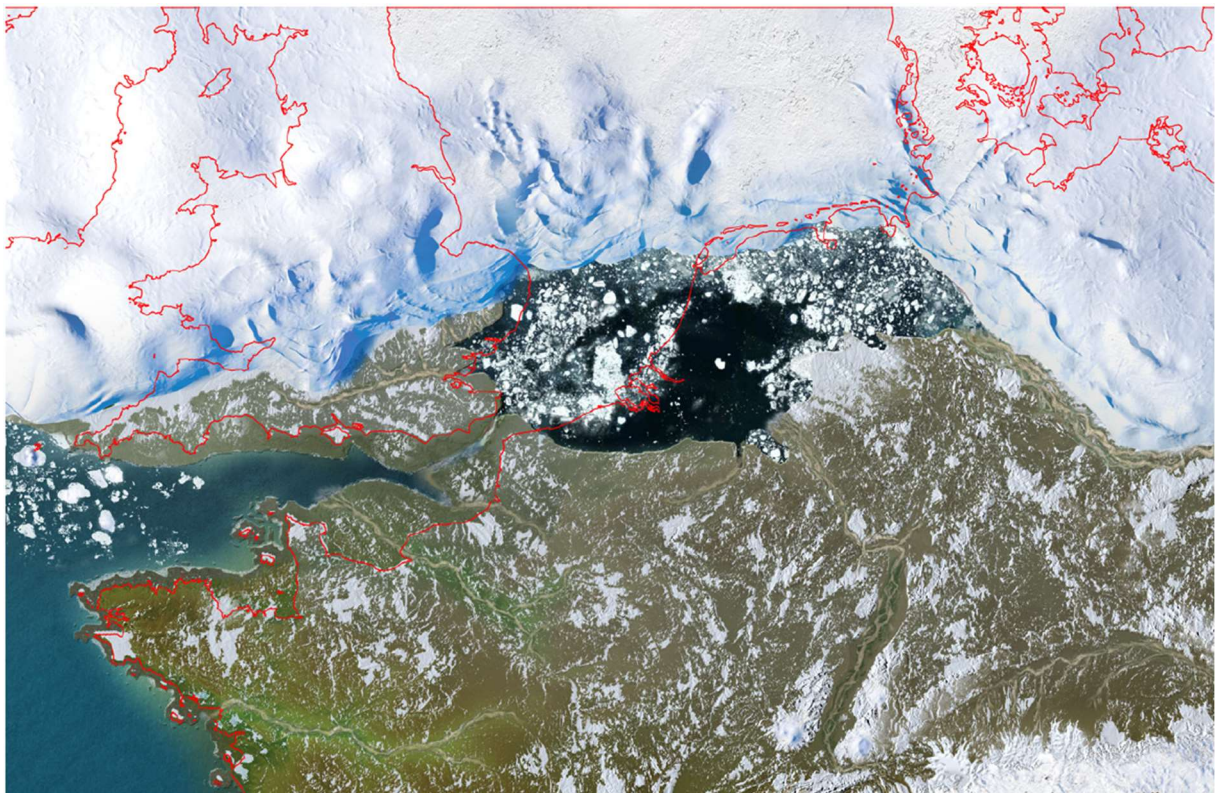


Figure 2: Reconstruction of the situation in the southern North Sea during the Elsterian/MIS 12, with ice sheets covering the largest part of the North Sea basin. A proglacial lake is enlocked between the ice sheets in the north and landmasses in the south, including the Weald-Artois anticline in the southwest (Amkreutz et al., 2022).

It has been hypothesized that during certain periods of maximum ice extent proglacial lakes must have developed in the southern North Sea basin, and glacio-lacustrine sediments that have been described in East Anglia and the Netherlands seemingly confirm this proglacial lake hypothesis (Gibbard, 1995). A proglacial lake would have existed during MIS 12 (Fig. 2, Fig. 3A) and during MIS 6 (Fig. 3B) (Murton and Murton, 2012). There is, however, still a lot of discussion on the lake limits, emphasising the importance of more sedimentological studies and modelling approaches that could settle the discussion. The proglacial lakes became gradually filled with water as ice sheets north of them prevented any outflow while the northward flowing European rivers continued to deliver water to them (Gibbard and

Cohen, 2015). The blockage of one of these proglacial lakes during MIS 12 by the Weald-Artois ridge at its southern border resulted in the overtopping, breaching and erosion of this barrier, which mostly consisted of chalk. As a result of erosion, the saddle migrated towards the north until it was no longer covered by chalk but by Eocene deposits such as the London Clay. The breaching process continued after formation of another ice-dammed lake during MIS 6 via degradation of the easier erodible London clay, and this caused the final opening of the Dover Strait (Gibbard and Cohen, 2015). Gibbard (2007) suggested that the more northward located MIS6 ridge could have been a bedrock outcrop or the remnant of a previously deposited moraine. Evidence for catastrophic proglacial lake drainage via the English Channel was already provided by Gupta et al. (2007) based on bathymetric studies. Gupta et al. (2017) further examined landforms present in the Dover Strait, such as features indicative of lake overspill and plunge pools resulting from waterfall erosion, and provided evidence supporting the two-stage opening hypothesis. The final dam breaching would have resulted in the connection of the English Channel to the North Sea, altering the courses of rivers such as Thames and Scheldt towards the Channel River. The Meuse and Rhine only followed the same evolution since the last interglacial (Gibbard and Cohen, 2015).

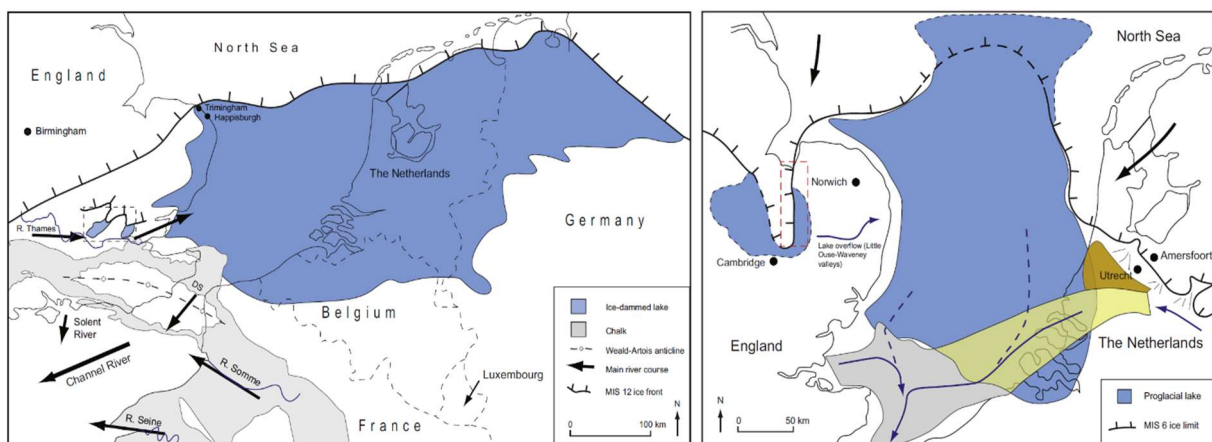


Figure 3: Reconstruction of proglacial North Sea Lake **A)** during MIS 12 and **B)** during MIS 6 (Murton and Murton, 2012).

De Clercq et al. (2018) provided morphological evidence suggesting that the presence of a forebulge at the end of MIS 6 may have affected the evolution of the Scheldt Paleovalley in the northwestern part of Belgium. GIA-related responses to the penultimate glaciation caused uplift on top of the already existing ridge situated in between Belgium and East Anglia and forming a barrier between the Dover Strait and a proglacial lake more to the north. The longer sea level rise in Belgium compared to the Netherlands suggests that Belgium might have been situated on top of the region of maximal uplift (De Clercq et al., 2018). Forebulge upwarping, drainage of the proglacial lake and subsequent bulge relaxation caused tilting of the river valley, as the northern part of the valley experienced more uplift than the more southerly located region. This resulted in a deflection of the paleo-Scheldt river north of Ghent towards the west, in the direction of Zeebrugge and Ostend. Subsequent valley incisions up to 20 m were filled in with Eemian sediments quite fast afterwards (De Clercq et al., 2018).

2.2.2 Last glacial cycle

During the previous interglacial, the coastal configuration in the southern North Sea was similar to that of today, with Britain being separated from the rest of Europe for the first time (Gibbard and Cohen, 2015). The sea level during the Eemian highstand was 4 to 9 m lower than during the Holocene. Moreover, the subsequent climatic evolution during the last glacial cycle was not at all characterised by a steady cooling from MIS 5e until the LGM, but there were large fluctuations in both global temperatures and ice volumes (de Boer and van de Wal, 2021). After the last interglacial, sea level gradually decreased before reaching several short cold periods at around 100 ka (Gibbard and Cohen, 2015). Eustatic sea levels decreased from -25 m during MIS 5a (85 ka) to -90 m during MIS 4 (70 ka) (Fig. 4) (Cohen et al., 2017). The LGM sea level lowstand of -120 m (Gibbard and Cohen, 2015) was followed by a fast termination within 10 kyr (de Boer and van de Wal, 2021). As can be seen on Fig. 5, inundation

of the North Sea after the LGM proceeded roughly from north to south (Amkreutz et al., 2022), which can be seen as a direct result of the general bathymetry of the North Sea basin. The Younger Dryas cold spell decelerated the flooding of the basin (Fig. 5), but this short period only caused a temporary delay of the rapid sea level rise accompanying the early part of the Holocene (Amkreutz et al., 2022).

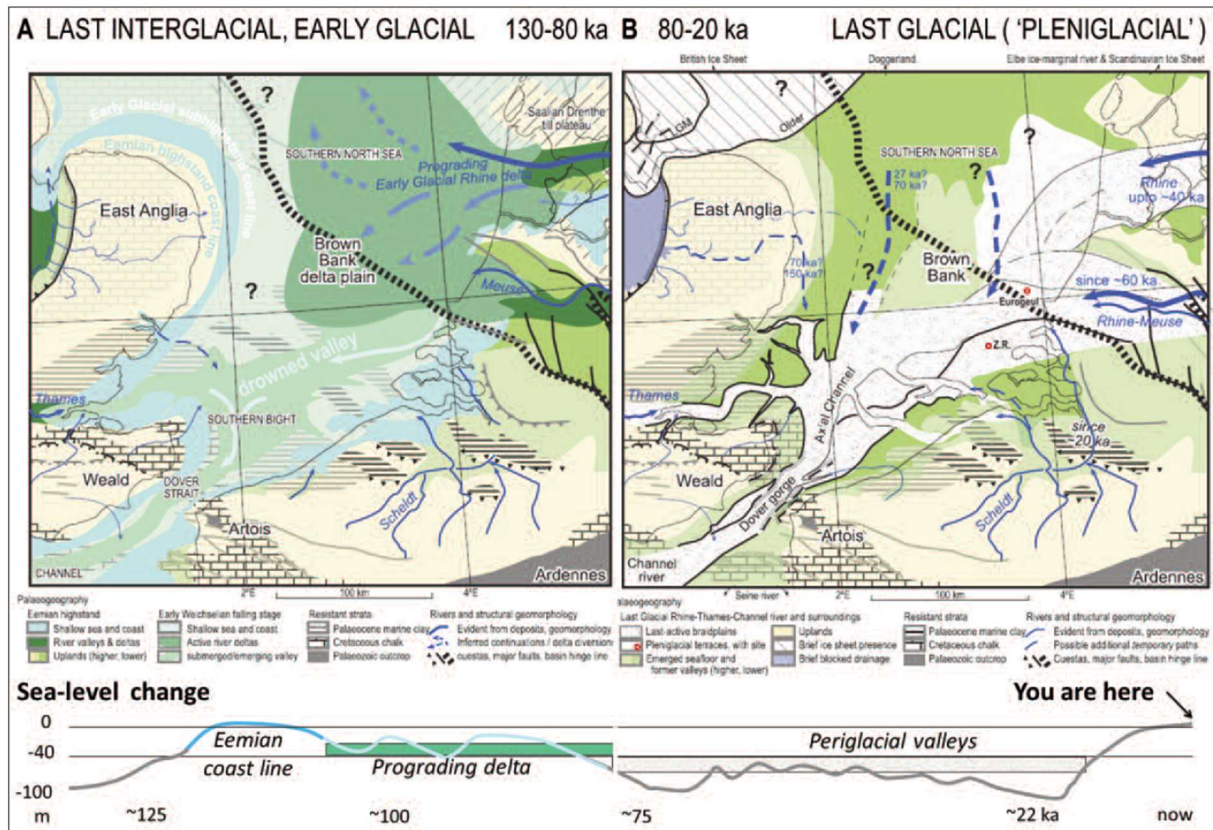
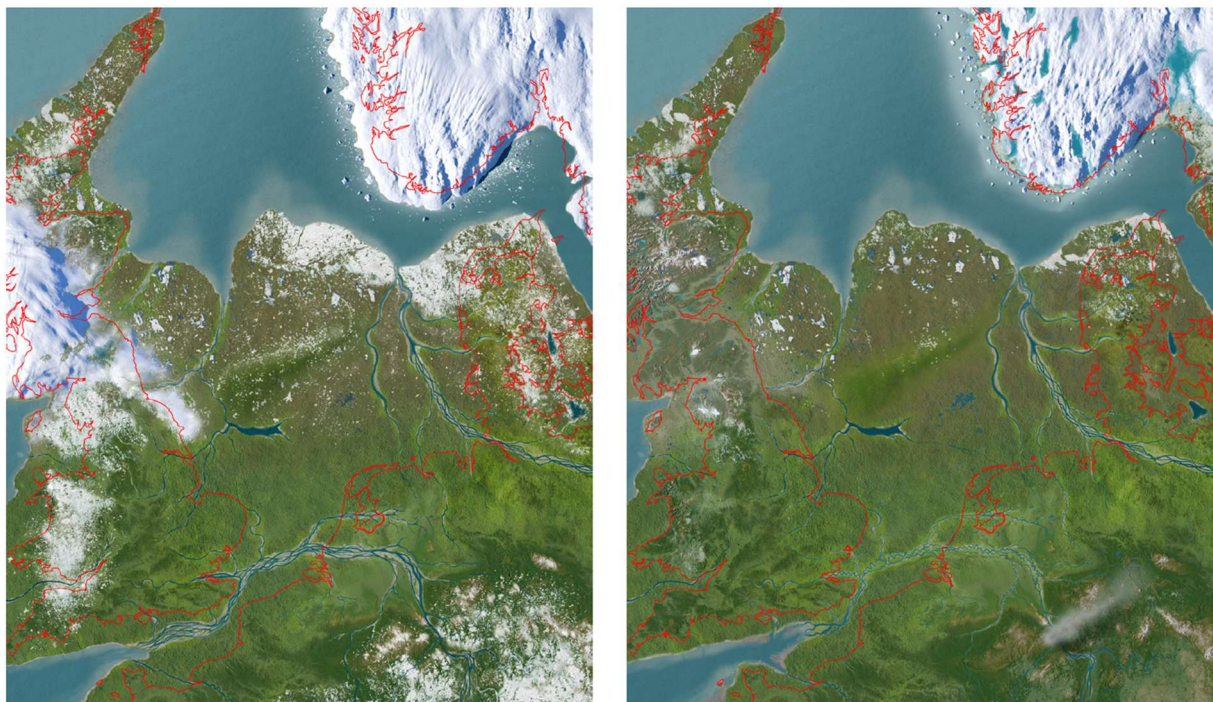


Figure 4: Paleogeography of the southern North Sea basin, **above: A)** at the start of the last glacial cycle and **B)** during later stages of the last glacial cycle, **below:** sea level curve for the last glacial cycle (Paleogeographical maps originally from Hijma et al. (2012), sea level curve added in Gibbard and Cohen (2015)).



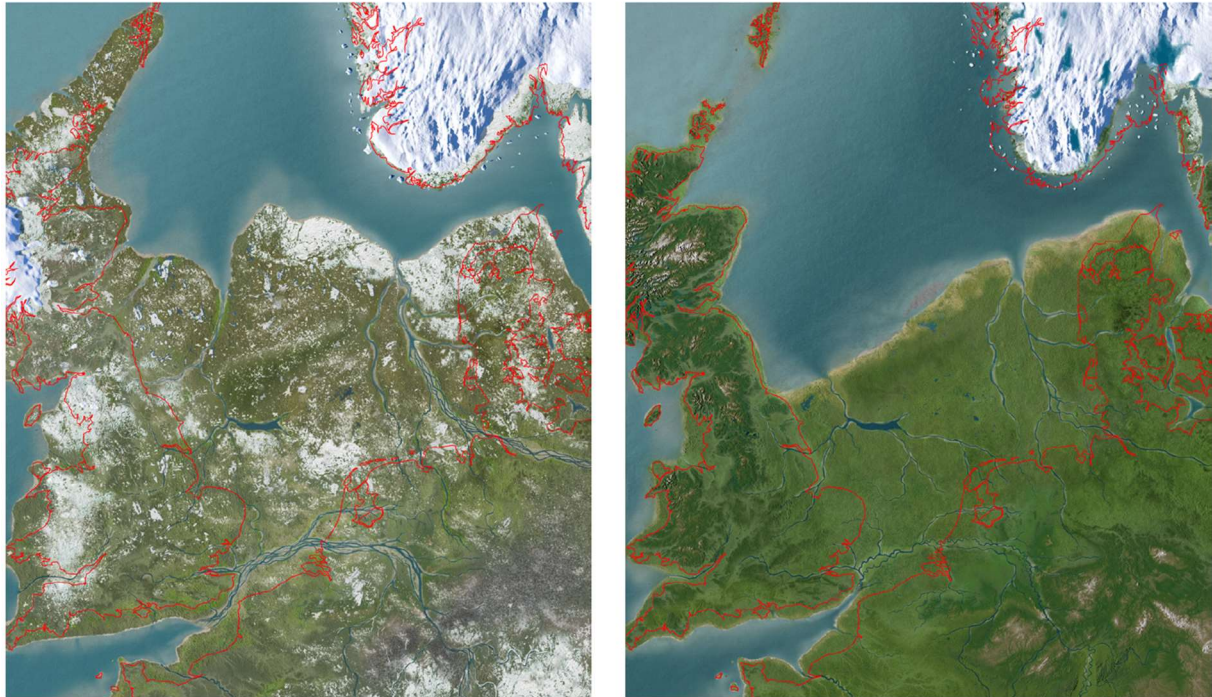


Figure 5: Reconstructions of the geographic and climatic situation of the North Sea basin at the end of the last glacial period at **A)** 14 ka, **B)** 13 ka, **C)** 12 ka and **D)** 11 ka (Amkreutz et al., 2022).

It is generally assumed that the LGM was reached at 21 ka, but this depends very much on the location and the used record. A minimum in CO₂ levels was reached at 20 ka, while at 22 ka the sea level already achieved its minimum and ice sheet models tend to date the LGM at 18 ka (de Boer and van de Wal, 2021). Within this glacial period, Heinrich events and Dansgaard-Oeschger events occurred. These are intervals of climate variability of several thousand years or hundreds to thousands of years respectively, related to ice discharge in the North Atlantic Ocean (de Boer and van de Wal, 2021). The end of the LGM is most likely linked to Heinrich event 2, which is characterised by a massive discharge of icebergs in the North Atlantic Ocean related to the collapse of the Laurentide and other North Atlantic ice sheets (Ehlers et al., 2018).

Not all ice sheets reached their maximum extent during MIS 2. While the Laurentide ice sheet was maximal during the LGM, the Cordilleran Ice Sheet, but also the Barents-Kara Ice Sheet and the western part of the Fennoscandian Ice Sheet were the most extensive during MIS 4 (Ehlers et al., 2018). The Eurasian ice volume during MIS 4 is indeed almost as large as during the LGM. These results are consistent with findings from Hughes et al. (2013) and with sediments in the North Sea, providing evidence for two phases of wide-ranging ice sheets within the last glacial period; one phase during MIS 4 and the other one during MIS 3/2 (Carr et al., 2006; Graham et al., 2011). According to Clark et al. (2012), the maximum ice sheet extent was reached at 27 ka in the largest parts of Great-Britain, Ireland and the North Sea. However, the BIIS is much smaller in volume than the FIS, which is one of the largest ice masses during the last glacial period and even merged with the Barents-Kara Ice Sheet and the British-Irish Ice Sheet (Graham et al., 2011; Hughes et al., 2013; Svendsen et al., 2004). The resulting ice sheet complex was the third largest ice mass during the LGM, only surpassed by the Antarctic and North American ice sheets and with a volume three times that of the current Greenland Ice Sheet contributing to approximately 20 m of eustatic sea level decrease (Hughes et al., 2016; Patton et al., 2016).

Hjelstuen et al. (2018) found evidence for drainage of an ice-dammed lake that would have existed when the BIIS and FIS were merged in the North Sea during the LGM. It was already suggested by Sejrup et al. (2016) that retreat of the two ice sheets after the LGM would have created a corridor for drainage of

the lake towards the north. The glacial lake outburst flood (GLOF) would have advanced towards the Ling Bank Drainage Channel and further in the direction of the Norwegian Channel. Hjelstuen et al. (2018) provided evidence for this theory by using sedimentological interpretations and interpretations of seabed features based on geophysical data to investigate the drainage route of the GLOF. The proglacial Late Weichselian North Sea Lake would have been situated south of the Dogger Bank (Fig. 6) (Hjelstuen et al., 2018) and would have had an estimated volume of 3900 km³ (Bigg et al., 2012). The northeastern part of the Dogger Bank is interpreted as the region of breaching, which is inferred to have happened at 18.7 ka when the sea level was at least 110 m lower than today. A 10 m thick sedimentary sequence 300 km downstream of this region, at the northern tip of the Ling Bank, is interpreted to represent the deposits resulting from this ice-dammed lake outburst (Hjelstuen et al., 2018).

2.3 Postglacial forebulge collapse

While ice sheet growth causes isostatic subsidence below the ice sheets and a zone of peripheral bulge formation surrounding the region of ice growth, the decay of ice sheets should result in the opposite effect, namely the uplift of areas once covered by ice sheets and the subsidence of the former forebulge. Several attempts have been made to localise the region of peripheral bulge collapse and to quantify the amount of subsidence. Based on the difference in vertical motion between Belgium and the Netherlands, derived from differences in rates of post-glacial relative sea-level rise, Kiden et al. (2002) suggest the Netherlands to be located closer to the former forebulge compared to Belgium. They argued that the difference in movements is mainly caused by the FIS, as other ice sheets have too small ice volumes (such as the BIIS) or are located too far away from both countries to have a significant impact. Belgium would possibly be located on the southern slope of the peripheral bulge, while there is no need to assume that the collapse of the forebulge would have been accompanied by a lateral (i.e. northward) migration of this bulge (Kiden et al., 2002). The subsidence rate of the western Netherlands compared to Belgium was calculated by Kiden et al. (2002) to be 0.66 m/kyr at 7.5 ka and 0.25 m/kyr since 5 ka. This subsidence also contains a tectonic component, estimated to account for 0.06 to 0.16 m/kyr of the sagging. The assumed position of the forebulge as a result of the FIS is displayed on Fig. 7. This figure also contains contour lines of the total post-LGM uplift in Scandinavia as calculated by Mörner (1979).

The absolute former forebulge subsidence estimates from Mörner (1979) count up to 170 m of collapse, but Steffen and Wu (2011) state that this value overestimates reality. They used a 3D model from Steffen et al. (2006) to calculate the subsidence and uplift as a result of the FIS, obtaining a forebulge of maximum 20 m crossing the northern parts of the Netherlands and Germany and extending towards the Dogger Bank, roughly in correspondence with the reconstruction from Kiden et al. (2002). According to Le Meur (1996), the elevation of the forebulge is roughly one-tenth of the central depression beneath the adjacent ice sheet. Calculations from Vink et al. (2007) on glacio-isostasy between 8 and 4.8 ka give estimations of northwest Germany to have subsided 7.5 m more compared to Belgium and 2.5 m more

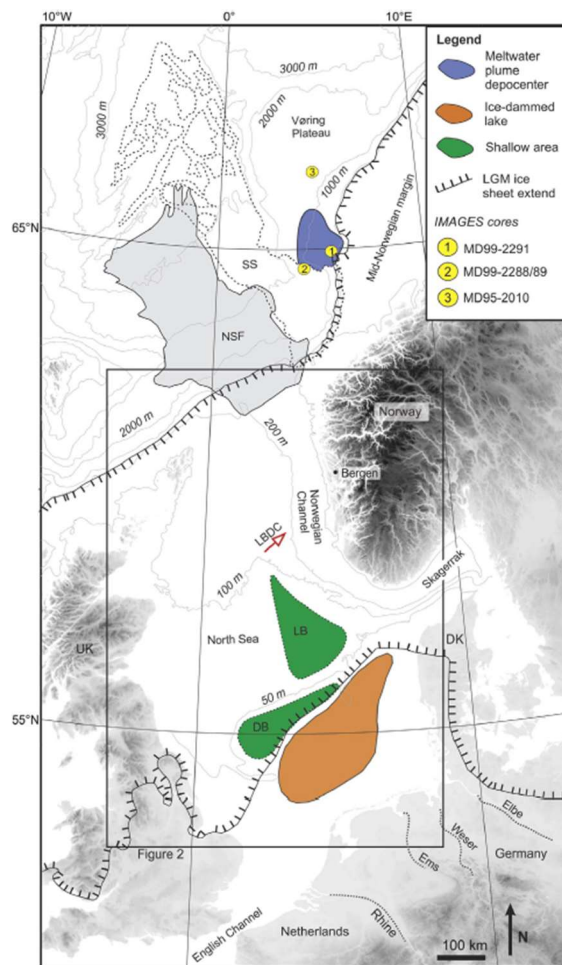


Figure 6: Map of the North Sea region with indication of Late Weichselian North Sea Lake (brown). DB: Dogger Bank; LB: Ling Bank; LBDC: Ling Bank Drainage Channel (Hjelstuen et al., 2018).

compared to the western part of the Netherlands. They estimated subsidence rates of 2.67 m/kyr in the western Netherlands and 3.33 m/kyr in northwest Germany between 9 ka and 7.5 ka, while these values decrease to 0.4 m/kyr for the Netherlands and 1.2 m/kyr for Germany for 7.5 to 5 ka. Since 5 ka, subsidence becomes negligible in the Netherlands and decreases to 0.4 m/kyr in Germany. These results confirm that postglacial forebulge subsidence related to decay of the FIS decreases towards the southwest and with time, as suggested by Kiden et al. (2002). Their reconstruction of the region of maximum forebulge collapse is confirmed by Vink et al. (2007) as well.

Major discrepancies between some forebulge studies (e.g. Mörner (1979) versus others) as well as the fact that the effects of the BIIS are often not taken into account highlight the importance of a comprehensive study on the forebulge evolution through time.

2.4 North American situation and relationship with global climate

GIA-effects resulting from ice unloading after the LGM in North America can still be observed today. Vertical velocities of 10 mm/year are observed near Hudson Bay, where the former Laurentide Ice Sheet reached its maximum thickness. The hinge line between subsidence and uplift crosses the Great Lakes, and subsidence rates up to 1-2 mm/year are observed south of this transition zone (Sella et al., 2007). Calculations from Argus et al. (2020) resulted in similar uplift and subsidence rates (Fig. 8). During glacial times, the lithosphere beneath the ice sheet and a small periphery surrounding it were depressed, while the region south of the hinge line experienced slight uplift compared to a situation in equilibrium. This geomorphological situation provided possibilities for proglacial lake formation fed by meltwater discharge during deglaciation, of which Lake Agassiz is the largest known example (Teller and Leverington, 2004). The southern ice margin formed the northern border of this lake, of which the size and volume were dependent on the migration of this ice limit (Mann et al., 1999; Teller and Leverington, 2004). As the glacio-isostatic depression was larger in the north, this part of the lake was also the deepest, while the lake was rather shallow towards to the south

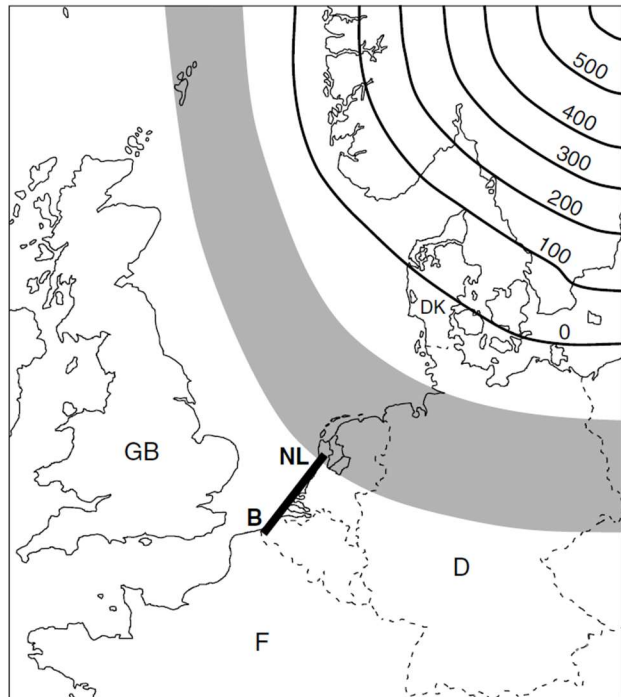


Figure 7: Position of the glacial forebulge as a result of GIA related to the Fennoscandian Ice Sheet (grey band) as assumed by Kiden et al. (2002) based on a transect studied to derive subsidence rates (black line); contour lines of total absolute GIA uplift after melting of FIS from Mörner (1979).

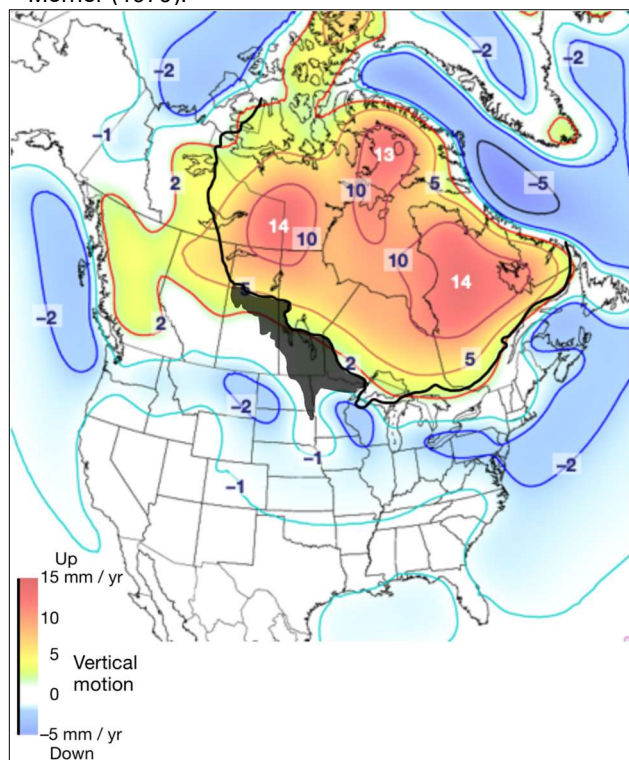


Figure 8: Current rates of vertical motion as a result of degradation of the Laurentide Ice Sheet between 15 ka and 5 ka (Supporting information from Argus et al. (2020)). Extent of Lake Agassiz (grey shading) and margin of the Laurentide Ice Sheet (solid black line) at the start of the Younger Dryas from Norris et al. (2021).

(Mann et al., 1999). The retreat of the Laurentide Ice Sheet together with accompanied changes in vertical motion resulted in a generally northward migration of Lake Agassiz (Teller and Leverington, 2004).

Different drainage routes from Lake Agassiz have been suggested. A change in meltwater drainage route from the Mississippi River into the Gulf of Mexico to drainage in the direction of the northern Atlantic Ocean via the St. Lawrence River was observed by Broecker et al. (1989), who suggested this being evidence for an outburst from Lake Agassiz having triggered the Younger Dryas by altering the global thermohaline circulation system. This drainage route was later confirmed by e.g. Carlson et al. (2007). Murton et al. (2010), however, provided evidence for the outburst flood being routed towards the northwest into the Arctic Ocean via the Mackenzie River during the start of the Younger Dryas. They did not rule out that the glacial advance during the Younger Dryas could have temporarily closed the northward draining route, causing subsequent meltwater release in a more southerly or easterly direction. The debate is not settled yet, as some recent publications reconfirm the possibility of a drainage route towards the east (Leydet et al., 2018), while other studies expand the evidence of an outburst flood towards the Arctic Ocean (Norris et al., 2021).

3. MATERIAL AND METHODS

3.1 Background

GIA calculations often rely on the so-called sea level equation, originally developed by Farrell and Clark (1976). This equation is solved to determine how relative sea level changes vary spatially and over time, as the meltwater redistribution resulting from ice and ocean feedbacks is calculated. The final isostatic adjustment is then determined by the response of the solid earth to the sum of the water and ice loading. The sea level equation thus combines the input, i.e. the applied load (ice and water) and the solid Earth model, with the output, which is the GIA response. Solving the sea level equation requires calculations on the interplay between ice masses worldwide and global variations in sea level, which can easily become quite complicated. For this reason, the sea level equation was not applied in this study as the water load has not been taken into account (see sections 3.2 and 4.2).

Le Meur and Huybrechts (1996) compared five different GIA Earth models to model the Antarctic Ice Sheet over the last glacial cycle. Their most complex model is a self-gravitating viscoelastic spherical Earth model (SGVE), originally developed by Le Meur (1996). This model considers the complete solid earth, containing an elastic lithosphere, a viscoelastic upper and lower mantle and an inviscid core. The results from this SGVE model were compared to the results from four more simplified models, which are combinations of two ways to represent the lithosphere and two ways to imitate the asthenosphere. A local lithosphere (LL) is considered to lack flexural rigidity resulting in a local response to loading, while an elastic lithosphere (EL) accounts for this flexural rigidity. Time-lagged redistribution of mantle material is considered using a diffusion equation (diffusive asthenosphere, DA) or an equation describing an exponential decay of the response (relaxed asthenosphere, RA). Combinations of the different treatments of the two layers are then LLDA, ELDA, LLRA and ELRA. Le Meur and Huybrechts (1996) showed that using the SGVE model resulted in the most correct representation of the Antarctic Ice Sheet. As they mention as well, the difference in results between the SGVE and ELRA models is however quite minimal. The ELRA model is computationally much more user-friendly than SGVE and is thus a decent alternative when one still wants to perform accurate GIA calculations.

3.2 GIA calculations

The Elastic Lithosphere-Relaxing Asthenosphere (ELRA) treatment of the solid Earth is the basis for the GIA model used in this study. As stated before, this model consists of two components. First, the “final” response i.e. the equilibrium state resulting from a load is calculated considering an elastic lithosphere. The deflection of the plate thus depends on its flexural rigidity, which depends on the lithospheric thickness. This “final” response is reached exponentially, so theoretically the deflection is only complete at an infinite time after emplacement of the load. It should be noted that this exponential decay considers a constant loading, which is not the case in reality as ice sheets grow and retreat during glacial cycles.

All calculations are carried out in MATLAB® (The MathWorks Inc., 2022). The code for calculating the steady-state Elastic Lithosphere component of the ELRA model is based on the code implemented in

Kori-ULB (Coulon et al., 2021; Pattyn, 2017), an ice flow model also accounting for GIA effects. The code for this model is completely publicly available.

3.2.1 Elastic Lithosphere component

The deflection of the elastic lithosphere w is calculated by a fourth order differential equation (Huybrechts and de Wolde, 1999; Le Meur and Huybrechts, 1996):

$$D\nabla^4 w + \rho_a g w = q \quad (1)$$

D is the flexural rigidity (expressed in Pa m³) and is calculated using (Brothie and Silvester, 1969):

$$D = \frac{ET_e^3}{12(1-\nu^2)} \quad (2)$$

where E represents the Young's modulus (100 GPa), ν is Poisson's ratio (0.25) and T_e the thickness of the effective elastic lithosphere (e.g. Tesauro et al., 2012). Le Meur and Huybrechts (1996) determined a flexural rigidity value of 10²⁵ Pa m³ for their ELRA model fitted to the SGVE model. This value corresponds to a lithospheric thickness of approximately 104 km.

Equation (1) further accounts for the density of the asthenosphere ρ_a (3300 kg m⁻³), the gravitational acceleration g (9.81 m s⁻²) and the load q . The total load is the sum of the ice and the water load. For this study only the ice load is taken into account. Accounting for the water load is rather complicated. One should use an accurate local sea level reconstruction for the complete glacial cycle, taking into account that not just all areas that are below sea level (according to this local relative sea-level curve) will be filled with water, as this is only the case when these areas are connected to other water bodies. For the study area, neglecting the water load when focusing on the period before the North Sea became inundated after the last deglaciation can be justified by the reasons further explained in section 4.2 (Model simplifications). The load used as input in the ELRA model for this study therefore only represents the ice load. This ice load is taken from different ice sheet reconstructions. These ice sheet reconstructions are further explained in section 4.1 (Ice sheet reconstructions). They consist of grids in which the value of a grid cell represents the ice thickness at a specific location. The load q thus only depends on the ice thickness h (expressed in m), the gravitational acceleration g and the ice density ρ_i (917 kg m⁻³):

$$q = \rho_i g h \quad (3)$$

As was shown by Le Meur and Huybrechts (1996) and Huybrechts and de Wolde (1999), equation (1) can be solved for the vertical lithospheric displacement w in function of the point load P_w ($= q \times \text{area}$) and the distance from the point load r (with $x = r/L_r$ the normalised distance from the point load):

$$w(x) = \frac{P_w L_r^2}{2\pi D} \text{kei}(x) \quad (4)$$

Here is L_r the so-called radius of relative stiffness, or in other words the flexural length scale:

$$L_r = (D/\rho_a g)^{1/4} \quad (5)$$

As can be seen, this value depends on the flexural rigidity (and thus on the lithospheric thickness) and approximates 130 km for the chosen rigidity value of 10²⁵ Pa m³.

Kei(x) is a Kelvin function of zero order, which is defined as the imaginary part of the modified Bessel function of the second kind (Brothie and Silvester, 1969; Oldham et al., 2008). This is an oscillating function, while the amplitude of its oscillation decreases exponentially (Oldham et al., 2008). Kelvin functions are often used to solve differential equations amongst others related to wave propagation and the deformation of plates.

Equation (4) only solves for the deflection of the lithosphere resulting from a point load. To calculate the total deflection at a certain location, the contributions of all neighbouring point loads are summed together.

3.2.2 Relaxing Asthenosphere component

The rate of change in vertical displacement is further determined by:

$$\frac{\delta w}{\delta t} = -\frac{1}{\tau_a}(w - w_{ss}) \quad (6)$$

where w_{ss} is the steady-state or “final” displacement towards which w decays exponentially if the ice-thickness would remain constant. τ_a is a time-constant depending on the properties of the asthenosphere, and is taken to be 3000 years in accordance with the suitable ELRA model of Le Meur and Huybrechts (1996), who state that this relaxation time corresponds to an upper-mantle viscosity of 5×10^{20} Pa s.

The solution of equation (6) gives the actual amount of displacement at a given time t after initiation of the ice load (e.g. Greve, 2001):

$$w(t) = w_{ss} + (w_0 - w_{ss})e^{-t/\tau_a} \quad (7)$$

with the initial displacement w_0 at $t = 0$, which will decay towards w_{ss} over time. The time-constant does not indicate the moment when displacement is in equilibrium with the ice load. As the decay is exponential, this equilibrium state will theoretically only be reached when time approaches infinity. When $t = \tau_a$ and the initial displacement was zero, the actual displacement will be $100 - (100/e) \%$, i.e. 63.212 %, of the equilibrium displacement (Fig. 9).

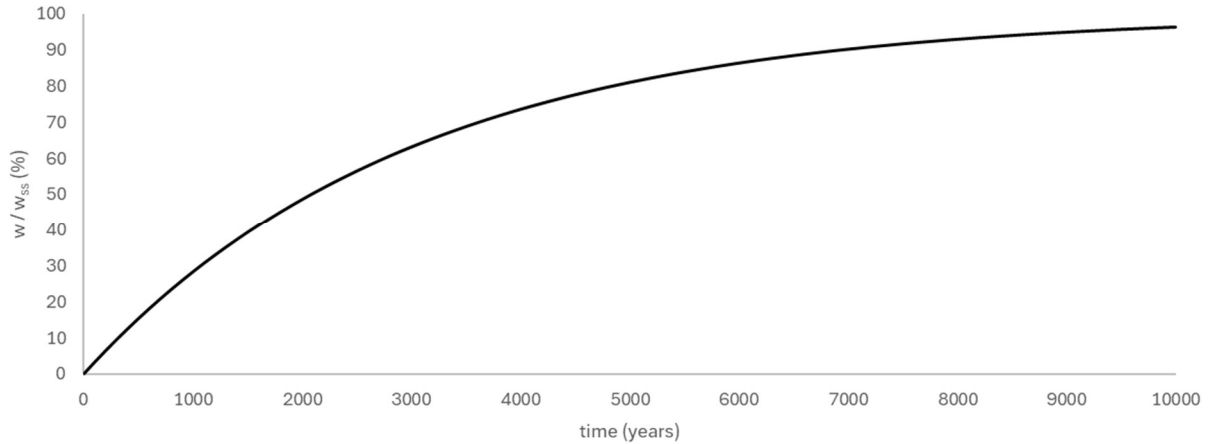


Figure 9: Exponential adjustment of the lithosphere to a constant ice load emplaced at time=0 years, with a time-constant of 3000 years.

The ice sheet reconstructions that were used as ice thickness input have different temporal resolutions between reconstructions at consecutive time slices, often of 1 kyr to several kyr in between two subsequent reconstructions (see details in section 4.1). To model the GIA response over time starting at X ka, one could calculate the equilibrium GIA response w_{ss} to the ice load at X ka, let the lithosphere adjust exponentially to the applied load until e.g. X-2 ka, where t would equal 2000 years. The GIA response at X-2 ka can then be used as the new w_0 . The equilibrium GIA response resulting from the ice load at X-2 ka can serve as the new w_{ss} , t would be reset to 0 years and this whole process can be repeated again. The only problem with this approach is that the ice load at X ka is assumed to remain unchanged until X-2 ka, when the original ice load is abruptly replaced by another load. To account for this problem, a constant time step of 100 years was used, and the applied ice load was changed every 100 years. Where the time step in between ice sheet reconstructions was 2000 years as for the example above, the ice load at time steps in between these 2000 years was linearly interpolated based on the closest available younger and older ice sheet reconstructions. If the ice load at X ka is load a and the ice load at X-2 ka is load b , then the ice load at X-0.1 ka equals $\frac{(2-0.1) \text{ kyr}}{2 \text{ kyr}} a + \frac{0.1 \text{ kyr}}{2 \text{ kyr}} b$ or $\frac{19}{20} a + \frac{1}{20} b$.

The lithosphere thus relaxes exponentially to ice load a for $t=100$ years, after which a new ice load is applied. This process starts at the oldest ice load available for every reconstruction and is run iteratively until the present day. A substantial difference in GIA response is found between a time step of 1 or several kyr and a more accurate time step of 100 years, while the GIA response between a time step of 100 years and smaller time steps such as 10 years becomes negligible. The enormous increase in computation time as time steps become smaller justifies the use of a time step of 100 years.

3.3 Map projections

Four different ice sheet reconstructions were used: PaleoMIST 1.0, ICE-6G_C, UMISM and BRITICE-CHRONO. The PaleoMIST 1.0 ice sheet reconstructions are available as Lambert azimuthal equal area (LAEA) projected grids with a spacing in of 5×5 km, where the value of every grid cell represents the ice thickness at that specific location. The ICE-6G_C reconstructions are available on a 1×1 degree grid, meaning that the area represented by grid cells is not always the same, while a constant area is a necessary requirement for the GIA calculations based on the model described above. The coordinates of the grid cells (geodetic latitude φ and longitude λ) were converted to LAEA projected easting and northing values based on the projection formulas as described by Snyder (1987):

$$easting = BD \cos \beta \sin(\lambda - \lambda_0) \quad (8)$$

$$northing = \frac{B}{D} [\cos \beta_0 \sin \beta - (\sin \beta_0 \cos \beta \cos(\lambda - \lambda_0))] \quad (9)$$

To be able to use the above equations, some intermediate calculations (see below) are necessary, where φ_0 and λ_0 are the latitude and longitude of natural origin (taken as 90° and 0° , respectively in accordance with the PaleoMIST 1.0 projection), a is the radius of the equator and e is the eccentricity, giving values of $a=6378137$ m and $e=0.081819191$ for the ETRS89-LAEA reference ellipsoid. All coordinate values are converted to radians first:

$$q = (1 - e^2) [\sin \varphi / (1 - e^2 \sin^2 \varphi)] - [(1/(2e)) \ln ((1 - e \sin \varphi)/(1 + e \sin \varphi))] \quad (10)$$

$$q_0 = (1 - e^2) [\sin \varphi_0 / (1 - e^2 \sin^2 \varphi_0)] - [(1/(2e)) \ln ((1 - e \sin \varphi_0)/(1 + e \sin \varphi_0))] \quad (11)$$

$$\varphi_P = \pi/2 \quad (12)$$

$$q_P = (1 - e^2) [\sin \varphi_P / (1 - e^2 \sin^2 \varphi_P)] - [(1/(2e)) \ln ((1 - e \sin \varphi_P)/(1 + e \sin \varphi_P))] \quad (13)$$

$$R_q = a \sqrt{(q_P/2)} \quad (14)$$

$$\beta = \sin^{-1}(q/q_P) \quad (15)$$

$$\beta_0 = \sin^{-1}(q_0/q_P) \quad (16)$$

$$B = R_q \sqrt{2/[1 + \sin \beta_0 \sin \beta + (\cos \beta_0 \cos \beta \cos(\lambda - \lambda_0))]} \quad (17)$$

$$D = a \cos \varphi_0 / \sqrt{1 - e^2 \sin^2 \varphi_0} / (R_q \cos \beta_0) \quad (18)$$

After projection of the coordinates to easting and northing values the area of grid cells is still not constant. A mesh grid with grid cells of 5×5 km was overlain on top of the projection and ice thickness values were linearly interpolated onto this mesh grid to obtain a matrix with the same dimensions and spatial resolution as the PaleoMIST 1.0 reconstructions. The same was done for the UMISM reconstructions, which were provided in grids with an original spatial resolution of 50×50 km. The BRITICE-CHRONO reconstructions were available as GIS files in the British National Grid reference system. They were first converted to coordinates using a build-in MATLAB® function and were in turn converted to the LAEA projection using the above-mentioned formulas. The projected ice thickness matrices are used as input for the P_w component of equation (4), where the area of every cell equals 25 km^2 .

3.4 Data visualisation and sensitivity tests

The displacement as a result of the different ice sheet reconstructions was visualised in plots made for every 5 kyr between the start of the reconstruction and 30 ka, and for every 1 kyr between 30 and 8 ka. For every plot a background displaying the present-day land-sea contours was used, and the GIA uplift was visualized by a white-to-red colour scale, while depression was displayed using a white-to-blue colour scale. These colour scales are used because they are perceptually uniform (the scaling from light to dark colours is linear) and are suitable to interpret for readers affected by colour vision deficiencies. To further improve the ability to grasp the amount of vertical displacement, contour lines of uplift are shown at intervals of 1 m, while depression contour lines are present at steps of 20 m. To have an idea

how the vertical displacement relates to the ice extent, ice sheet margins are visualised, while the GIA below the ice sheets is not shown as its interpretation is not the focus of this thesis.

The GIA response at the aforementioned time slices is also plotted on top of a modern bathymetry map of the southern North Sea to be able to discuss ancient drainage patterns, the possible formation of basins and other related depositional processes during the last glacial cycle. Only the now inundated North Sea region roughly between 51°N and 57 °N is considered, as larger depths more to the north and high elevations on land would dramatically decrease the interpretability of the maps.

The paleotopography of this region for that time can be derived by accounting for i) changes in sea level elevation, ii) differences in sedimentary cover between today and the past, and iii) vertical isostatic adjustment. The stratigraphy of this region is too complex and not enough understood to be able to “backstrip” sedimentary units deposited (and/or eroded) during the timeframe of interest. However, these sediments are considered, in general terms, for the interpretations in the Discussion chapter. The created maps display the sum of the modern bathymetry and the paleo ice-induced vertical motion for the North Sea region that is not covered by ice and is located above the paleo-sea level. The elevation of past sea levels is based on a sea level curve for the last 35 kyr published by Lambeck et al. (2014) and a sea level reconstruction extending back to the Eemian (and older) reconstructed by Lambeck et al. (2002) (Table A1, see appendix). Now-inundated areas on these maps which were also inundated at the moment of reconstruction are visualised in grey. Those would have been marine at the time. Areas lying beneath the past sea level but fully surrounded by emerged areas and ice are not coloured grey in order to facilitate paleogeomorphological interpretations. With no connection to the world ocean, those areas would not be marine despite lying below sea level. The topography of now-inundated areas emerged at the timing of reconstruction is visualised using a stepwise rainbow colour scale incremented every 10 m. GIA contour lines are shown as described above and the ice margin is indicated using a blue contour line.

It has to be mentioned that the paleo-sea level values used to display the sea level history over time are in fact global values and they might differ from the history in the North Sea region. However, the use of local sea level reconstructions would cause some problems. For the locations and time slices where local sea level reconstructions are available, these reconstructions are often poorly constrained. Moreover, they depend on the exact location and thus also vary in between the North Sea basin. This means that for every grid cell a different reconstruction would be required. Even if such high resolution data would be available (which is not the case), this approach would enormously increase the calculation time. Lastly, as these local reconstructions often already include a GIA component, their use would increase the risk of circular reasoning. As a GIA component is already included in the vertical land motion calculations from this study, it was decided to use eustatic sea level reconstructions.

Vertical displacement of the separate BIIS and FIS was calculated as well by calculating the GIA after successively emplacing a mask on the FIS and BIIS to evaluate the influence of ignoring one of these two ice sheets. The exact location on where to split the two ice sheets varies slightly between the reconstructions and is based on limits of the separate ice sheets when they were not merged and on ice thickness distribution over the North Sea during moments when the ice sheets were merged. This test is important as some earlier forebulge reconstructions have for example only focused on the influence of the FIS on GIA. Moreover, some ice sheet reconstructions are only available for one of the two ice sheets, so an evaluation of their usefulness for reconstructing the forebulge in the North Sea region is necessary as well. As for the BRITICE-CHRONO reconstruction only the southern part of the FIS (together with the complete BIIS) is reconstructed (see section 4.1). The impact of running the other ice sheet models only for this region is assessed as well to be able to quantify the reliability of using BRITICE-CHRONO for GIA reconstructions in the North Sea region.

For this study, a flexural rigidity of 10^{25} Pa m³ was used. There is, however, still much discussion about the European lithospheric thickness as very contrasting values are mentioned in literature. Bradley et al. (2011) found a thickness of 71 km for the British Isles. Milne et al. (2006), however, used a thickness of 95 km for the same geographical region. Vink et al. (2007) found a large variety in lithospheric thickness values in western Europe, ranging from 140 km under Belgium over 100 km underneath the Netherlands to 80 km below the northwest German coast. Studies from Bergstrand et al. (2005) and Lidberg et al. (2010) confirm a lithospheric thickness of 120 km below Fennoscandia based on GPS

observations. Several studies also report large variations of the lithospheric thickness underneath Europe (e.g. Artemieva, 2019). A sensitivity test was carried out to evaluate the influence of different lithospheric thickness values on the GIA response. The model was run for the different ice sheet reconstructions using flexural rigidity values of $3.0 \cdot 10^{24}$ Pa m³, $6.5 \cdot 10^{24}$ Pa m³ and $1.5 \cdot 10^{25}$ Pa m³, corresponding to lithospheric thickness values of 70 km, 90 km and 120 km, respectively.

The upper mantle viscosity of 5×10^{20} Pa s, corresponding to a time-constant of 3000 years, is often used in GIA studies, also when these studies make use of other models than ELRA (e.g. Bradley et al., 2011; Lidberg et al., 2010; Milne et al., 2006). A sensitivity analysis was carried out to evaluate the GIA response to different upper-mantle viscosities. Calculations are performed for τ_a values of 1000, 6000 and 10000 years, covering the range of acceptable upper-mantle viscosity values.

To compare the different ice sheet reconstructions, the ice volume over time was calculated for the three reconstructions going back to at least 80 ka and containing all Eurasian ice sheets based on the sum of all thickness values and on the knowledge that every grid cell represents an area of 25 km². Similarly, the ice-covered area and average ice thickness over time are calculated, however by excluding ice thickness values lower than 1 m, because they influence these last two variables too much and are only of a minor importance.

4. MODEL CHOICES

4.1 Ice sheet reconstructions

Three different available ice sheet reconstructions of the British-Irish, Fennoscandian and Barents-Kara ice sheets were used as input for the GIA model: ICE-6G_C (VM5a), PaleoMIST 1.0 and UMISM. They were calibrated against a BIIS reconstruction also containing the southern part of the FIS; BRITICE-CHRONO. Their characteristics and the input data that was used to create them are described below.

The ICE-6G_C (VM5a) model presented by Argus et al. (2014) and Peltier et al. (2015) is a global model based on the sea-level equation. This model actually consists of two parts, where VM5a is the name of the model incorporating the viscoelastic structure of the interior of the planet. These viscoelastic properties are used to create the actual model displaying variations in thickness of the continental ice sheets: ICE-6G_C (Peltier et al., 2015). The ICE-6G_C model was preceded by earlier models in the ICE-nG series. These global models are based on dated ice sheet margins and the global and local sea level curves. The subsequent models in this series build upon refinements from the previous version in the sequence. Together with the ICE-nG models, VMx viscosity models are generated based on effects from the ice masses used in the previous ICE-nG model on the earth's lithosphere and mantle. The new ICE-nG model is calibrated using the updated VMx model, after which the new ice sheet model can be used to validate the latest viscosity model. The ICE-6G_C (VM5a) model is an update of the ICE-5G (VM2) model and was constructed by using available GPS measurements of crustal vertical motion, dating of erratic trim lines based on their exposure age to infer information on ice cover thickness and timing and by using data on relative sea level changes (Argus et al., 2014; Peltier et al., 2015). The ice margins reconstructed by Gyllencreutz et al. (2007) are considered as well. The VM5a viscosity model was described by Peltier and Drummond (2008) and contains an upper mantle with a viscosity of 5×10^{20} Pa s overlain by a stratified lithosphere. This lithosphere has a thickness of 100 km, with the top 60 km being perfectly elastic and the bottom 40 km having a viscosity of 10^{22} Pa s. The ICE-6G model is so far the latest version in the series for which reconstructions are available for the complete last glacial cycle. Ice sheet model ICE-7G_NA (VM7) is already available, but reconstructions derived from this model only go back until 26 ka at this moment. For consistency, it was decided to rely on the reconstructions derived from ICE-6G_C (VM5a). This model provides reconstructions from 122 ka to the present day, with time steps of 2 kyr between 122 ka and 32 ka, increasing to a temporal resolution of 1 kyr between 32 ka and 21 ka, further increasing to 0.5 kyr before 21 ka and even to 0.25 kyr before 15 ka. The reconstructions are available on a 1×1 degree grid. They were downloaded from <https://www.atmosp.physics.utoronto.ca/~peltier/data.php>.

Quite recently, another ice sheet model was presented by Gowan et al. (2021): PaleoMIST 1.0 (Paleo ice sheet Margins, Ice Sheets and Topography). Like the ICE-6G_C model, PaleoMIST 1.0 is a global model where ice sheets everywhere on earth are reconstructed together. This new model, however,

does not rely on far-field sea level records nor on indirect $\delta^{18}\text{O}$ proxy records, as is often the case for pre-LGM ice sheet reconstructions for which other information becomes scarce. On the contrary, PaleoMIST 1.0 is constructed using ice physics, geological observations and knowledge of local changes in relative sea level (Gowan et al., 2021). As is the case for the ICE-6G_C reconstruction, ice sheet reconstructions for every time slice are run iteratively several times until they fit the response of the Earth structure and the calculated sea level. Sea level calculations were performed using the SELEN software (Spada and Stocchi, 2007). The Earth structure was represented by a 1D, spherically symmetric three-layered body. This model contains an elastic lithosphere with a thickness of 120 km, an upper mantle with a viscosity of 4×10^{20} Pa s and a 4×10^{22} Pa s viscosity lower mantle. These parameters were used to represent the complete Earth and are consistent with values often used in GIA studies in Eurasia (e.g. Steffen and Wu, 2011). The ice sheets themselves were reconstructed by using the ICESHEET program (Gowan et al., 2016). This software requires three main sets of input parameters; position of the ice sheet margins, values of basal shear stress and the basal topography. Most of the margin reconstructions that were input for PaleoMIST 1.0 and at least these used for the ice sheets in Eurasia are based on existing reconstructions. The Eurasian margins are taken from reconstructions by Helmens (2014), Hughes et al. (2016) and Ingólfsson and Landvik (2013). The initial values of basal shear stress were based on the current geology and topography. These values were adjusted during the subsequent iterations to reduce the error, i.e. to make sure that the sea level calculations aligned with the independent sea level indicators (Gowan et al., 2021). The final PaleoMIST model provides ice sheet reconstructions from 80 ka to the present day, with a temporal resolution of 2.5 kyr. The reconstructions are available on a Lambert azimuthal equal-area projection grid with a spatial resolution of 5×5 km and were downloaded from Gowan (2019).

A third model that was used as input for the GIA studies is a reconstruction based on the University of Maine Ice Sheet Model (UMISM) (Näslund, 2010). In contrast to the ICE-6G_C and the PaleoMIST 1.0 global ice sheet models relying on the sea level equation, the UMISM-based reconstruction is only carried out on a regional scale (i.e. focusing on the ice sheets in Eurasia). This reconstruction described in Näslund (2010) is created using UMISM (Fastook, 1994; Fastook and Chapman, 1989; Fastook and Prentice, 1994), which is a model based on thermo-mechanics. Three important characteristics of the ice sheets are considered in the model; mass balance, ice flow and ice temperatures. The model uses differential equations to solve the conservation of mass, momentum and energy, respectively (Näslund, 2010). The model requires the input of data on past temperatures. For this reconstruction, a temperature proxy record of the last 120 kyr from the Greenland ice core GRIP (Dansgaard et al., 1993) was used. A mass balance empirical relationship derived from data on the Antarctic Ice Sheet was adopted to mathematically describe precipitation, which also depends on distance from the pole, the saturation vapour index (which is a function of the altitude and the lapse rate) and the surface slope (Fastook and Prentice, 1994). The spatial pattern of air temperature and precipitation patterns depend on the topography, which is described using a Digital Elevation Model (Näslund, 2010). To accommodate for this, an Earth model is necessary. In contrast to the previous models, the UMISM reconstruction relies on a hydrostatically supported elastic plate model. This is necessary to take account for the altitude of the ice sheet, which influences the mass balance by affecting air temperature and precipitation (Näslund, 2010). The reconstruction starts at 120 ka and was run iteratively on 5 year time steps until the present day using the aforementioned thermo-mechanical parameters. The output data from the reconstruction were provided by Jens-Ove Näslund and Peter Schmidt on an equidistant grid of 50×50 km. The reconstructions from 120 ka to the present day were provided with a temporal resolution of 1 kyr between 120 and 20 ka and a temporal resolution of 0.1 kyr between 20 and 0 ka. A small area of potential ice cover on Ireland lies outside the provided equidistant grid, but this area accounts to a maximum of 1.5 % of Eurasian ice volume in the other ice sheet reconstructions; and as the largest part of the BIIS is located in between Ireland and the zone of interest in the North Sea region, no or only a negligible influence on the GIA results within the region of investigation is expected.

These three ice sheet reconstructions are calibrated against the well-constrained BRITICE-CHRONO reconstruction recently presented by Clark et al. (2022b). This ice sheet reconstruction is based on a

geochronometric dating programme across the British Isles, their continental shelves and in the North Sea. The research project resulted in the gathering of 690 geochronometric ages (relying on radiocarbon, optically stimulated luminescence and terrestrial cosmogenic nuclides) based on geomorphologic and stratigraphic land samples and seafloor sediment cores of which the locations were chosen based on available geophysical information. The amount of sites dated to constrain ice sheet retreat has increased by a factor three due to BRITICE-CHRONO. This information resulted in 23 publications on ice advance and retreat, assembled in the recently published map-based reconstruction of the BIIS and its coalescence with the FIS (Clark et al., 2022b). Ice extent, basal topography and basal shear stress were used as input for the ICESHEET 1.0 program (Gowan et al., 2016), which was used as well for the PaleoMIST 1.0 reconstruction. As the BIIS interacted with the FIS, the model was also run for a region covering Norway, Denmark and Southern Sweden, aligned with their ice margins reconstructed by the DATED-1 project (Hughes et al., 2016). The output of this model was used to run the more complicated numerical dynamical Parallel Ice Sheet Model (Bueler and Brown, 2009; Winkelmann et al., 2011), to match the ice sheet dynamics with the thicknesses derived from the ICESHEET 1.0 program. The modelled thickness reconstructions thus suit the empirical data derived from the field campaigns. The reconstructions only cover roughly MIS 2, with ice sheet extent and thickness data available from 31 ka to 15 ka in time slices of 1 kyr. They can thus not be used to reconstruct GIA during the complete last glacial cycle, but as the most recent and probably most well-constrained reconstruction the BRITICE-CHRONO reconstruction can be used to validate the other ice sheet reconstructions. The reconstructions were downloaded from Clark et al. (2022a) and were available as GIS data in the Ordnance Survey National Grid (OSNG), also known as the British National Grid (BNG) reference system.

A comparison of the different Eurasian ice sheet models demonstrates that the differences between reconstructions are quite substantial. The UMISM and ICE-6G_C models both go back until ca. 120 ka, but where the ice sheet volume remains negligible until MIS 4 according to UMISM, ice volumes in ICE-6G_C already grow quite quickly to a third of the volume they reached during the LGM (Fig. 10). The PaleoMIST 1.0 reconstruction starts at 80 ka, the moment at which the three reconstructions show an ice build-up towards the MIS 4 peak. This maximum is reached at 60 ka for PaleoMIST 1.0, a few thousand years after ICE-6G_C and a few thousand years before UMISM. The MIS 2/LGM peak is reached in the same order and would have occurred at 20 ka according to PaleoMIST 1.0. Where ice disappears almost completely between MIS 4 and MIS 2 according to PaleoMIST 1.0 and UMISM, the volume decreases only slightly according to ICE-6G_C. After the LGM, ice volumes in the three reconstructions decrease rapidly, resulting in an ice-free region of interest since the start of the Holocene. Similar patterns are observed regarding ice extent and mean ice thickness (Fig. A1 and A2,

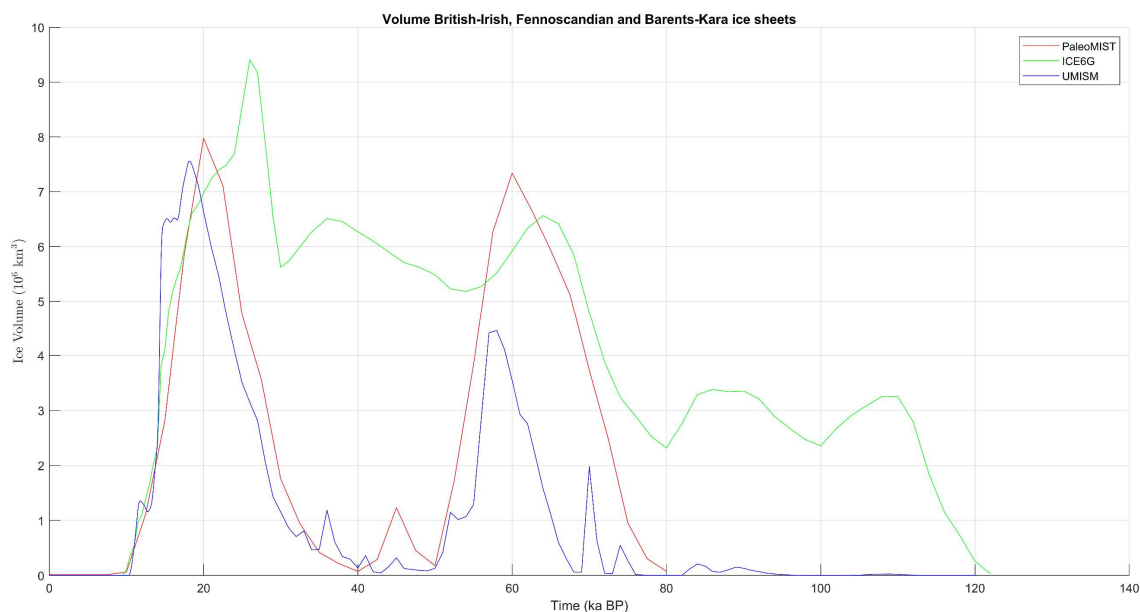


Figure 10: Evolution of ice volumes over time for the three Eurasian ice sheet models used in this study.

see appendix). It should be kept in mind that the UMISM reconstruction does not show the western part of Ireland, but this region only accounts for maximum 1.5 % of the ice volume in the other two reconstructions.

Fig. 11 shows the ice sheet reconstructions of the different Eurasian models at 20, 40 and 60 ka to visualise the regional differences in ice extent and thickness. The fact that the PaleoMIST 1.0 reconstructions were originally available in a much higher resolution than ICE-6G_C is clearly visible. At 20 ka, UMISM creates the largest ice extent and thickness in the vicinity of the region of interest, while ice sheets in the Barents-Kara are clearly less developed compared to the other reconstructions. The ICE-6G_C reconstruction shows a reduced BIIS-FIS connection compared to the other reconstructions. Similar patterns can be observed at 60 ka. At 40 ka, ICE-6G_C displays a quite extensive ice cover and thickness in the North Sea region, while the other reconstructions indicate no ice on the British Isles and only a minimal ice cover in Fennoscandia and the Barents-Kara region.

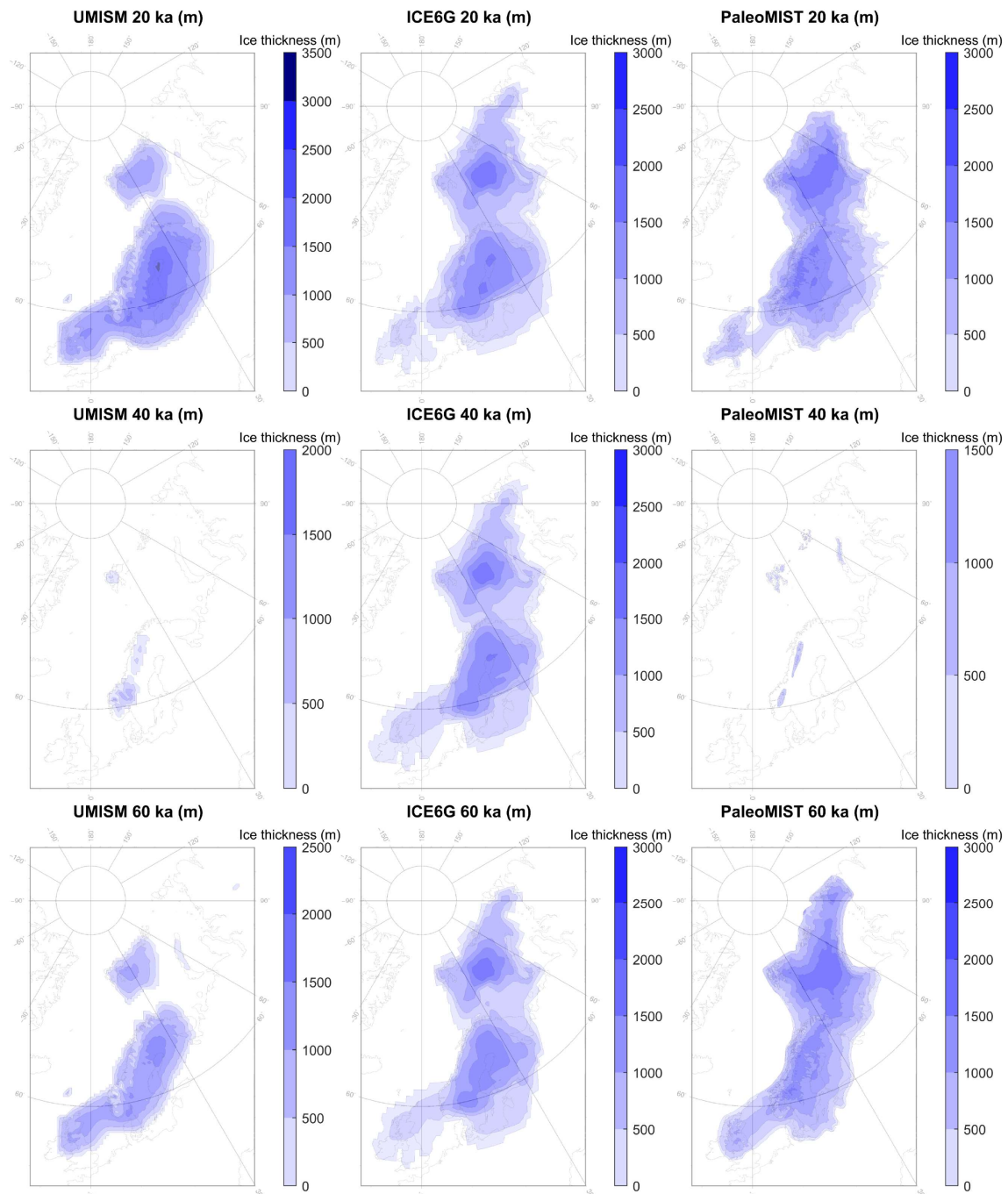


Figure 11: Ice cover at 20, 40 and 60 ka according to the different used Eurasian ice sheet models.

4.2 Model simplifications

An important aspect that is not taken into account is the hydro-isostatic loading. During periods where (a part of) the North Sea was inundated, the weight of the water load would have depressed the lithosphere beneath it further. As incorporating this effect would complicate the calculations enormously, it was decided to only focus on the isostatic compensation as a result of ice sheet loading. During periods of extensive ice sheet growth, the hydro-isostatic effect is expected to only account for a rather small portion of the total isostatic response. The largest part of the North Sea is only up to 100 m deep, while the ice sheets were often several kilometres thick. The slightly higher density of seawater compared to ice cannot compensate for the thickness difference between the water and ice column. The fact that the region of interest is now inundated and has been as well during certain periods in the past has to be taken into consideration. However, the largest part of the North Sea basin, including the main region of interest for this study, was not covered by water roughly since 80 ka until the start of the Holocene (Fig. 4 and 5). As water and thus sea level changes are not taken into account, neither are rotational feedback mechanisms, which in turn have an influence on the shape of the sea surface and geoid shape (Milne and Mitrovica, 1998) and will thus induce deformation of the solid earth.

Lateral variations in earth parameters such as lithosphere thickness and mantle viscosity are not taken into account as well. One constant value is used for the complete area of investigation. A sensitivity test is, however, carried out to estimate to what extent variations in these parameters influence the final GIA result. It is possible to incorporate regional variations of these parameters in the ELRA model, but then the rather simple solution using the kelvin function cannot be used anymore. The model would have to be solved numerically, which would require much higher computational power. The introduced complexity would result in a level of accuracy of the isostatic rebound predictions that lies beyond the scope of this thesis.

Furthermore, as only isostatic responses to ice (un)loading are calculated, uplift or depression as a result of geological processes such as sedimentation and tectonism are not considered either. River systems supplied a non-negligible amount of sediments to the North Sea basin during the Quaternary, which will have resulted in additional subsidence within this basin. Crustal movements related to e.g. graben systems have also been important during the Cenozoic history of the region (Ziegler and Louwerens, 1979), thus altering the total amount of subsidence or uplift.

5. RESULTS

5.1 Influence of ignoring northern part of FIS on the study area

The difference between running the GIA model with the full ICE-6G_C, PaleoMIST 1.0 and UMISM reconstructions and only using the same reconstructed region as for the BRITICE-CHRONO ice sheet model as input is shown on Fig. A3 (see appendix). The fact that the difference in vertical displacement is negligible for the region of interest justifies the use of the BRITICE-CHRONO reconstruction as a fourth equally useful reconstruction, albeit within the limited available timeframe.

5.2 Influence of separate ice sheets

Calculations of the displacement when only taking the BIIS or FIS into account give quite different results from calculations relying on both ice sheets together. As can be seen on Fig. 12, the area of maximal forebulge uplift passes through Scandinavia when only the BIIS is used as input, while the British Isles experience uplift when only the FIS is taken into account. When both ice sheets are considered together, the area of maximal uplift at 20 ka is situated in the Netherlands and adjacent areas in Belgium and Germany. While the FIS contributes for at least more than half of the uplift in this area for the different reconstructions, the contribution of the BIIS to this uplift is not negligible. Apart from the significant contributions of both ice sheets, ignoring one of these ice sheets would also result in a completely different orientation of the forebulge, which in turn can alter interpretations of paleo-drainage patterns.

5.3 Sensitivity of Earth parameters

Based on the GIA model that was run using the different ice sheet reconstructions and the different lithospheric thickness values of 70 km, 90 km and 120 km as input, it can be concluded that a thinner lithosphere results in a stronger depression than what would be the case for a thicker lithosphere.

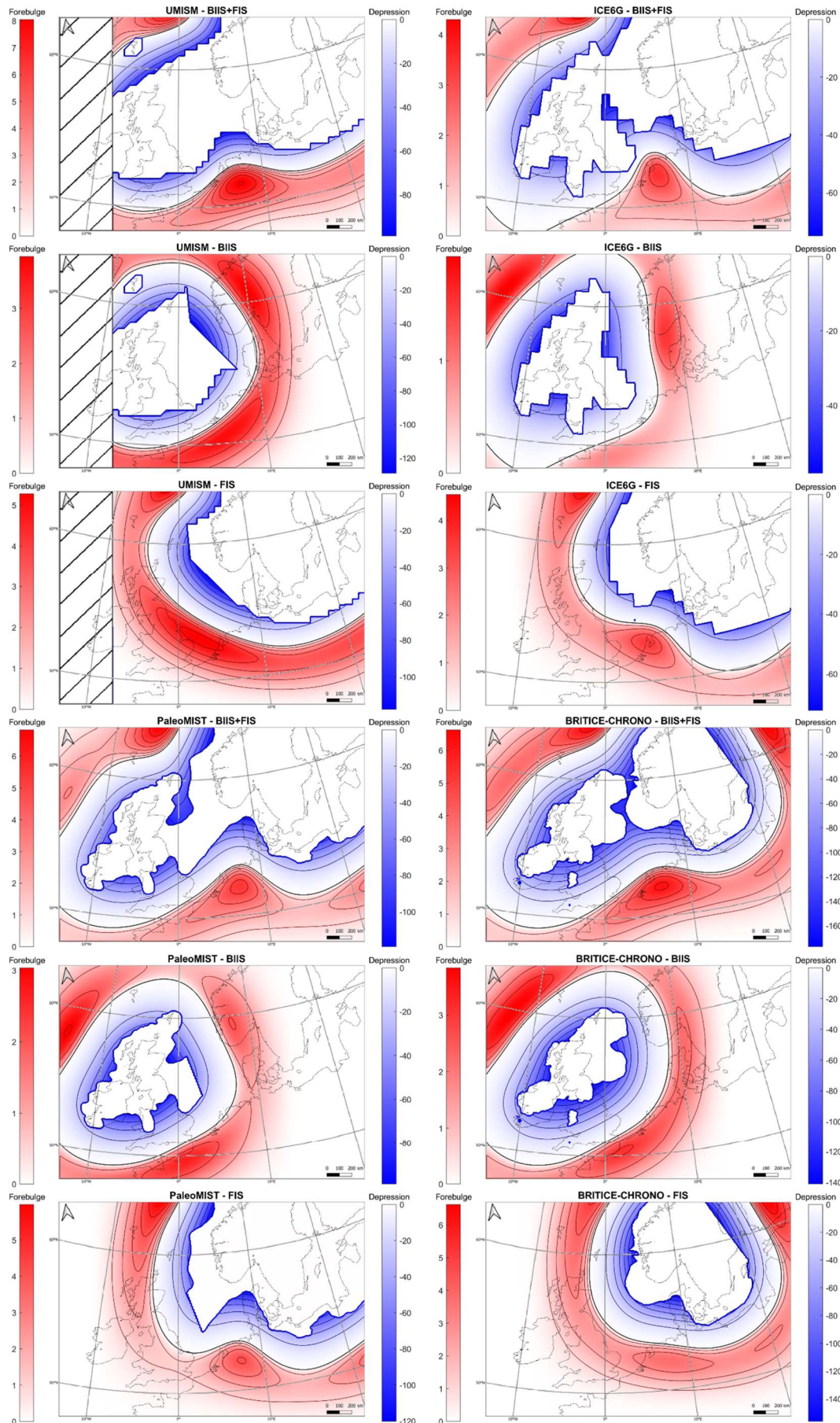


Figure 12: Vertical displacement (m) in the study area at 20 ka for the different ice sheet reconstructions when considering both the British-Irish and Fennoscandian ice sheets, only the BIIS and only the FIS.

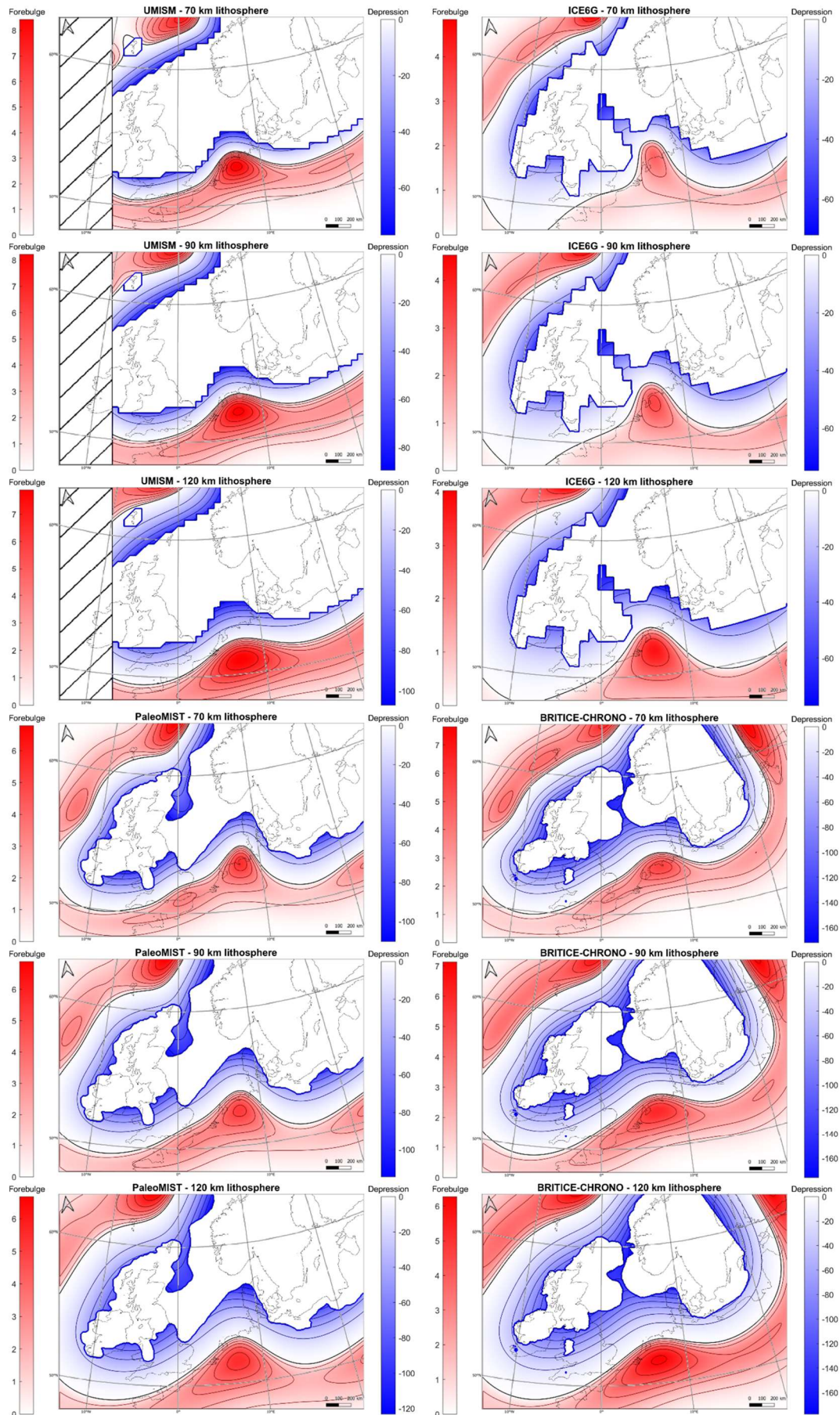


Figure 13: Reconstruction of the vertical displacement (m) at 20 ka for lithospheric thickness values of 70 km, 90 km and 120 km for the different ice sheet reconstructions.

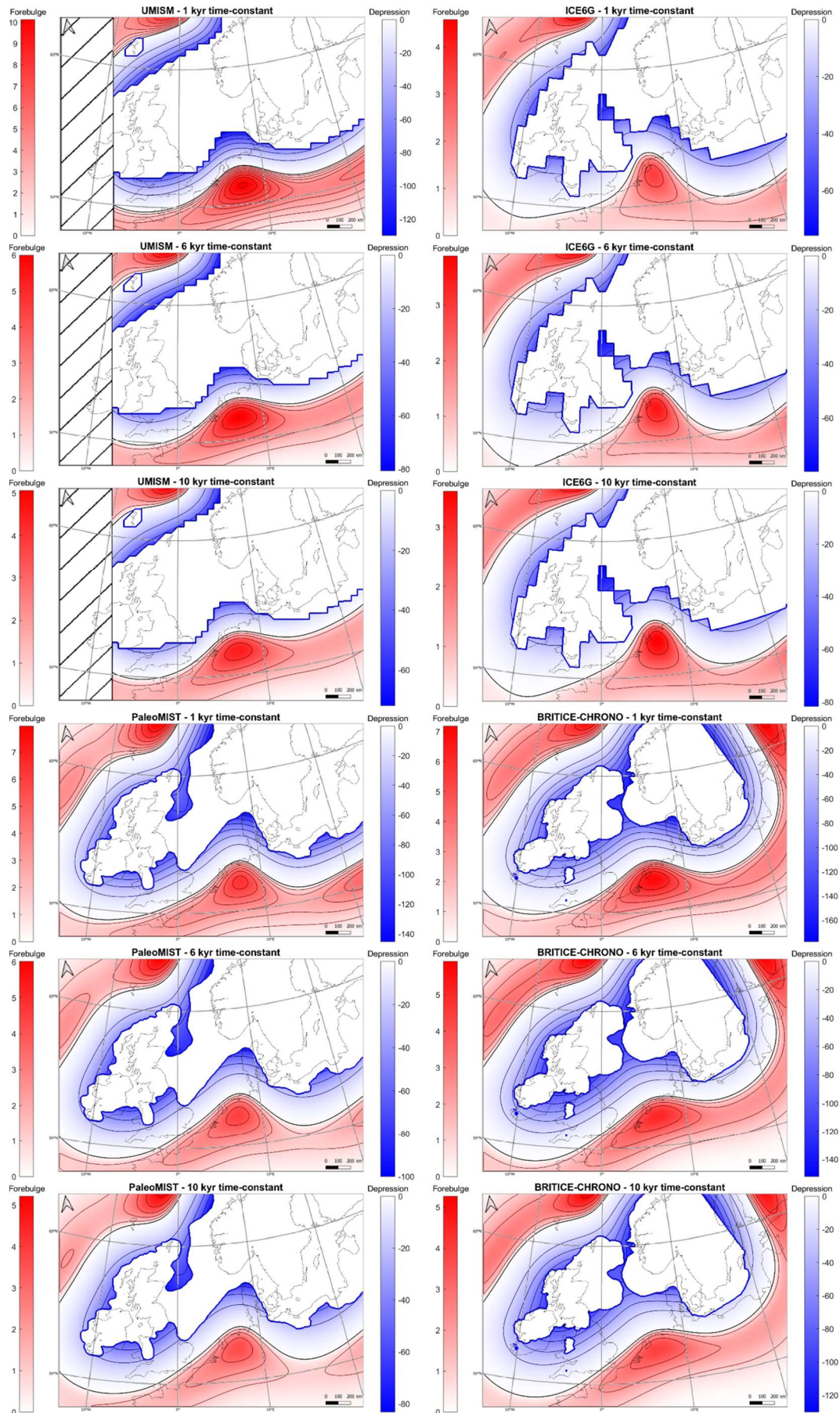


Figure 14: Reconstruction of the vertical displacement (m) at 20 ka for time-constant values of 1000 years, 6000 years and 10 000 years for the different ice sheet reconstructions.

Regarding the maximal forebulge uplift, a thicker lithosphere seems to result in a more elevated forebulge compared to a thinner lithosphere. This correlation is, however, less pronounced than for the depression. The evolution of the maximal uplift in the region of interest and the maximal depression over time for the different ice sheet models and lithospheric thicknesses is shown on Fig. A4 (see appendix). Fig. 13 displays the geographical configuration of the vertical displacement at 20 ka for the different lithospheric thickness values. A thicker lithosphere results in a forebulge being located further away from the ice sheet.

The influence of the viscosity of the asthenosphere, represented by variations in relaxation time, on the maximal uplift and depression over time is displayed on Fig. A5 (see appendix). A shorter time-constant results in more pronounced vertical displacement during times when ice cover is more substantial, while the opposite effect is observed during interstadials. Then, the vertical displacement is smaller compared to reconstructions relying on a slower relaxation time. The geographical distribution of vertical displacement at 20 ka is shown on Fig. 14. When the relaxation time is longer, the forebulge is located closer to the ice sheet compared to reconstructions relying on a shorter relaxation time.

5.4 GIA response in the North Sea region throughout the last glacial period

The GIA response is modelled starting at the beginning of the last glacial period up until when the North Sea has gained more or less its present-day dimensions during the Early Holocene. Reconstructions younger than 8 ka will become increasingly inaccurate when hydro-isostasy is ignored and thus lie beyond the scope of this thesis. Only the plots depicting certain time slices that are essential for understanding the general evolution of vertical displacement over time are displayed in section 5.4. The remaining time slices are shown in Fig. A6 (see appendix).

Even though ice build-up towards the MIS 4 peak only started around 80 ka, the substantial differences in MIS 5 ice extent and volume between UMISM and ICE-6G_C are reflected in varying GIA responses (Fig. 15). The first noteworthy ice sheet on the British Isles only appeared around 75 ka according to UMISM. This resulted in a semicircular forebulge of maximum 1 m uplift surrounding a limited Fennoscandian Ice Sheet, and amongst others passing through southern Denmark and over the central North Sea towards the Orkney and Shetland Islands. Depression in the northern North Sea in between the forebulge region and the FIS only amounted to a few meters. On the contrary, according to ICE-6G_C, ice sheets on Scandinavia and the British Isles were already quite extensive during MIS 5 and even reached thicknesses of up to 2 km around the Gulf of Bothnia. This resulted in the hinge line between uplift and depression also passing through Denmark towards Dogger Bank, but from there deflecting towards East Anglia, completely going around the British Isles and again coinciding with the UMISM hinge line north of Shetland. The North Sea east and south of Dogger Bank experienced uplift locally exceeding 2 m. The eastern and northern parts of the North Sea were a few meters depressed.

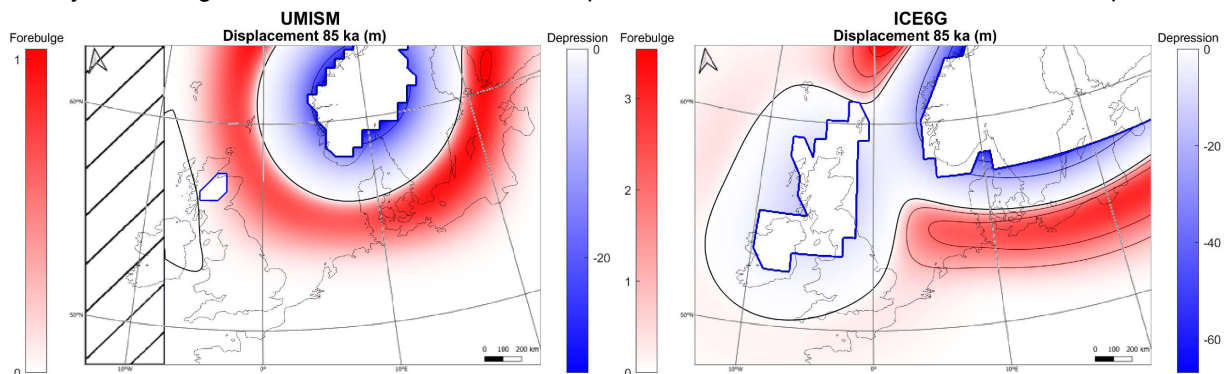


Figure 15: Reconstruction of the vertical displacement (m) at 85 ka based on the UMISM and ICE-6G_C ice sheet models.

At the MIS 5 - MIS 4 transition, the UMISM reconstruction also shows the development of a BIIS, having induced a similar vertical displacement signal as was already earlier observed for the ICE-6G_C reconstruction. According to PaleoMIST 1.0, ice on the British Isles only appeared shortly before its MIS 4 volume peak at 60 ka, having caused the development of a forebulge surrounding the FIS in similar

areas as was observed for UMISM prior to 80 ka. However, as ice at ca. 75 ka was more voluminous than during MIS 5, the forebulge could locally exceed 4 m uplift, while depression in the northern North Sea already exceeded a few tens of meters. As ice sheets were expanding, the region of maximal forebulge uplift resulting from the UMISM and ICE-6G_C reconstructions in the southeastern North Sea basin started moving towards the present-day northern part of the Dutch-German border.

At 60 ka, all three available reconstructions show a well-developed British-Irish Ice Sheet, merging with the Fennoscandian Ice Sheet in the North Sea (Fig. 16). All reconstructions at that moment place the summit of the forebulge somewhere in the Netherlands. The reconstruction based on ICE-6G_C depicts a forebulge peak of over 4 m in the northernmost part of the Netherlands, while the other two reconstructions place a forebulge apex of over 5 m more in the central Netherlands. Uplift in the northern part of Belgium would still have been around 3 to 4 m. The hinge line between uplift and depression ran over the southeastern tip of England towards the Frisian Islands before deflecting back towards the (south)east in Germany. Depression in the North Sea region between this hinge line and the ice margin reached up to 50-100 m.

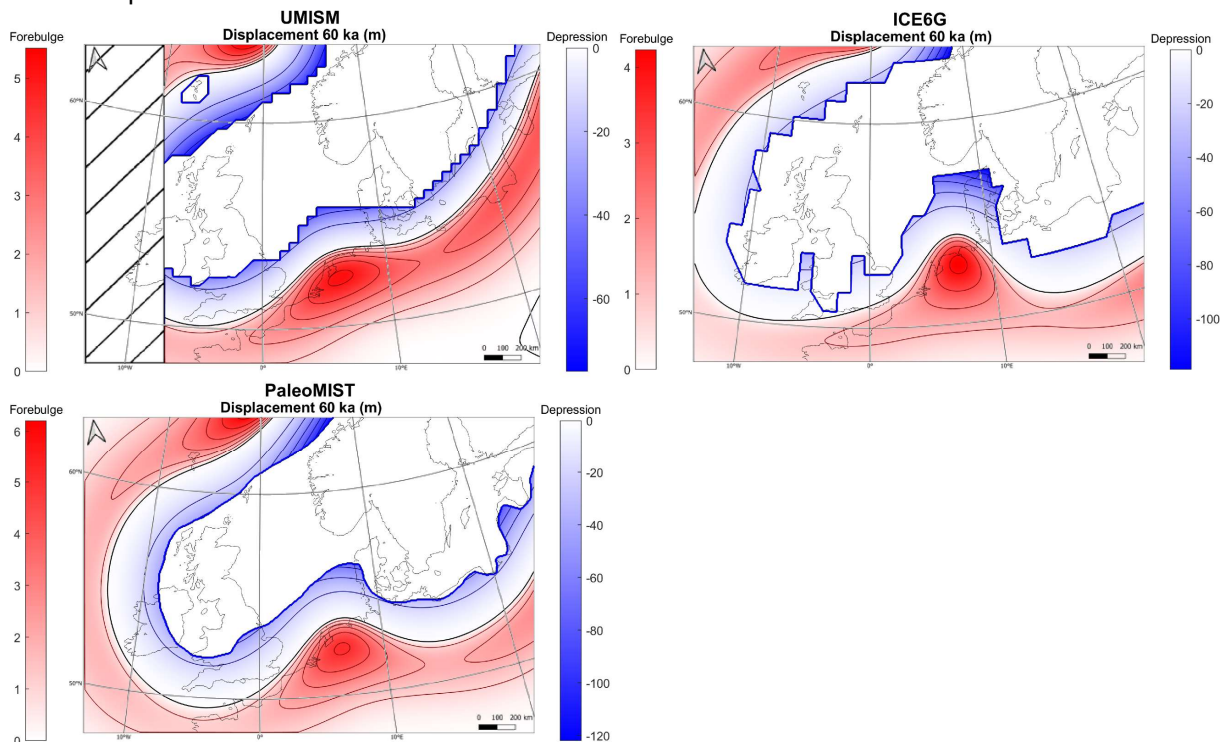


Figure 16: Reconstruction of the vertical displacement (m) at 60 ka based on the UMISM, ICE-6G_C and PaleoMIST 1.0 ice sheet models.

A reduced to non-existent BIIS and a strong reduction of the FIS until 30 ka according to UMISM and PaleoMIST 1.0 caused the forebulge to migrate again in a northwestward direction (and even again directly connecting with the forebulge north of Scotland over the North Sea in the case of PaleoMIST 1.0). These reconstructions also show an attenuation of the vertical uplift and depression. On the contrary, the ICE-6G_C reconstructions only display a minimal reduction in ice masses, resulting in very limited variations of the hinge line and of the magnitude of uplift and depression.

Starting from 30 ka, BRITICE-CHRONO reconstructions are available as well. All reconstructions indicate the presence of a BIIS, which was only still merged with the FIS according to ICE-6G_C (Fig. 17). While the forebulge summit was still located at the northern Dutch-German border for ICE-6G_C, the ice-free North Sea region in the other reconstructions resulted in the maximum forebulge uplift having been located at the Dutch and German offshore North Sea sectors.

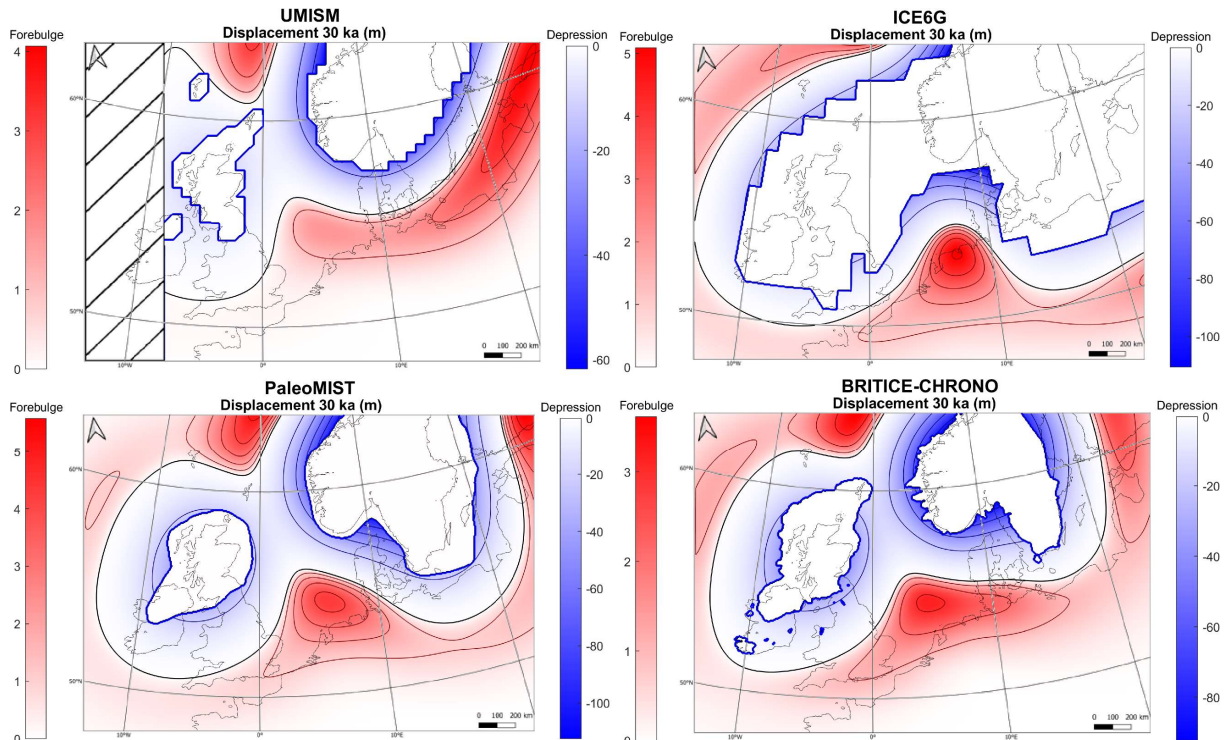


Figure 17: Reconstruction of the vertical displacement (m) at 30 ka based on the different ice sheet models.

Merging of the BIIS and FIS at 29 ka according to UMISM and PaleoMIST 1.0 and at 26 ka according to BRITICE-CHRONO caused a migration of the forebulge from the now inundated southeastern North Sea region towards the Netherlands and northwestern Germany.

Fig. 18 displays the evolution of the maximum forebulge over time for the available reconstructions since 30 ka. There were some fluctuations, but all reconstructions apart from UMISM show rather stable forebulge elevations in the range of 3 to 6 m of uplift until 15 ka. The only exception is UMISM, according to which the BIIS was larger in both extent and thickness compared to the other reconstructions, remained quite voluminous for a significant period of time and only disconnected from the FIS at 14 ka. The forebulge based on UMISM increased quite steadily between 30 and 20 ka and even experienced an acceleration in uplift after 20 ka. All available reconstructions show a steady decrease in forebulge uplift since 15 ka until the youngest available reconstructions at 8 ka, when the forebulge accounted for 1 to 2 m of uplift.

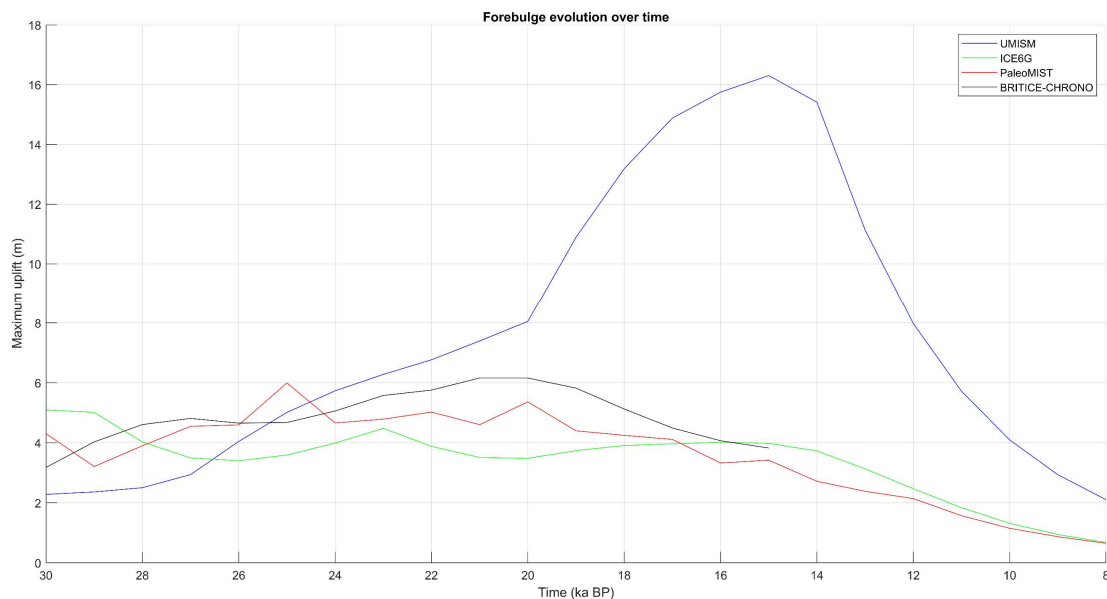


Figure 18: Evolution of maximum uplift over time between 30 and 8 ka for the different reconstructions.

At 20 ka, BRITICE-CHRONO is the first reconstruction to show a disconnection between BIIS and FIS (Fig. 19). The forebulge hinge line again passed through the southeastern tip of England for PaleoMIST 1.0 and BRITICE-CHRONO and slightly more to the south for UMISM and ICE-6G_C. This uplift-depression transition was also relocated over the Frisian Islands and deflected towards the southeast into Germany. Only for UMISM the hinge line was located south of the Frisian Islands. The Netherlands again experienced the largest amount of uplift. Even though the ice sheet configurations differ between the reconstructions, close to the southern ice margin in the North Sea region the land was depressed up to about 100 m.

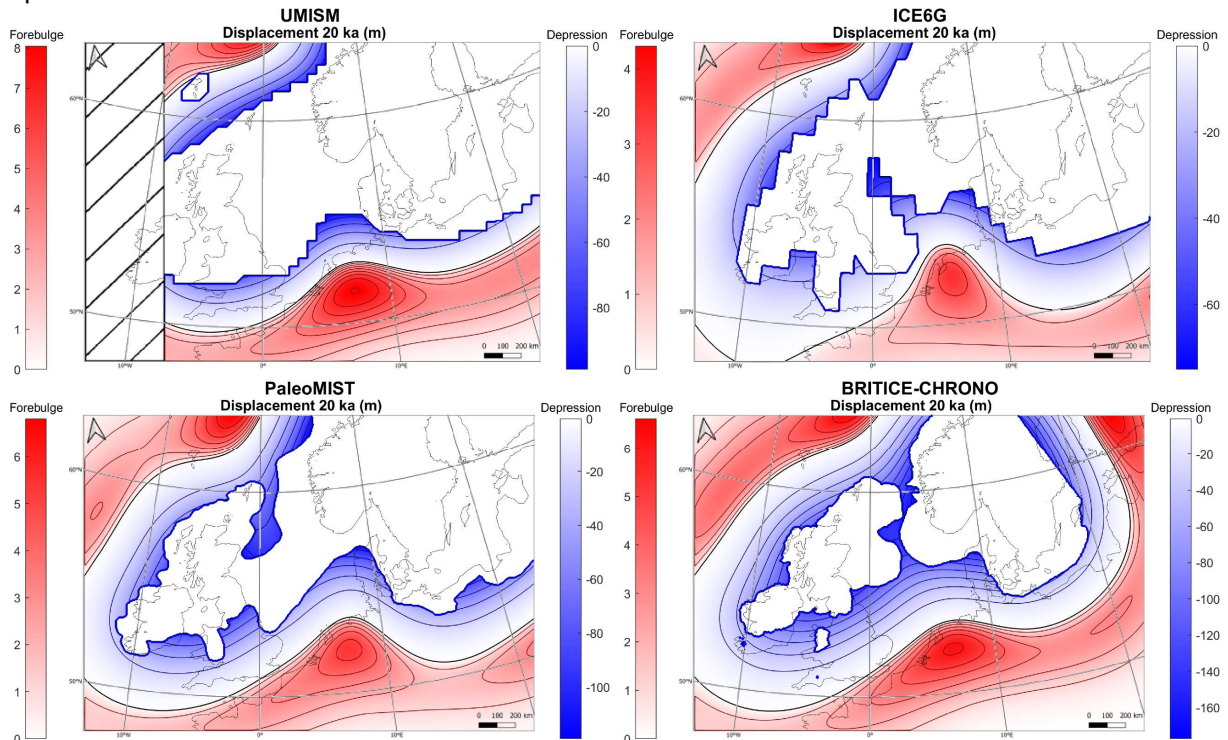
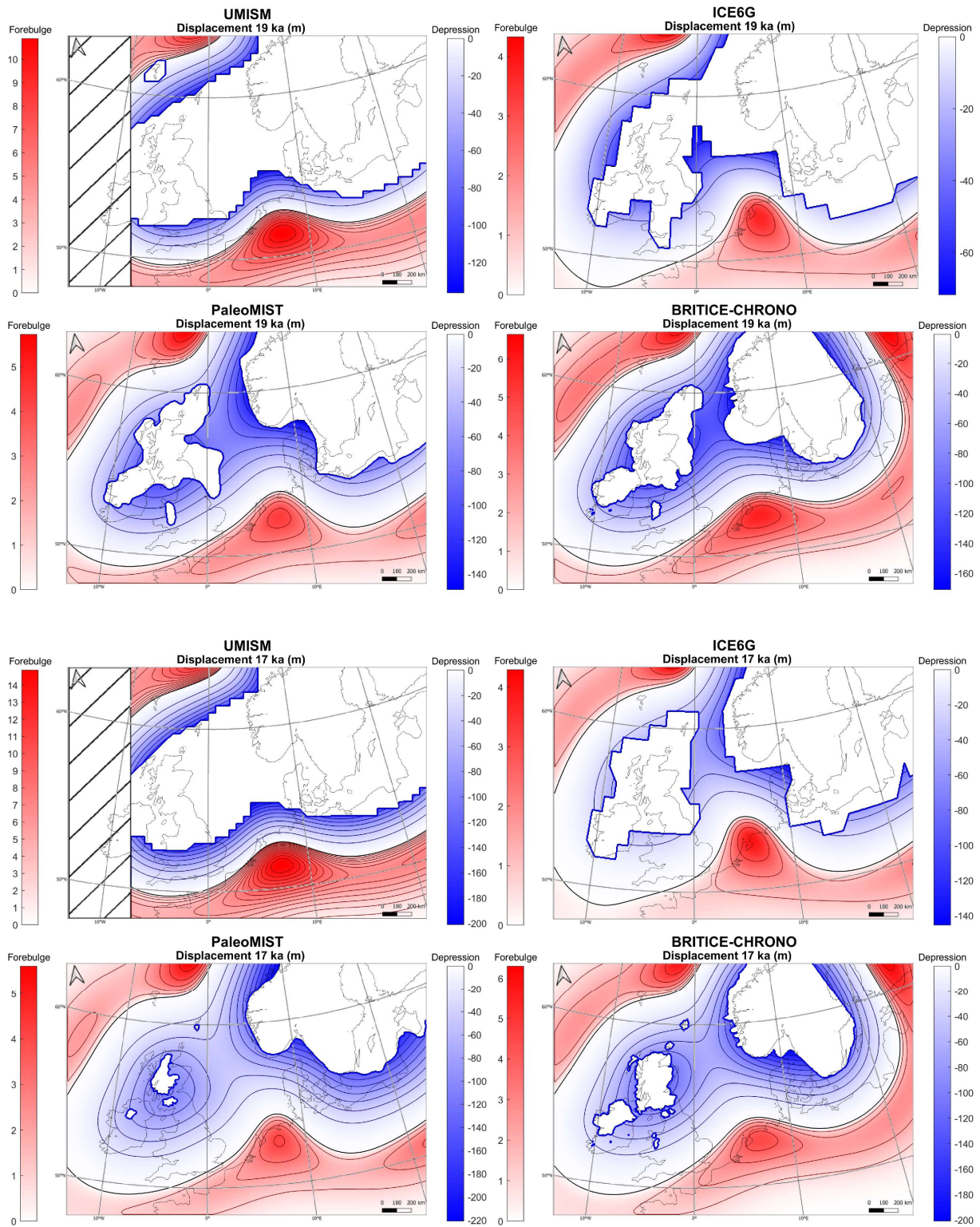


Figure 19: Reconstruction of the vertical displacement (m) at 20 ka based on the different ice sheet models.

Disconnection of the BIIS and FIS occurred around 19 ka according to PaleoMIST 1.0, at 18 ka according to ICE-6G_C and as already mentioned, only at 14 ka according to UMISM (Fig. 20). Unzipping of the two ice sheets and the rapid retreat of the ice masses roughly since 20 ka again resulted in a migration of the hinge line in the direction of the offshore sectors from the Netherlands and Germany. This seaward retreat is the most obvious for the PaleoMIST 1.0 and ICE-6G_C based reconstructions. Once the ice was removed from the study-area, a lateral migration of the uplift-depression transition is no longer observed. At this moment, the lithosphere only continued to adjust vertically to the new ice-free equilibrium state and is expected to exponentially relax towards zero vertical displacement. However, at this moment sea level rise will have gained in importance in the study area and the lithosphere will have started to experience hydro-isostatic adjustment.



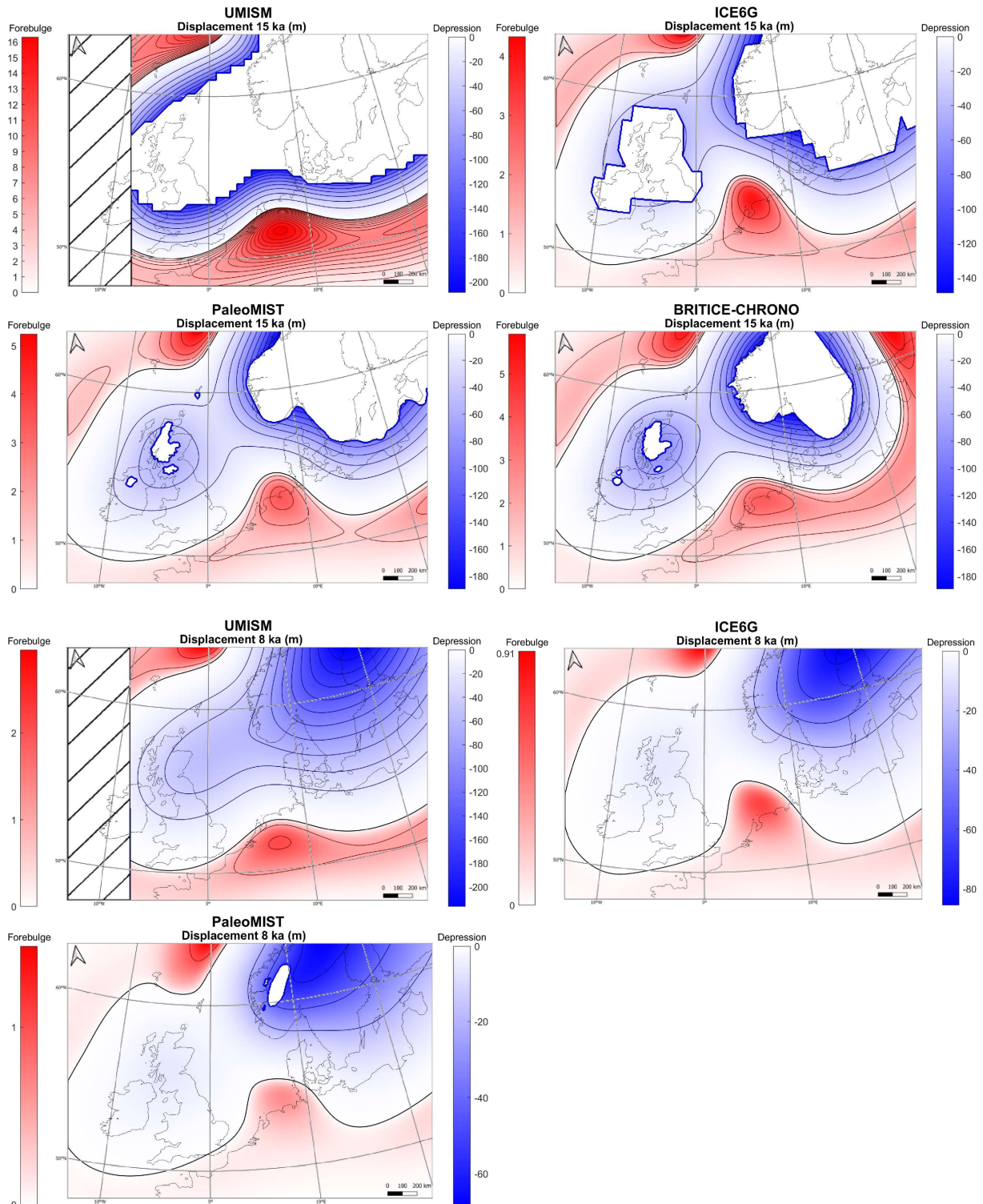


Figure 20: Reconstruction of the vertical displacement (m) at 19, 17, 15 and 8 ka based on the different ice sheet models.

5.5 Paleotopographic reconstructions of the southern North Sea region throughout the last glacial period

Throughout the last glacial period, the present-day North Sea was (partly) emerged. A map of the current bathymetry (Fig. 21) can be compared to the paleotopography maps created based on the different GIA-reconstructions (Fig. 22 to Fig. 28), for which the same stepwise rainbow colour scale is used.

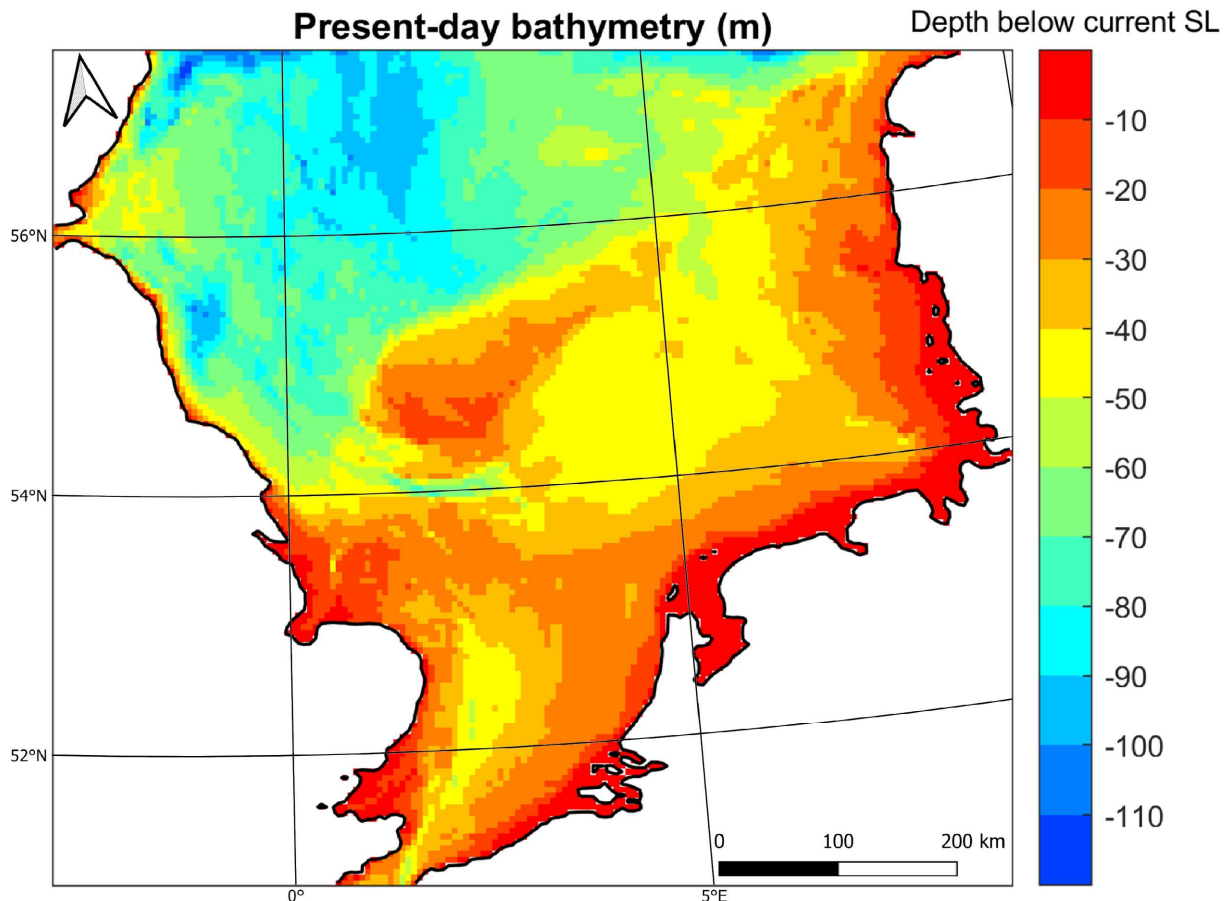
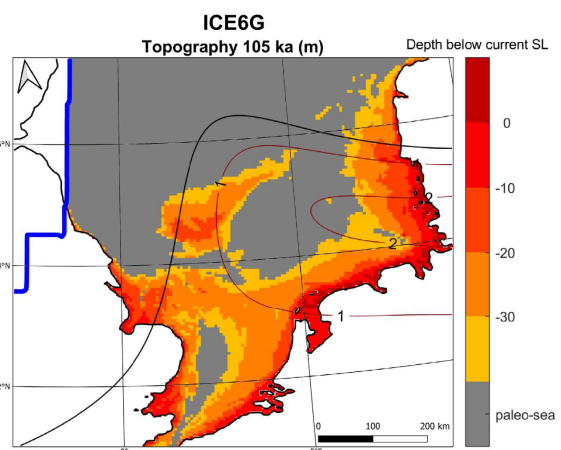
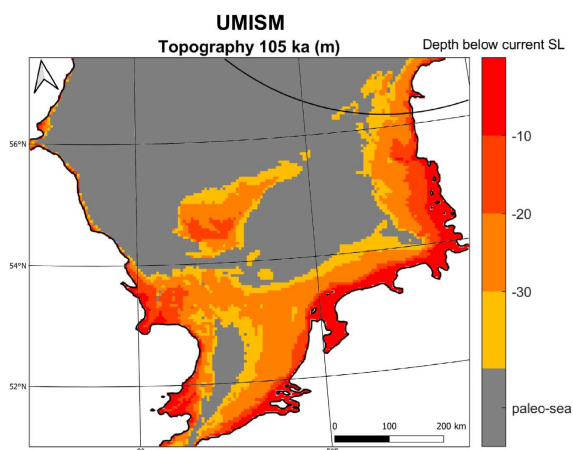
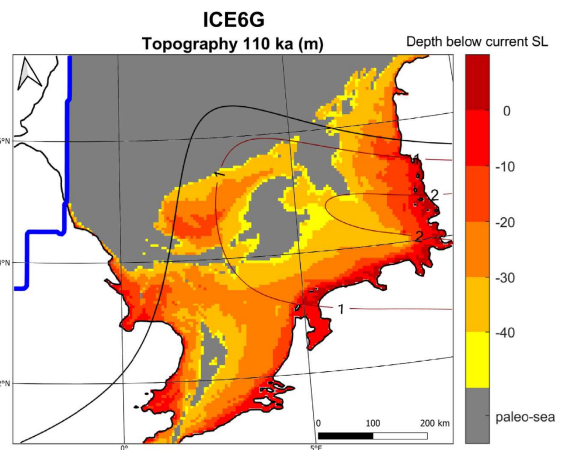
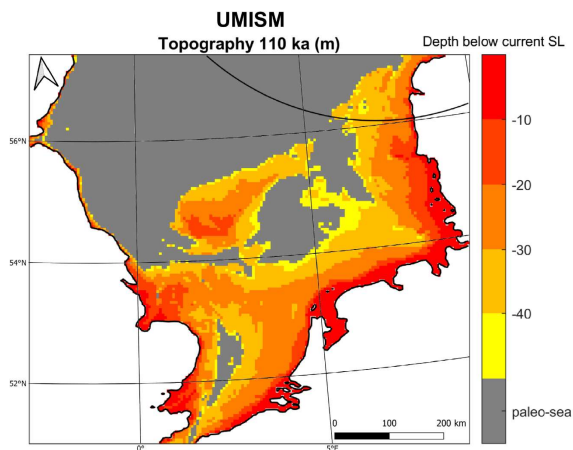
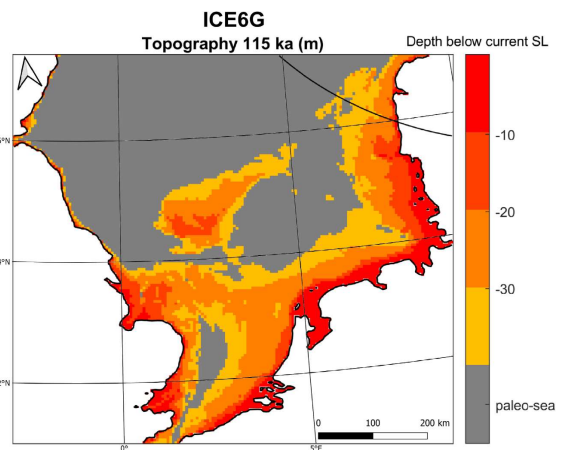
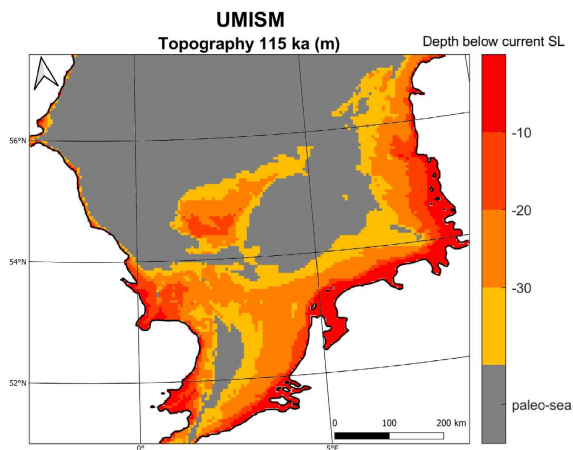
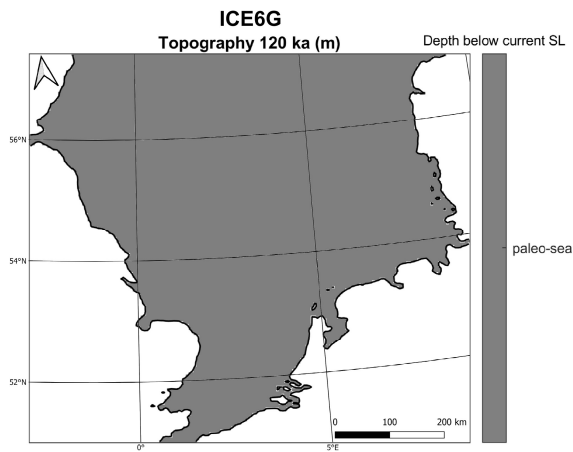
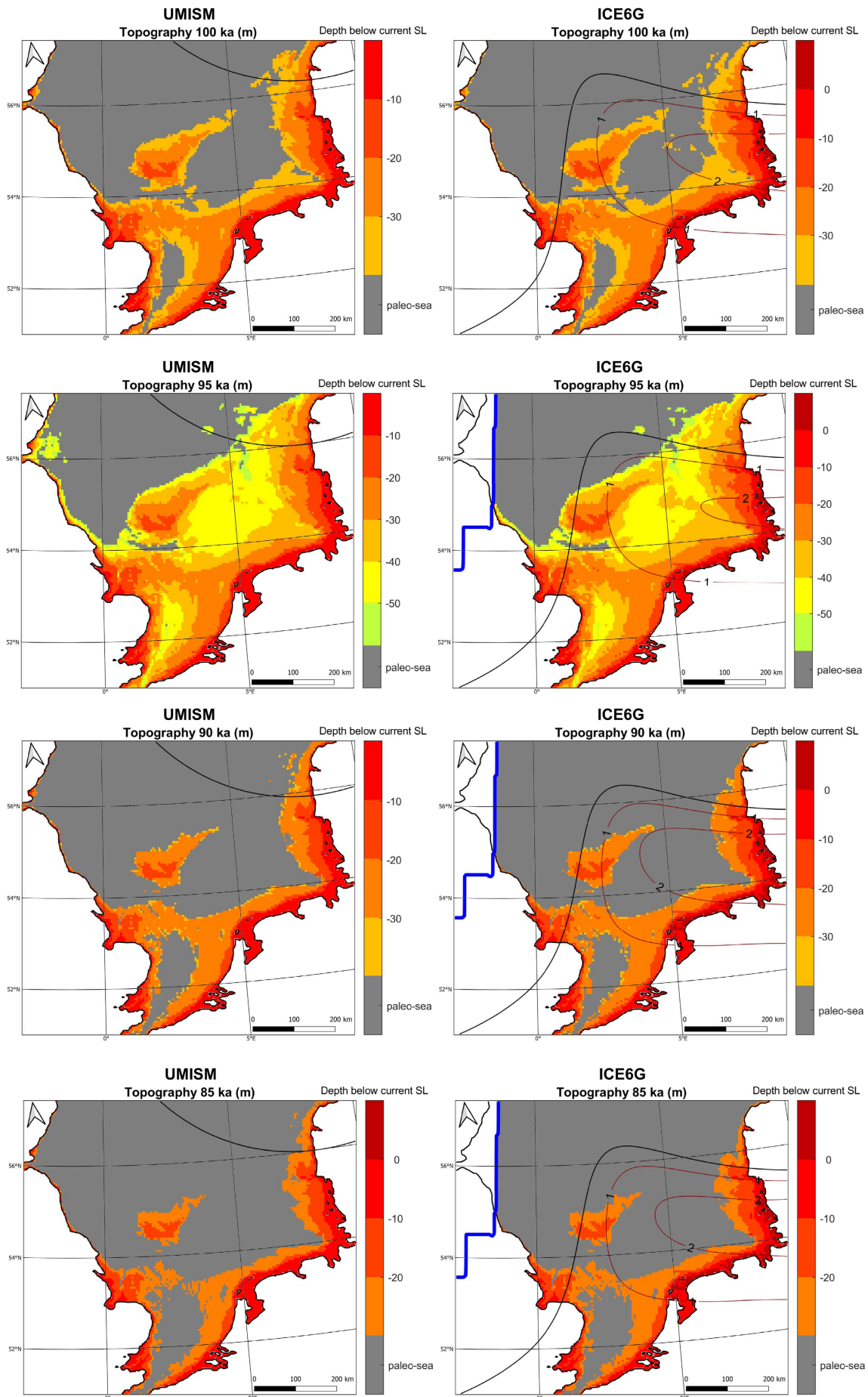


Figure 21: Current bathymetry map of the North Sea region covering the same area and using the same spatial resolution and colour scale as the paleotopography maps.

Paleotopography maps of MIS 5 mainly show a reflection of changes in eustatic sea level (Fig. 22). The forebulge according to ICE-6G_C only slightly exceeded the 2 m uplift contour, located in the offshore German North Sea sector. According to UMISM, uplift in the region of interest never exceeded 1 m during MIS 5. The region of interest experienced more than 20 m depression only for the first time during the last glacial period at 70 ka (Fig. 23). At 120 ka, the sea level was a few meters higher than today, resulting in a completely inundated North Sea (Fig. 22). During the following time slices, the Southern Bight was almost completely emerged. Only the surface of the Axial Channel was still located below sea level, but whether this area was inundated or not depends on to what amount its apparent disconnection from the inundated parts of the North Sea could be attributed to a Holocene sedimentary cover. Depending on the exact time slice, a zone of roughly 100 km offshore the current Dutch, German and Danish coasts was emerged as well, and so was the Dogger Bank. Due to a drop in sea level at 95 ka the entire area east of Dogger Bank suddenly became emerged, while the base of the Outer Silver Pit was still located below sea level. Sea level rose again after 95 ka, inundating a large part of the Southern Bight and of the eastern side of the southern North Sea. Another sea level drop at 80 ka created a situation very similar to the 95 ka configuration, but with the Axial Channel being located above sea level as well. At 75 ka, sea level had already dropped 60 m compared to the interglacial situation of today, further emerging some limited areas along the British coast and northeast of Dogger Bank (Fig. 22). This was also the first moment during the last glacial period when the study area experienced the presence of ice. The BIIS entered the North Sea region from the west according to UMISM and ICE-6G_C, while UMISM also displays ice from the FIS in northern Denmark and even ice on top of Dogger Bank. PaleoMIST 1.0 does not yet indicate ice in the region of interest at 75 ka.





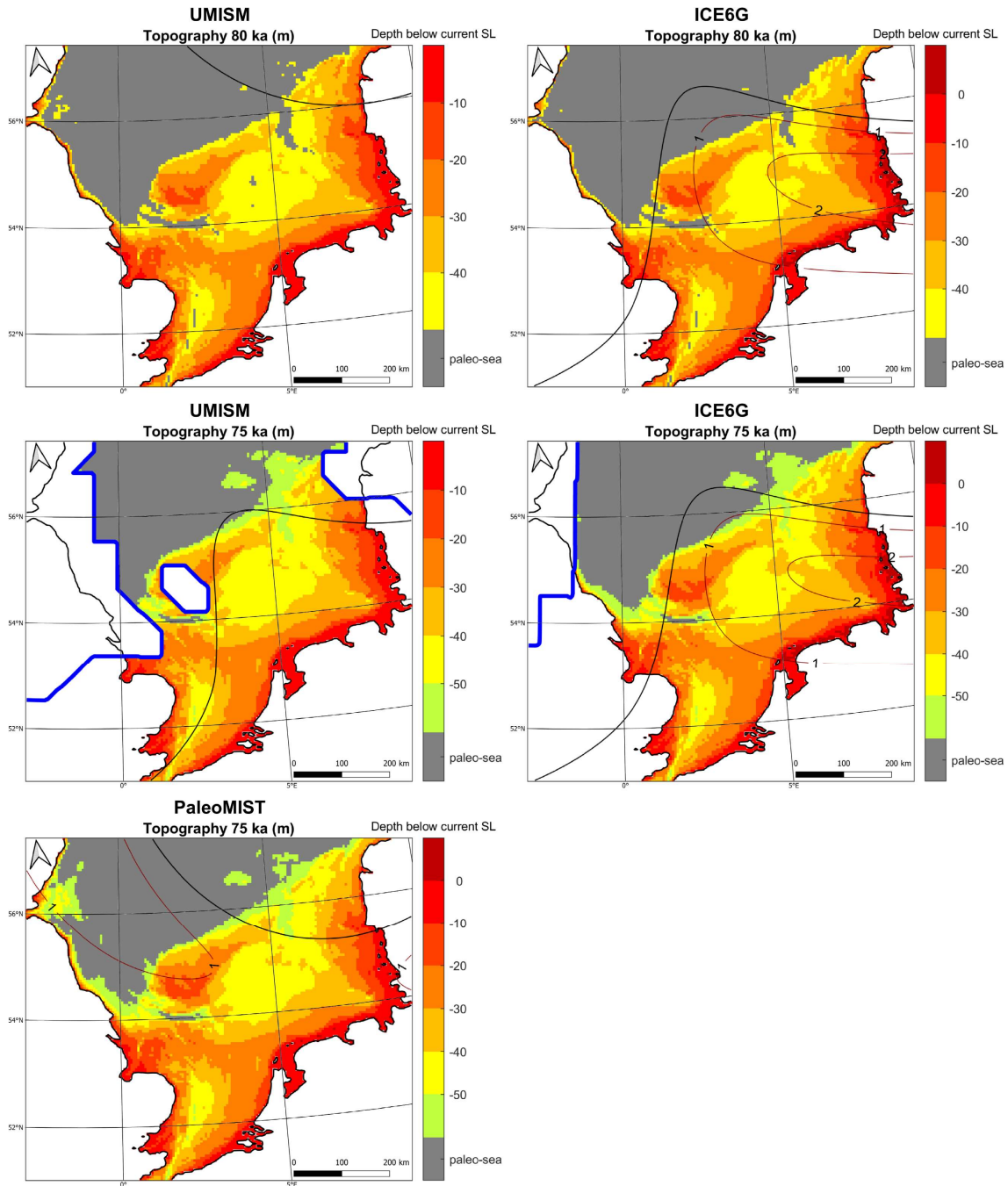


Figure 22: Evolution of the topography of the North Sea basin from 120 ka to 75 ka based on the current bathymetry and the GIA as calculated based on the UMISM and ICE-6G_C (120 ka to 75 ka) and PaleoMIST 1.0 (75 ka) ice sheet models.

At 70 ka, well into MIS 4, glacial influence started to gain in importance in the study area (Fig. 23). While PaleoMIST 1.0 and ICE-6G_C show no expansion of ice in the region, the approaching ice sheets now for the first time induced a depression exceeding 20 m in the northernmost regions of the study area. According to UMISM, the BIIS and FIS were even connected via a narrow ice bridge, covering the Dogger Bank and large areas west, north and northeast of it. However, where the connection was the narrowest, ice thickness only amounted up to a few hundred meters.

5 kyr later, ICE-6G_C indicates a similar situation as was shown for UMISM at 70 ka, with ice thicknesses of approximately 350 m at the narrowest zone of connection (Fig. 23). The ice bridge, however, temporarily disappeared according to UMISM, and PaleoMIST still shows no ice in the study area. A further sea level drop of ca. 15 m caused the emersion of certain areas north of Dogger Bank.

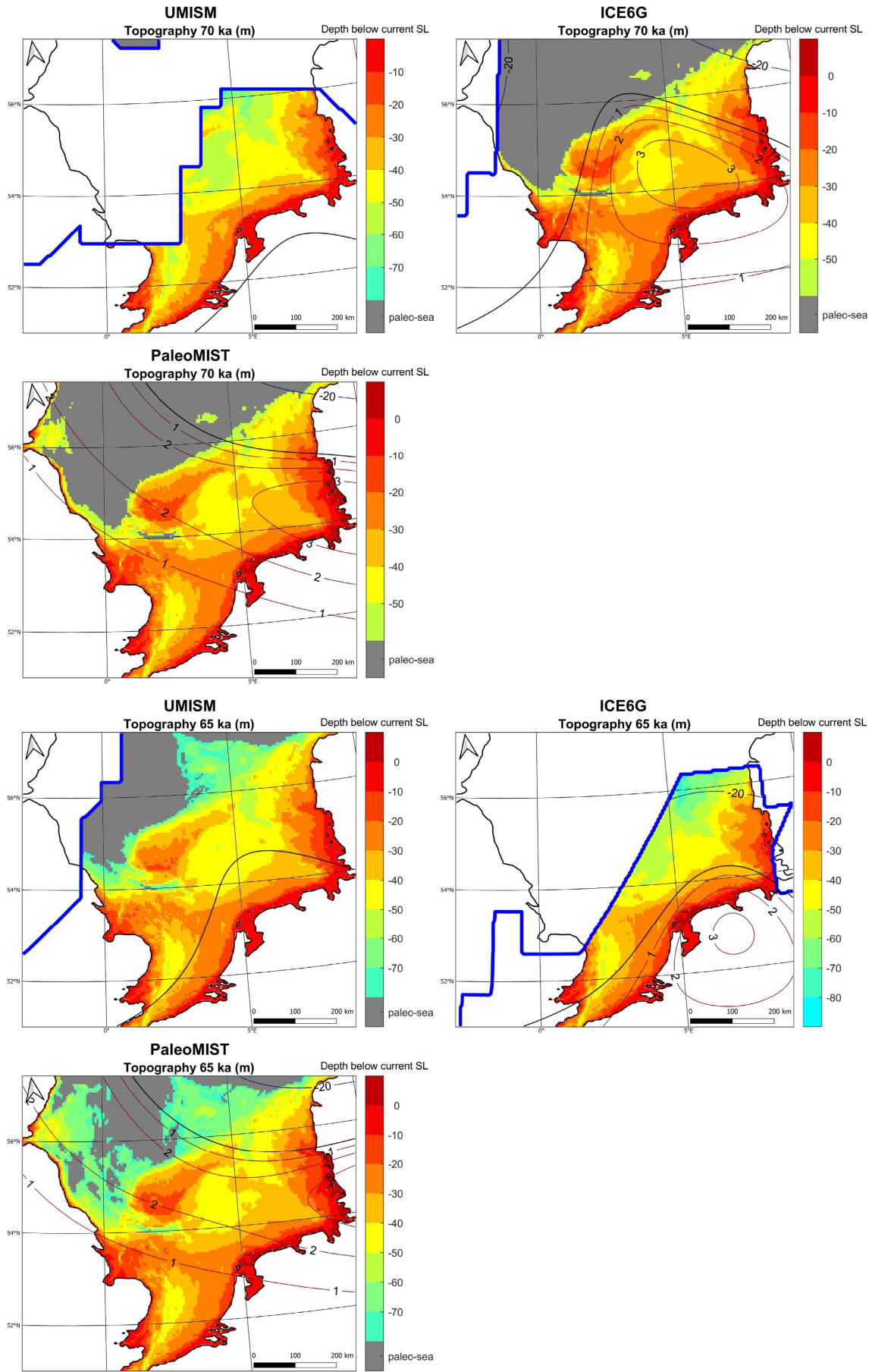


Figure 23: Evolution of the topography of the North Sea basin from 70 ka to 65 ka based on the current bathymetry and the GIA as calculated based on the UMISM, ICE-6G_C and PaleoMIST 1.0 ice sheet models.

At 60 ka, all reconstructions indicate the study area being subjected to full glacial conditions (Fig. 24). Extensive ice sheets on the British Isles and on Scandinavia were connected, covering a large part of the North Sea with ice. The southern ice limit roughly ran in a SW-NE direction over the southern edge of Dogger Bank. As mentioned earlier, the Netherlands experienced GIA-related uplift of ca. 5 m, but also the surrounding regions of northern France, Belgium and a part of the current German coast experienced uplift. The UMISM and PaleoMIST 1.0 reconstructions indicate the ice margin running over central Denmark. Right south of this margin, GIA-related depressions of over 40 m and 60 m respectively caused the land to be located 100 m below the current sea level. The ICE-6G_C ice margin only ran over the northern tip of Denmark, and with a depression of over 80 m the land over there was located almost 200 m below the current sea level.

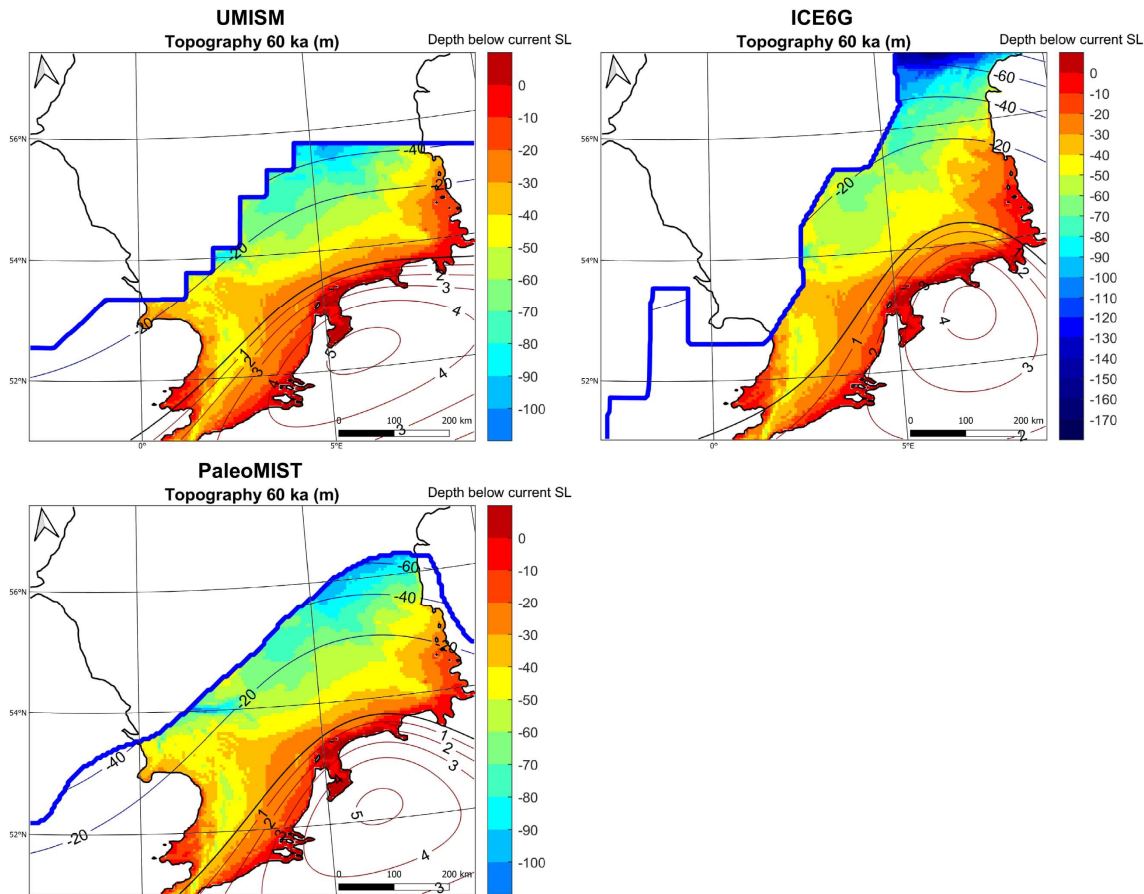
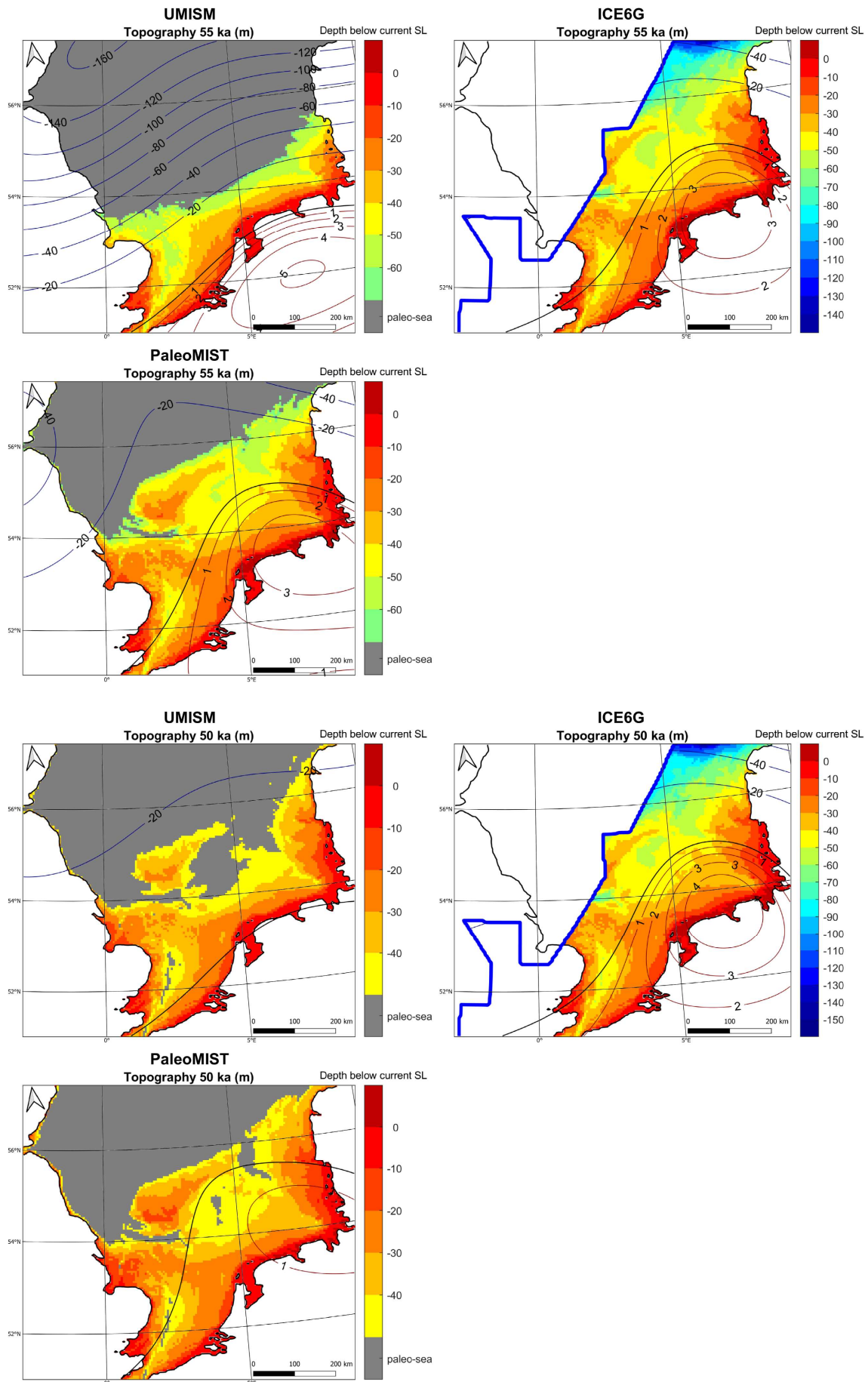
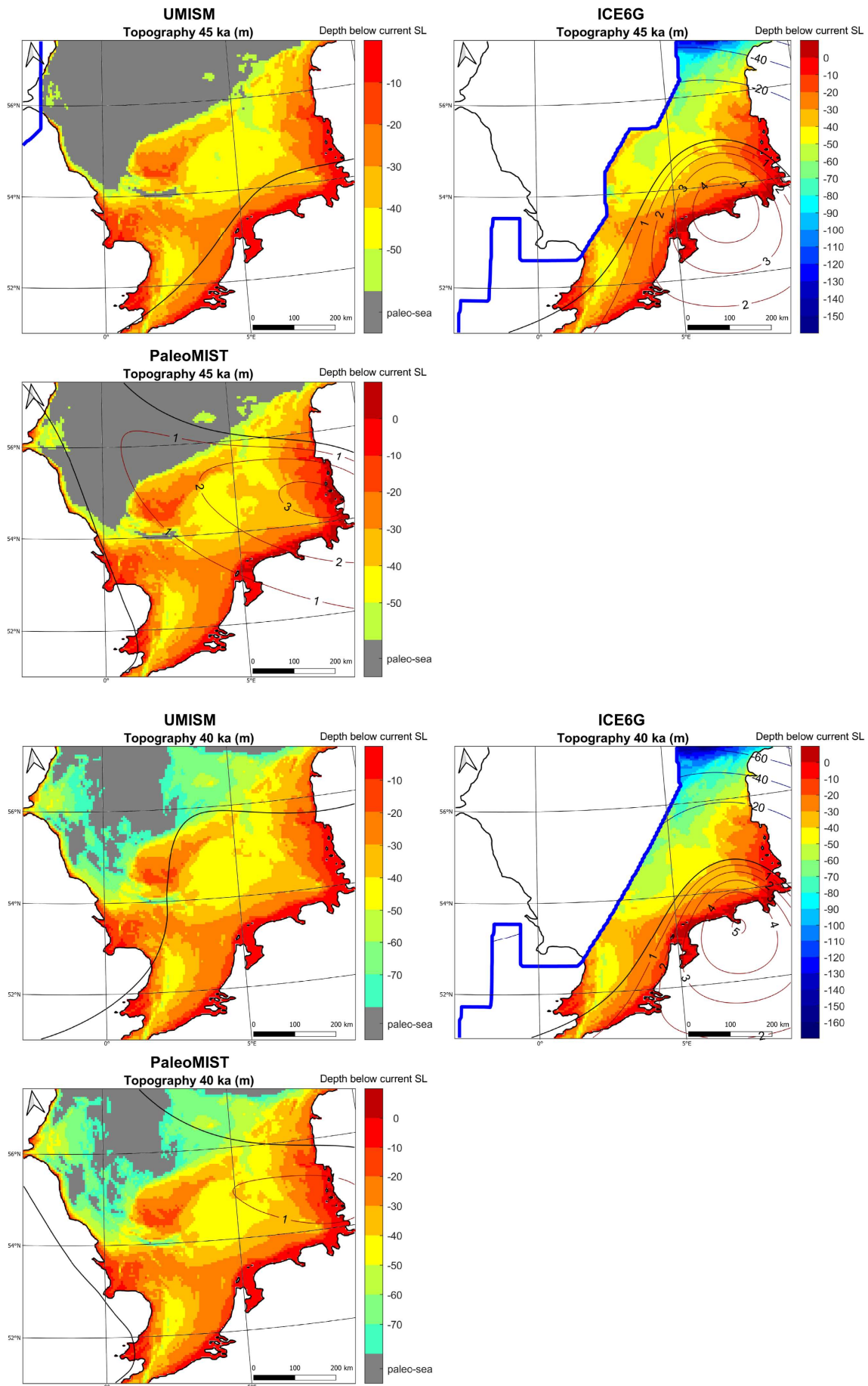


Figure 24: Topography of the North Sea basin at 60 ka based on the current bathymetry and the GIA as calculated based on the UMISM, ICE-6G_C and PaleoMIST 1.0 ice sheet models.

At 55 ka, the situation had changed dramatically according to UMISM and PaleoMIST. Influence of ice in the North Sea region had completely disappeared as a BIIS was non-existent (Fig. 25), and a significant ice sheet on the British Isles would only appear again around 30 ka. The same situation as was already observed earlier was achieved according to PaleoMIST 1.0; the coastline ran over a SW-NE axis north of Dogger Bank. Due to a much thicker BIIS-FIS connection according to UMISM, the North Sea region was depressed more than 100 m in certain areas, causing the inundation of Dogger Bank and Oyster Ground. In contrast to the ice retreat according to UMISM and PaleoMIST 1.0, glacial conditions only attenuated a little bit according to ICE_6G_C, and its situation remained similar until 30 ka, with only a small weakening of the GIA-related depression. By 50 ka, the forebulge according to UMISM and PaleoMIST 1.0 had almost completely disappeared, and Dogger Bank started to re-emerge. 5 kyr later, the North Sea depression due to the former ice bridge was no longer noticeable in the UMISM reconstruction, resulting in a completely emerged Dogger Bank and Oyster Ground. A sea level drop at 40 ka again emerged extensive regions north of Dogger Bank.





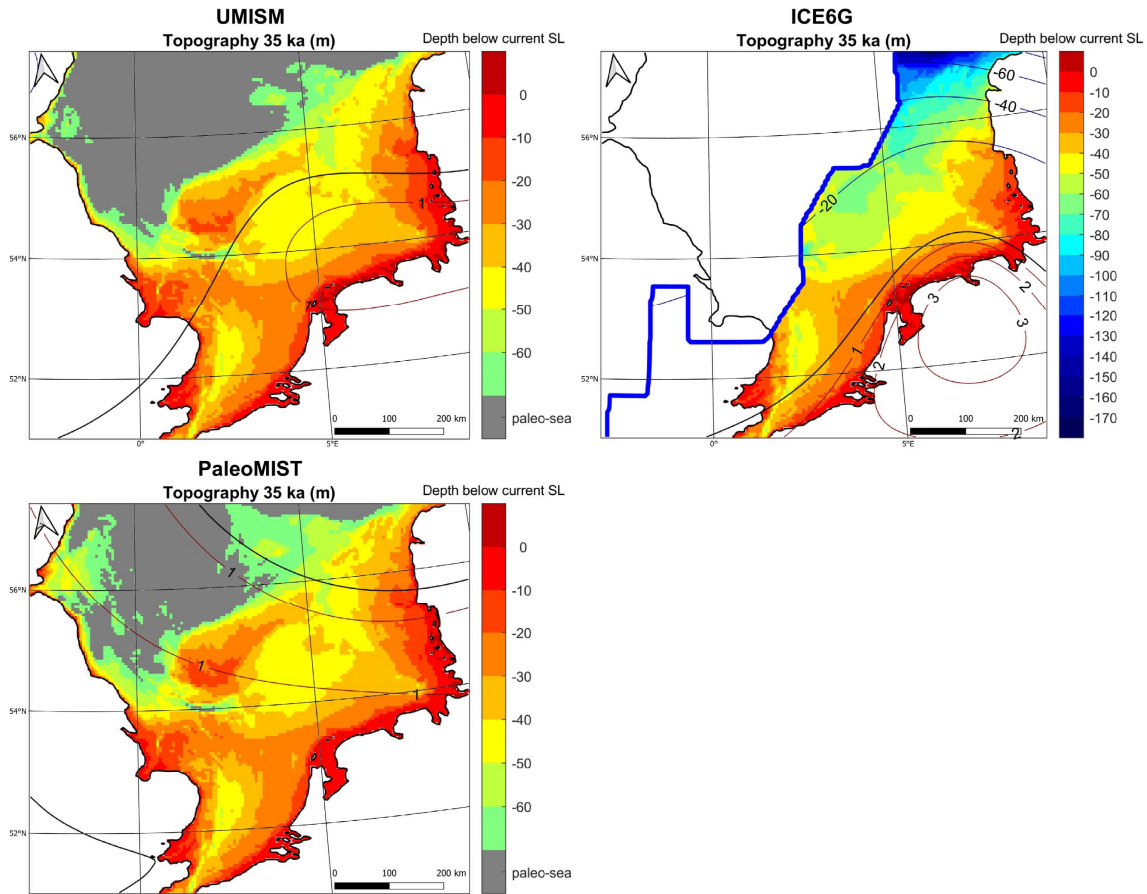
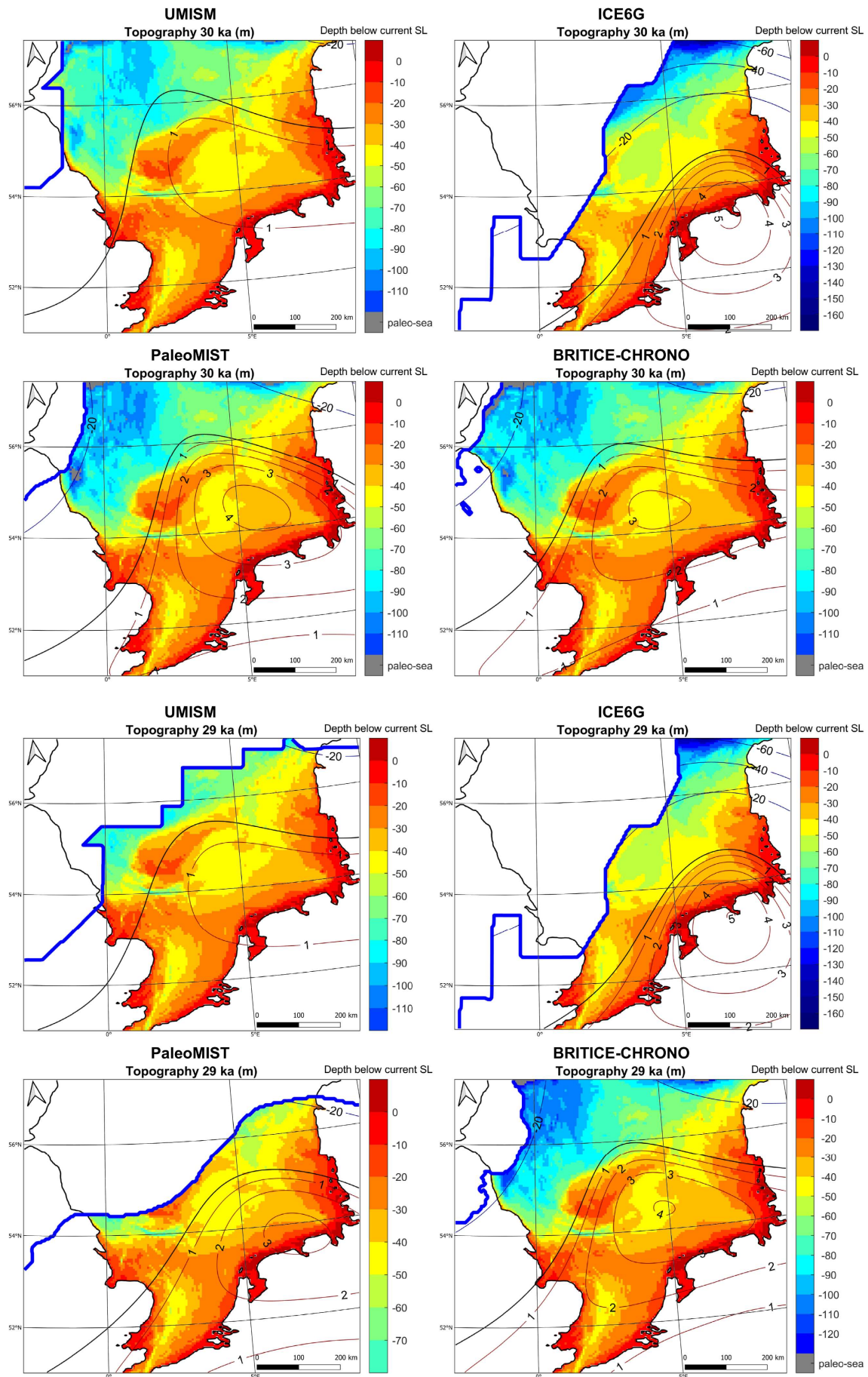


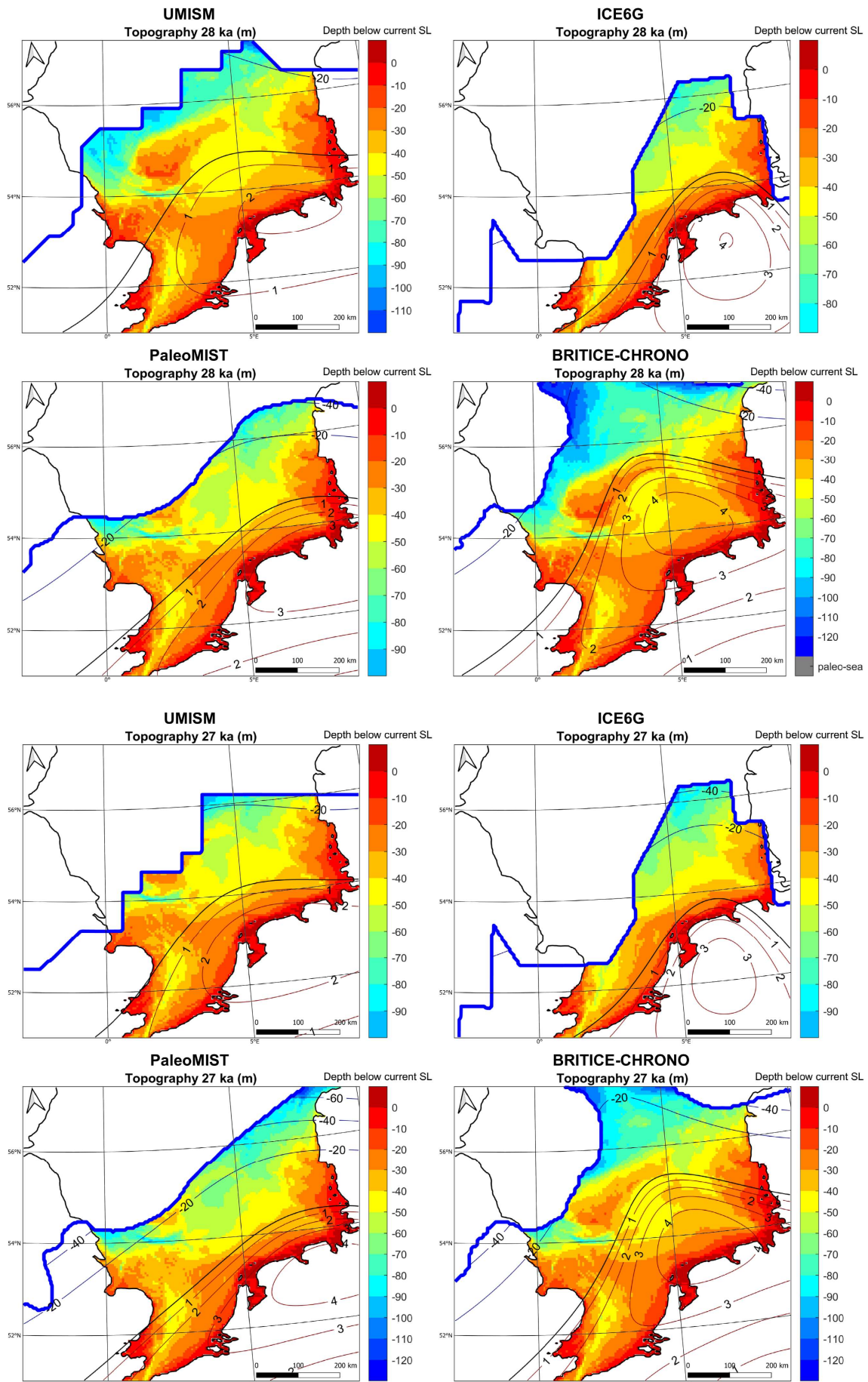
Figure 25: Evolution of the topography of the North Sea basin from 55 ka to 35 ka based on the current bathymetry and the GIA as calculated based on the UMISM, ICE-6G_C and PaleoMIST 1.0 ice sheet models.

At 30 ka, the eustatic sea level was more than 100 m below the current sea level for the first time during the last glacial period, and would remain at least that low until ca. 15 ka. This resulted in the area of interest almost completely being emerged (Fig. 26). The UMISM, PaleoMIST 1.0 and BRITICE-CHRONO reconstructions indicate the presence of a BIIS, still disconnected from the FIS. According to ICE-6G_C, the same ice bridge that was established around 65 ka still connected the two ice sheets at 30 ka. A hinge line already quite close to the present-day coast caused the Oyster Ground and the ice-free part of Dogger Bank to be more depressed according to ICE-6G_C. A smaller forebulge according to UMISM resulted in the Oyster Ground in turn having been less elevated compared to PaleoMIST 1.0 and BRITICE-CHRONO.

At 29 ka, the BIIS and FIS merged according to UMISM and PaleoMIST 1.0, with the SW-NE oriented southern ice margin located north of or passing over Dogger Bank, respectively (Fig. 26). This ice margin would gradually move southwards for the next 7 to 8 kyr. The ICE-6G_C ice margin started to move southwards as well, and also the not-yet merged BIIS and FIS started to occupy larger areas in the region of interest.

When the two ice sheets merged at 26 ka according to BRITICE-CHRONO, the forebulge resulting from all reconstructions had already remarkably shifted in a southeastward direction, amongst others resulting in a depressed Dogger Bank and Oyster Ground area (Fig. 26). Maximum GIA-related depression in the northernmost ice-free regions now counted up to 40 to 60 m, so that this southern North Sea region at this moment was already at least 100 m below current sea level.





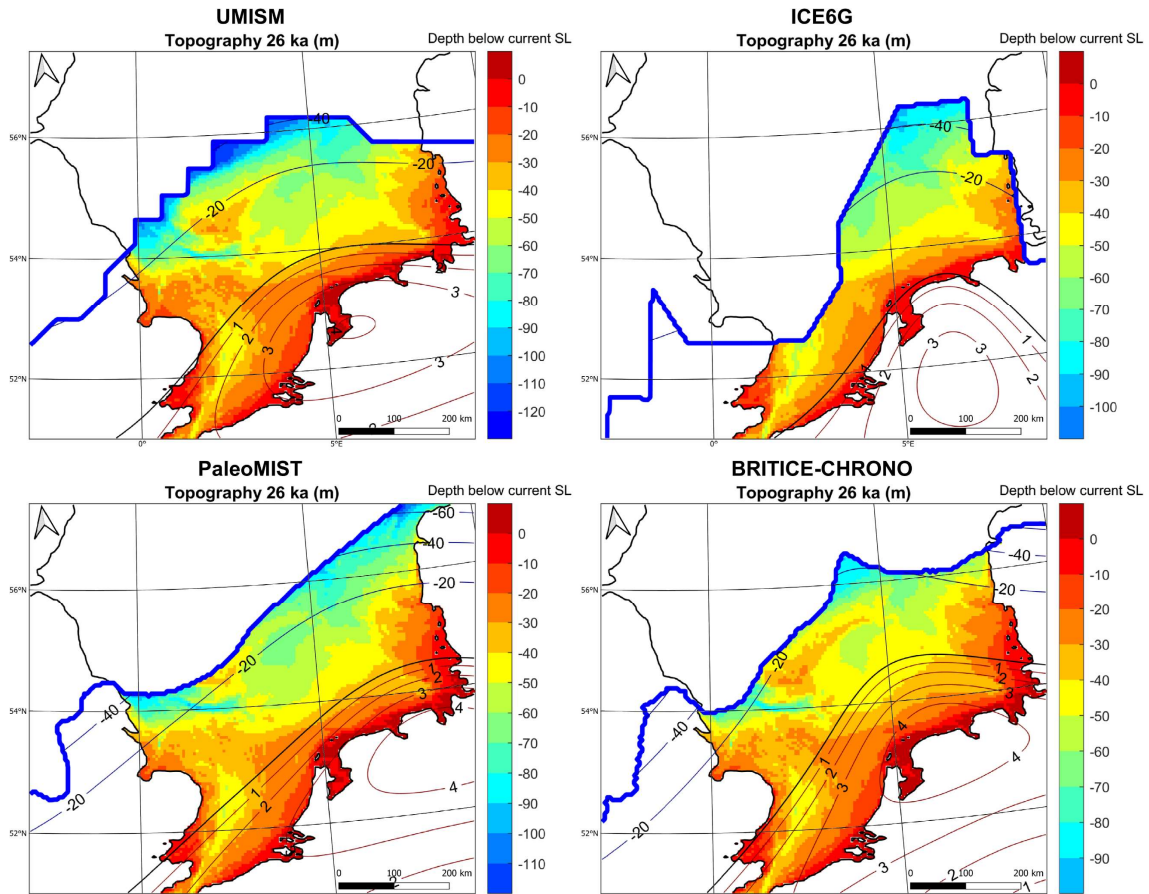
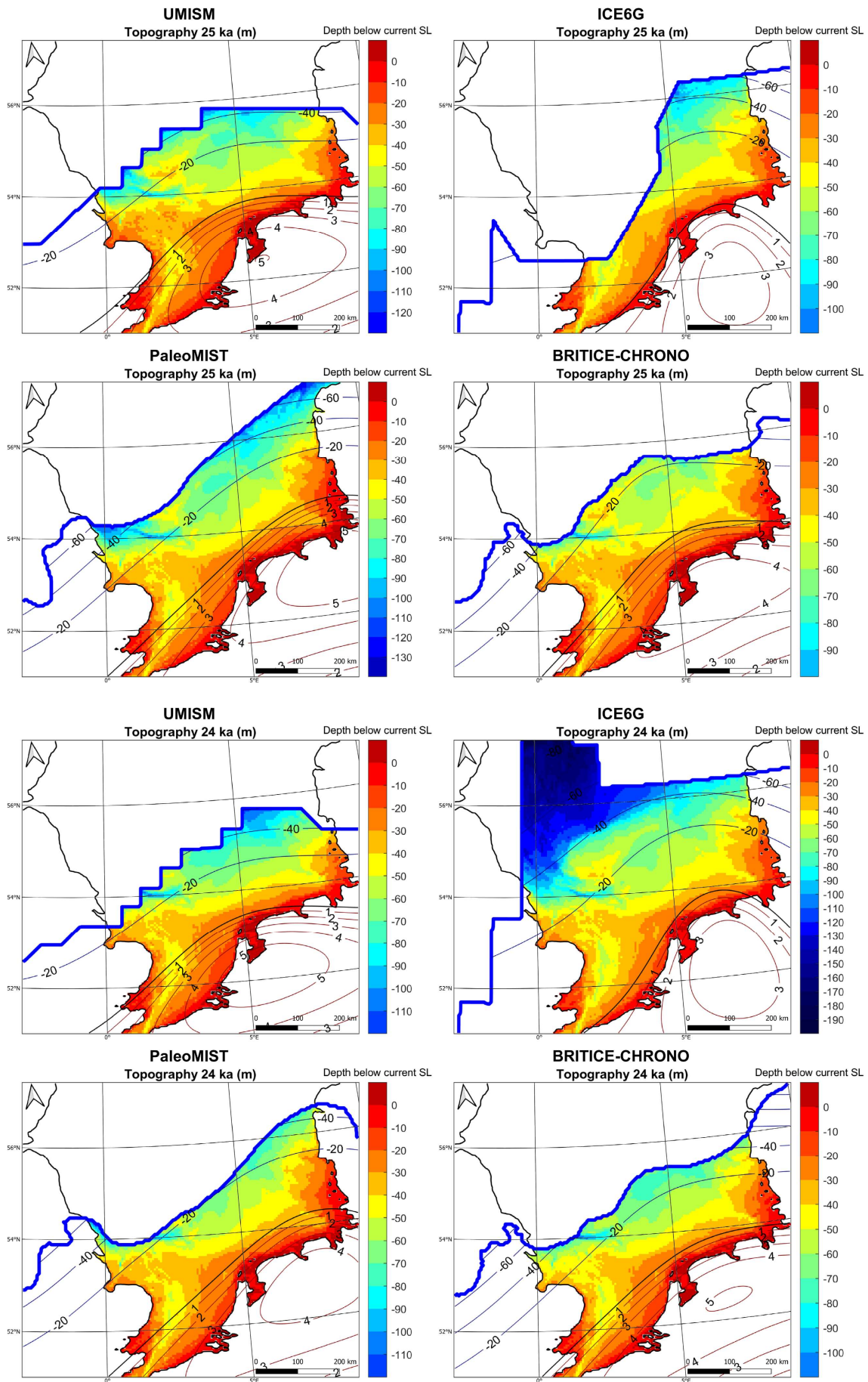
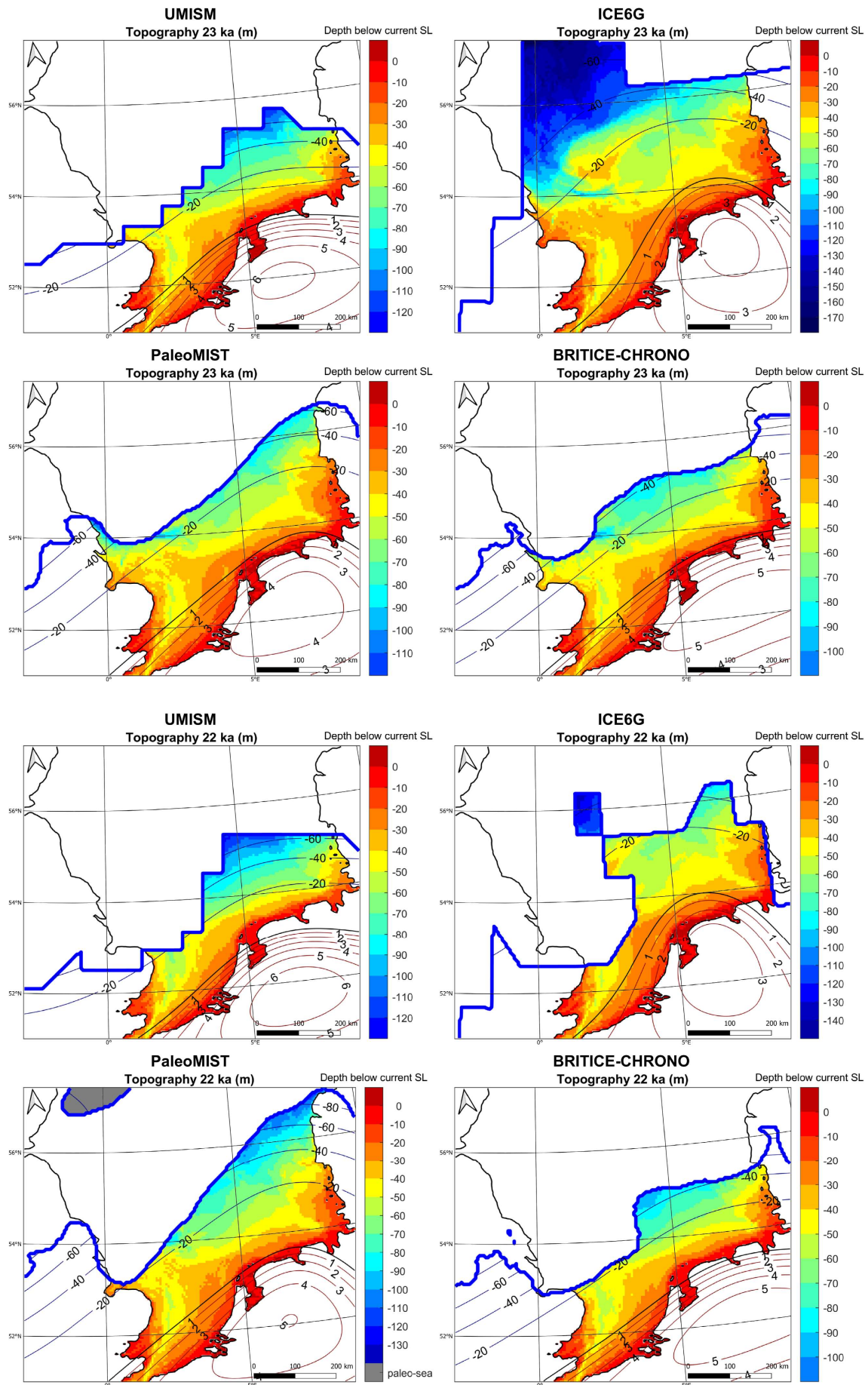


Figure 26: Evolution of the topography of the North Sea basin from 30 ka to 26 ka based on the current bathymetry and the GIA as calculated based on the UMISM, ICE-6G_C, PaleoMIST 1.0 and BRITICE-CHRONO ice sheet models.

The ice-volume peak was reached earliest for the ICE-6G_C reconstruction, and retreat was already observed after 25 ka (Fig. 27). At the 23 ka BRITICE-CHRONO ice-volume maximum, all reconstructions apart from ICE-6G_C indicate a very southerly located ice-margin, covering the entire Dogger Bank region. The barrier between the Axial Channel and the North Sea just north of the Southern Bight became more depressed due to the GIA-response of expansion of the ice sheets towards the south. Directly south of this ice margin a depression of at least 100 m below current sea level could still be observed. On the contrary, ICE-6G_C indicates a much larger depression north of Dogger Bank, due to the ice retreat in that area. At 22 ka, a glacial re-advance occurred according to ICE-6G_C, while the other reconstructions only show a minimal southward expansion of the ice and a slightly more depressed southern North Sea basin (Fig. 27). Ice retreated at 21 ka according to BRITICE-CHRONO, suddenly creating an ice-free depression of at least 170 m below present-day sea level. This retreat continued at 20 ka, when BRITICE-CHRONO is the first reconstruction to indicate the unzipping of the BIIS and FIS, having allowed seawater originating from the north to inundate the deeply depressed area northwest of Dogger Bank or to mix with the proglacial lake water that might have been present there.





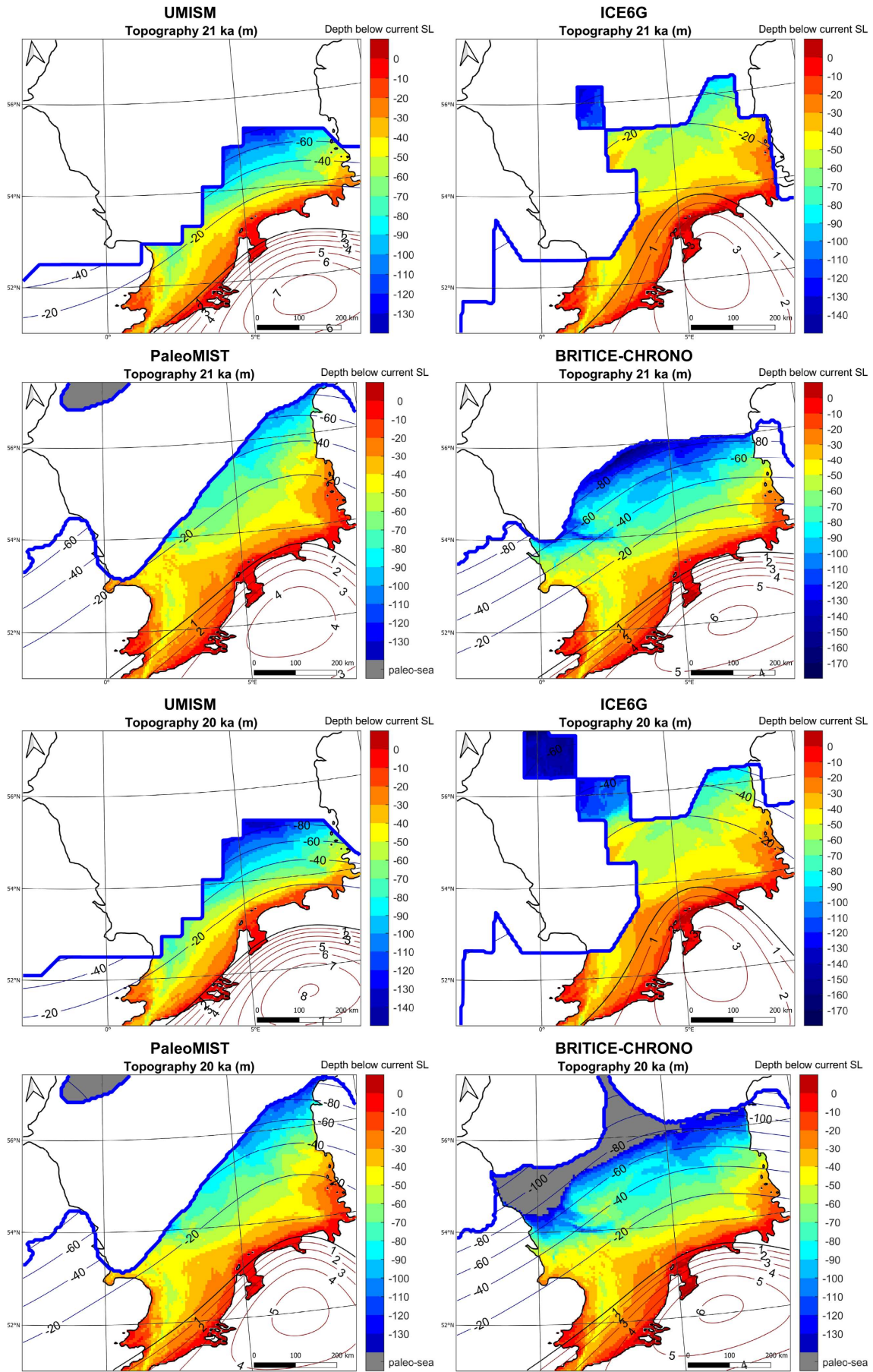
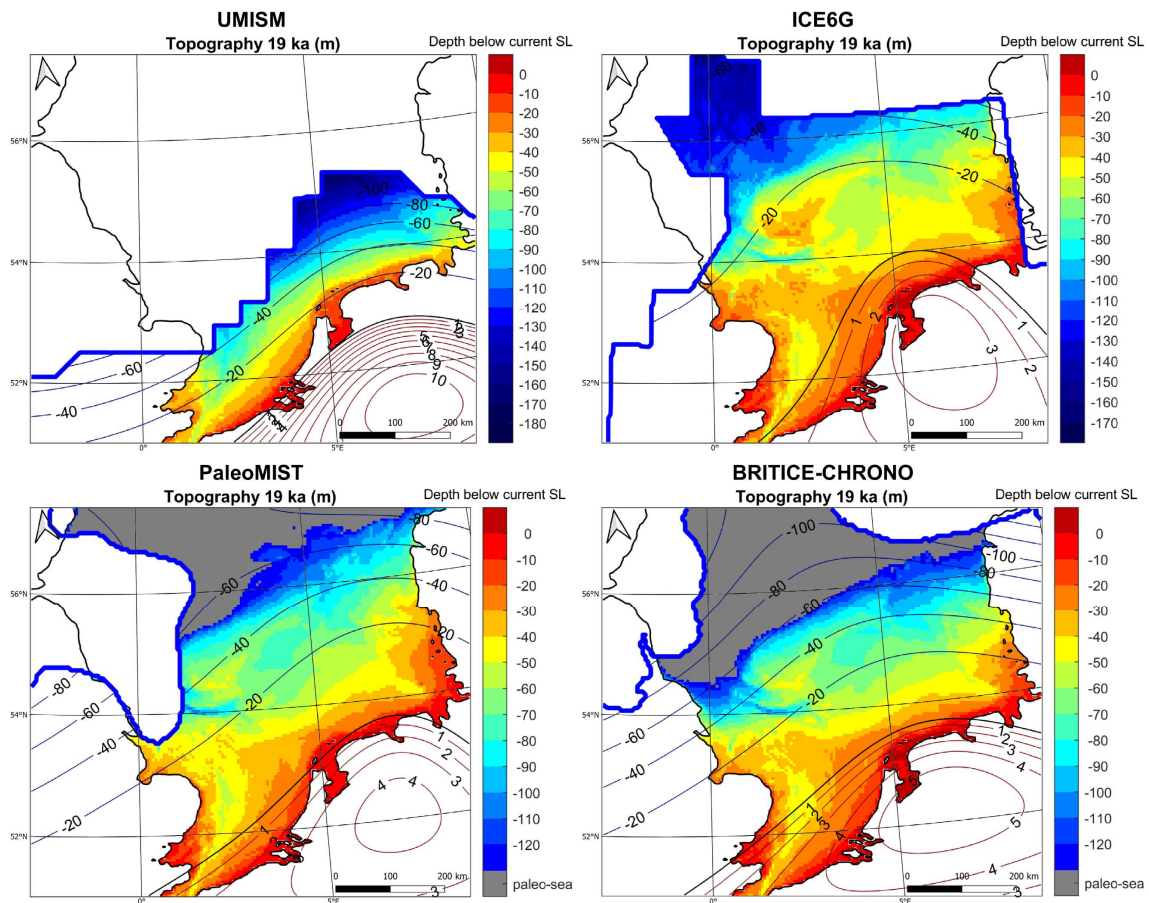


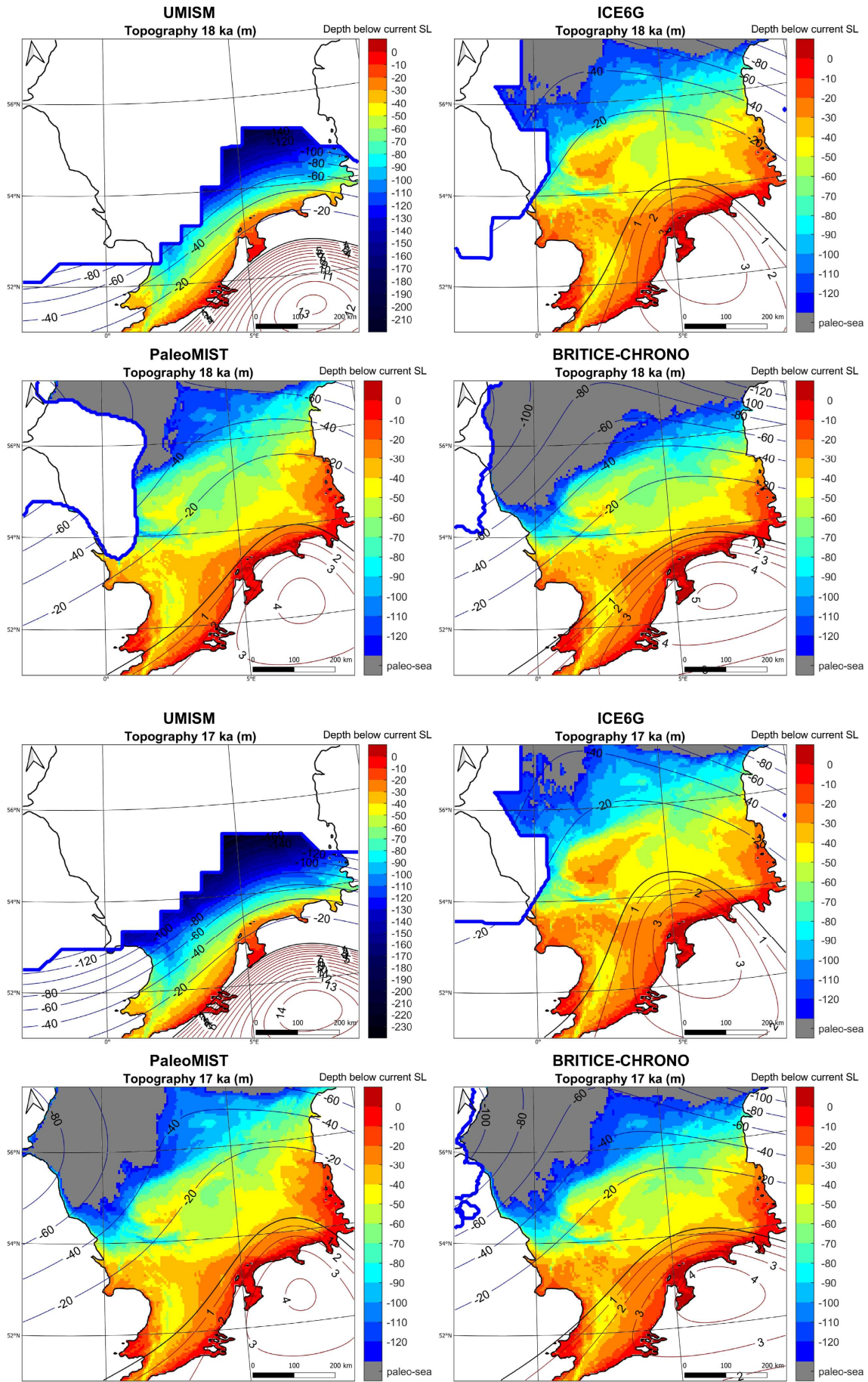
Figure 27: Evolution of the topography of the North Sea basin from 25 ka to 20 ka based on the current bathymetry and the GIA as calculated based on the different ice sheet models.

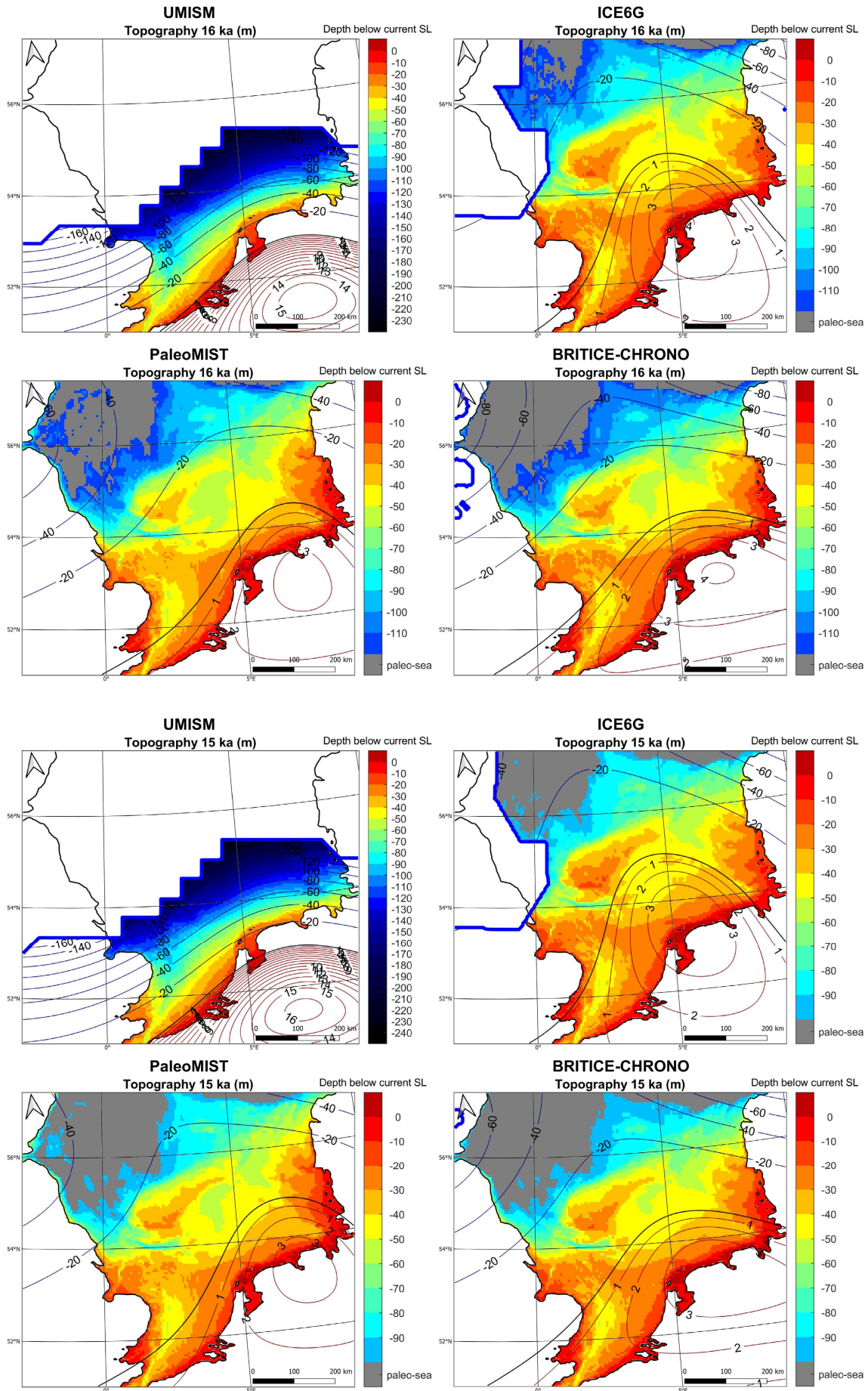
A similar situation occurred at 19 ka according to PaleoMIST 1.0 and at 18 ka according to ICE-6G_C (Fig. 28). As the deeply depressed areas in the north of the region of interest became less depressed due to the removed ice load, areas in the north became emerged again, even when sea level started to rise gradually. The competition between sea level rise and land rise continued up until 16-15 ka according to PaleoMIST 1.0, ICE-6G_C and BRITICE-CHRONO, when sea level rise started to surpass GIA-related uplift, again resulting in the gradual inundation of the southern North Sea region.

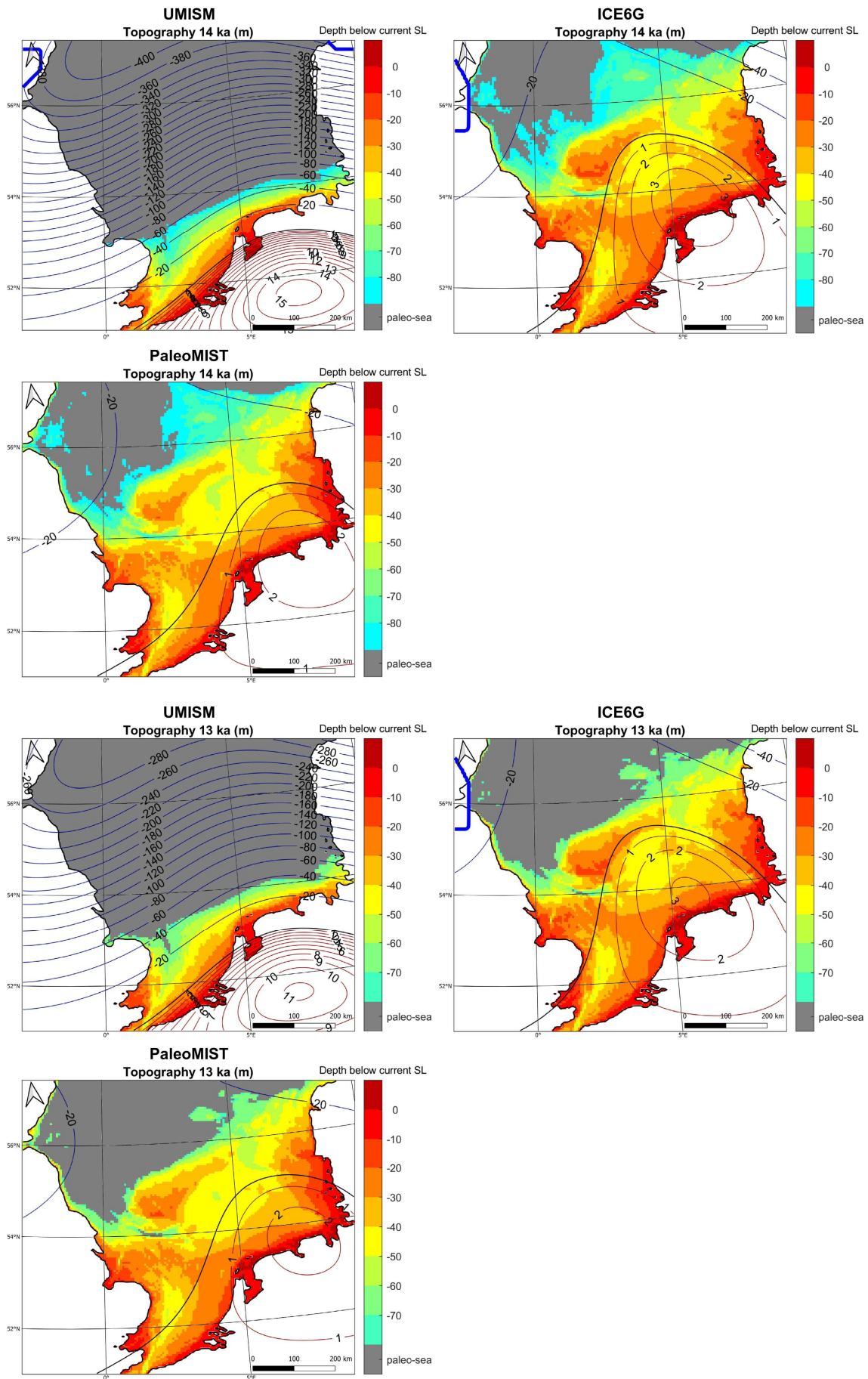
A very different signal is observed for the UMISM-based reconstructions (Fig. 28). As its ice load remained rather constant until ca. 15 ka, the lithosphere could almost completely adjust to the equilibrium-state, causing a forebulge uplift of over 16 m, and a depression of 200 m in the northernmost ice-free areas. This resulted in these areas having been located almost 250 m below the current sea level. At 14 ka, almost all ice on the British Isles and the North Sea had vanished, so that the largest part of the North Sea suddenly became flooded. Only the Southern Bight and some areas currently offshore the Netherlands and Germany were still located above sea level. Due to the subsequent GIA-related uplift, inundation of these areas only proceeded from ca. 10 ka onwards.

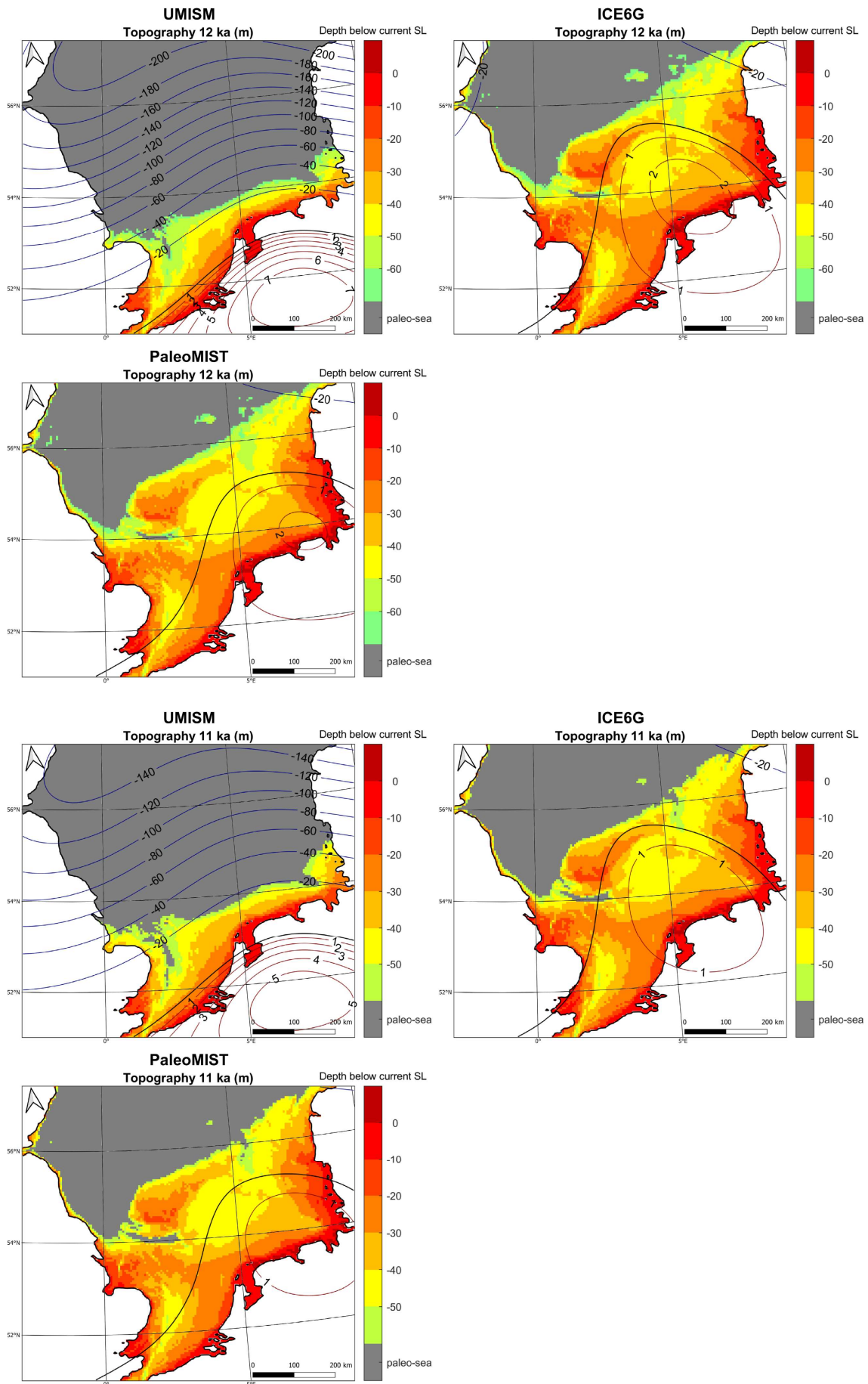
The inundation process of the PaleoMIST and ICE-6G_C based reconstructions after 15 ka occurred more gradually (Fig. 28) and could mainly be attributed to eustatic sea level rise, as the lithosphere had already largely adjusted to the removed ice load. Dogger Bank became an island around 10 ka and flooded completely around 8 ka, while the Southern Bight was largely inundated by 9 ka. At 8 ka, the coastlines were very similar to today, with mainly the Wadden Sea still being emerged, together with some areas offshore the present-day Thames and Rhine-Meuse-Scheldt estuaries.

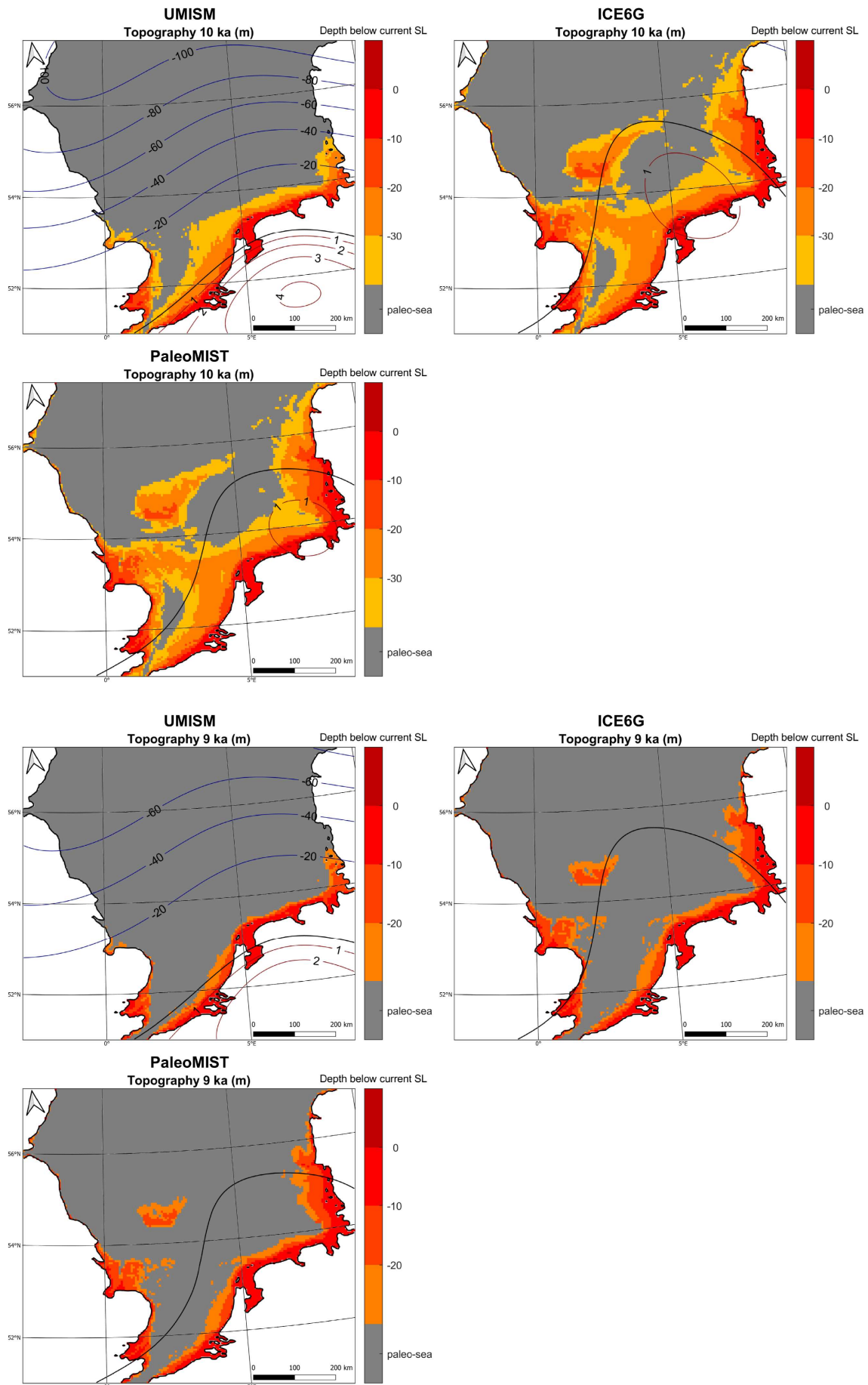












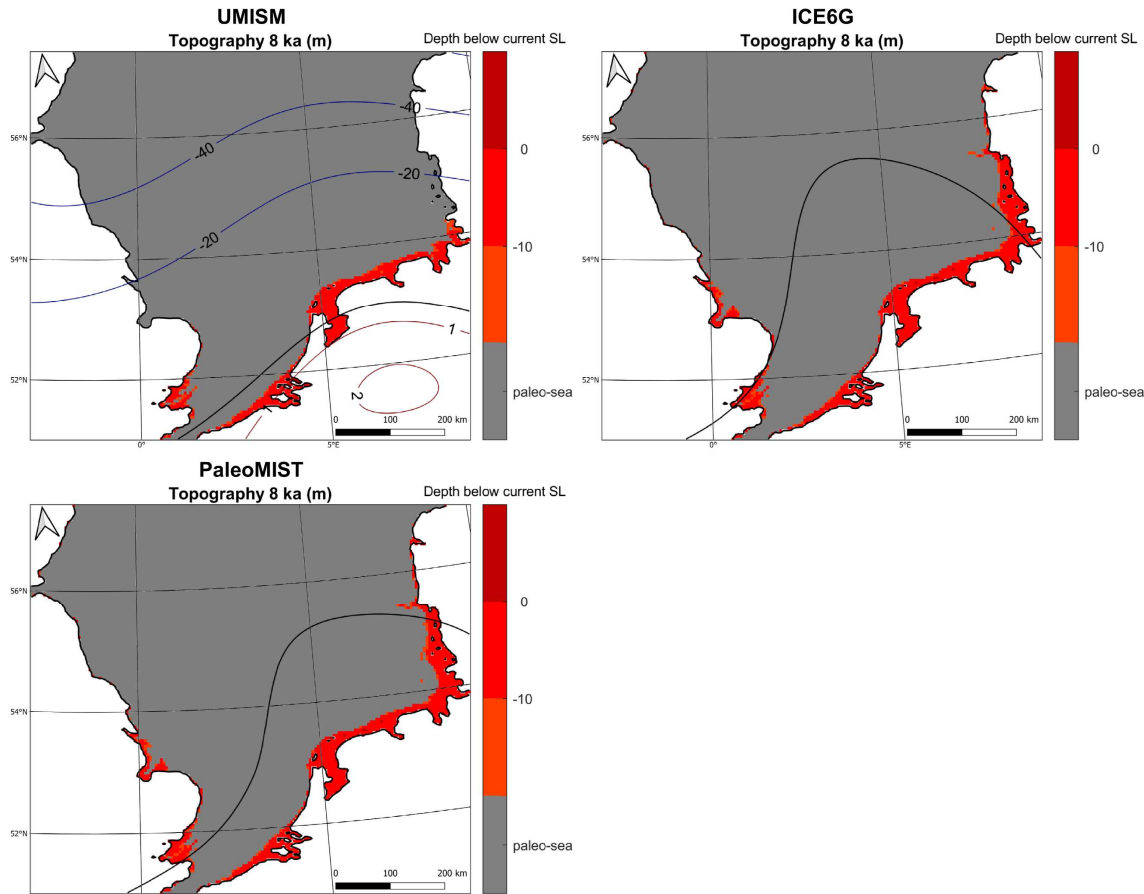


Figure 28: Evolution of the topography of the North Sea basin from 19 ka to 8 ka based on the current bathymetry and the GIA as calculated based on the UMISM, ICE-6G_C and PaleoMIST 1.0 (19 ka to 8 ka) and BRITICE-CHRONO (19 ka to 15 ka) ice sheet models.

6. DISCUSSION

6.1 Data sensitivity

The maps created based on the four different ice sheet reconstructions are all shown. Deciding which ice sheet model(s) is/are more reliable than others lies beyond the scope of this thesis, and is also not possible by only relying on the available information that is used for this study. As a result, different emersion/inundation histories and paleo-drainage patterns are compared. This means that not one solution is provided for every time slice, although the general story is similar for the different reconstructions.

For the evaluation of the separate influence of the BIIS and FIS on the GIA response in the North Sea region, the exact location of where the two ice sheets were split differs between the different reconstructions and was based on lateral ice thickness variations over the North Sea. These discrepancies in ice sheet splitting as well as true variations in ice thickness distributions might explain the disagreements in relative contributions of the two ice sheets to forebulge uplift. However, generally it can be concluded that ignoring one of the two ice sheets would result in a serious misconception of the total isostatic response to the influence of ice sheet growth and retreat over time.

Lateral variations in forebulge location due to changes in lithospheric thickness are the result of a higher flexural rigidity when the lithosphere would be thicker, which would result in a longer bending wavelength of the response to the ice weight. A thinner lithosphere also results in stronger vertical displacement compared to a thicker lithosphere. This can be explained by a thicker lithosphere causing more attenuation of the ice load signal, as the lithosphere is more rigid and viscous than the asthenosphere.

Variations in the amplitude of vertical displacement when comparing different time-constants are the direct result of the relaxation time defining the dampening effect of the response. A short relaxation time

results in a fast response to ice loading, while the response will be more smeared out over time for slower relaxation times. Geographical changes of the forebulge front between different time-constants can be explained by the ice sheet build-up prior to 20 ka. A short relaxation time will mainly depend on the ice sheet at 20 ka and shortly before this moment, while longer relaxation times will much more depend on the ice sheet configurations further back in time, when ice cover was less extensive and would have resulted in a forebulge located more northwards.

A comparison of the reconstructions based on different lithospheric thicknesses and asthenosphere viscosities illustrates differences in sensitivity of the two layers. The migration of the forebulge is only minimal for relaxation time differences, and is more substantial for variations in lithospheric thickness. Contrarily, the amount of vertical displacement varies little based on lithosphere variations, while being much more pronounced for changes in relaxation time. The migration of the forebulge is thus mainly sensitive to the lithospheric thickness, while its magnitude is mainly sensitive to the viscosity of the asthenosphere.

The vertical displacement plots very clearly confirm the intuitive thought that the depression surrounding the ice sheets becomes deeper closer to the ice margin. Depression is indeed the most pronounced below the ice sheets, although subglacial depression was not shown on the plots in order to be able to visualise the ice extent. Moreover, the study of the depression in these regions lies outside of the scope of this research. Depression further attenuates towards the depression-uplift hinge line, while the “slope” of depression decreases as well. Contrarily, the uplift is steepest at the ice-proximal side of this hinge line and decreases much more gently at the side furthest away from the ice sheet(s). Furthermore, it can be observed that when an ice sheet margin contains straight, concave and convex sections, the magnitude of uplift will be smallest along convex sections, the largest along concave sections and intermediate along straight ice margins. This can be explained as follows; displaced mantle material due to a sagging ice-covered lithosphere can diverge in different directions along a convex margin and will be converged towards one ice-free region along a concave margin. This demonstrates that the exact reconstructed configuration of an ice sheet will strongly influence the lateral distribution of vertical displacement and thus emphasises the need of very accurate ice sheet models in order to be able to make reliable GIA reconstructions.

6.2 Holocene sediments

As paleobathymetry data is quite poor, most GIA models are forced to rely on modern bathymetric maps to illustrate past topography. This creates anomalies from the true situation in the past as certain depositional or erosional features may have only formed quite recently. Isopach maps derived from borehole measurements can be used to take this effect into account for coastal regions only (Cohen et al., 2014), but this information is not always available. As mentioned earlier, the reconstructed paleotopography maps indeed did not take the presence of recent sediments into account. Hoebe et al. (2024) created a paleo-digital elevation model for the eastern part of the southern North Sea region (i.e. the complete Belgian, Dutch and German offshore sectors and onshore parts of these countries). This paleobathymetry map represents the top of the Pleistocene and was constructed based on various regional datasets. The difference between this map and the current bathymetry gives a thickness map of the present-day Holocene sediments. This Holocene thickness map is available at a lateral resolution of 5×5 km (the same resolution as was used for the paleotopography maps of this study) and at a very high vertical resolution. For the British North Sea sector, data on Holocene sediment thickness from the British Geological Survey is partly available (Balson et al., 1991; Balson, 1988; Balson and D’Olier, 1990; Cameron et al., 1984; Harrison et al., 1987; Jeffery et al., 1990, 1988). Isopach maps are largely available for the central part of the southern North Sea, but cover only to a small extent the western part. The lateral resolution of the available isopach maps is higher than for the grid from Hoebe et al. (2024), but the available isopaches are very limited (contour lines are often only drawn for 1, 5, 20 and 30 m thicknesses). Holocene sediment thicknesses north of Dogger Bank could not be considered, but are also of a lesser importance for the rest of the discussion. It has to be emphasised that taking the present Holocene sediments into consideration will still not give a completely correct image of the paleotopography during the last glacial period. Processes such as erosion, which strongly impact

emerged areas, will have altered the current bathymetry as well. Moreover, glaciogenic sediments deposited during the last glacial period were not yet present during its earliest phases.

Due to the limited availability of data, only regional variations in Holocene thickness are considered here. North of the Axial Channel, a unit of Holocene sediments more than 5 m in thickness and locally increasing up to a thickness of 15 m is present. More to the north, Holocene sediments at the eastern parts of Dogger Bank span 5 to 20 m in thickness. Sedimentary deposits reaching 5 to 20 m are also present in the German offshore North Sea sector, and are distributed along a NNW-SSE oriented depocentre. This is the representation of the Elbe Paleovalley, along which meltwater from the Fennoscandian Ice Sheet and the ice-free landmasses of northern Europe was transported seawards in the direction of the Norwegian Channel during MIS 2 (e.g. Özmaral et al., 2022). The valley drained, amongst others, rivers such as the Eider, Elbe, Weser and Ems. This northward drainage via the Elbe Paleovalley can only have occurred in absence of the coalescence of the BIIS and FIS over the North Sea and in absence of a proglacial lake in the same area (Özmaral et al., 2022). The activation of this valley likely was accompanied by valley incision due to a much lower sea level. Its inundation during the postglacial sea level rise, which is described by Hoebe et al. (2024), on the other hand caused the deposition of Holocene sediments in this area. Furthermore, the Wadden Sea is covered by 5 to over 20 m of Holocene sediments, as well as estuarine areas of rivers such as the Weser, Ems, Rhine, Meuse and Scheldt. In contrast to what was expected, the Holocene sediment cover at the Frisian Front is negligible. The creation of sandbanks such as those offshore Belgium also caused a local increase in Holocene sediment thickness. Taking the Holocene sediments north of the Axial Channel, at the eastern side of Dogger Bank and along the Elbe Paleovalley into account is important to make correct interpretations on past drainage patterns based on (paleo)topographic constraints. These areas are indicated with dotted lines on Fig. A7 (see appendix).

6.3 Drainage patterns during the last glacial period

Evidently, during periods when the BIIS and FIS were merged, the drainage of rivers usually entering the current North Sea basin will have been completely different compared to periods that were characterised by an ice-free corridor towards the northern North Sea.

The existence of (a) proglacial lake(s) located at the southern edge of the merging ice sheets has already been hypothesised by several authors (e.g. Sejrup et al., 2016), and possible evidence of the drainage of such a lake after the LGM was provided by Hjelstuen et al. (2018) (see Fig. 6 for the assumed dimensions of this lake).

Hjelstuen et al. (2018) also referred to Bigg et al. (2012) for the volume estimation of this lake, which was calculated to have been approximately 3900 km³. One should consider two important issues when relying on this volume calculation. First of all, its extent is based on the reconstruction of maximum ice sheet margins and bathymetric variations, thus not taking GIA-effects into account. Secondly, the area of the lake reconstructed by Bigg et al. (2012), extending from the ice margin well into the Southern Bight and covering almost the entire now inundated areas in between, is much larger than the lake reconstructed by Hjelstuen et al. (2018). Indeed, when calculating the lake volume based on the current bathymetry and using a similar lake extent as was used by Bigg et al. (2012), but relying on the BRITICE-CHRONO 23 ka southern ice margin, a result very similar to 3900 km³ can be obtained.

Based on amongst others isotopic data from a core from the northern North Sea margin and a comparison with other cores taken in the region, Lekens et al. (2006) found proof for meltwater release in the southern Norwegian Sea not related to Heinrich events around 34 ka. It is this timeframe that Bigg et al. (2012) suggested as the possible timing for the northward drainage of their suggested proglacial North Sea lake, and does thus not coincide with the drainage suggested by Hjelstuen et al. (2018), i.e. resulting from ice sheet unzipping after the LGM. The ICE-6G_C indeed indicates continuous ice cover over the North Sea from MIS 4 onwards, but only shows ice sheet unzipping around 18 ka. The more recent and spatially much higher resolved PaleoMIST 1.0 and BRITICE-CHRONO reconstructions only indicate ice sheet merging after 30 ka. The only used reconstruction showing ice sheet zipping and

unzipping roughly around 34 ka is UMISM, according to which a very brief period of ice sheet merging occurred around 36 ka. The zipping zone of the two ice sheets did not exceed ice thicknesses of 300 m, and more or less the complete BIIS had disappeared by 35 ka. It is thus uncertain whether this very short-lived North Sea ice sheet barrier could have resulted in a proglacial lake with the dimensions suggested by Bigg et al. (2012).

Research focussing on the now inundated Elbe Paleovalley - also referred to as Elbe Urstromtal or Heligoland Channel - is quite limited. The area was first mapped by Figge (1980), and today it is generally assumed that the formation and enlargement of this valley only occurred during MIS 2, as a result of the nearby location of the FIS delivering its outwash and meltwater, but also because the valley was draining parts of eastern, western and central Europe (Özmaral et al., 2022; Papenmeier and Hass, 2020; Westaway and Bridgland, 2010). Even though there is no proof of its existence during earlier glaciations, it can be assumed that during the Saalian and subsequent periods of sea level lowstand the Elbe and Weser would have drained more or less in a similar direction.

At the end of the Last Interglacial, when sea level was slightly more elevated than today, the Elbe and Weser can be expected to have entered the North Sea more or less at the same location as their present-day mouths. However, the courses of the more southerly located rivers were different from their current pathways. The Rhine entered the North Sea much more northwards, east of the present-day Lake IJssel, as the area west of it was completely inundated, while the Meuse estuary was located more southwestwards compared to the Rhine estuary, but still north of its present-day mouth (Busschers et al., 2007). The Scheldt river entered the North Sea via re-occupation of its paleovalley deflecting westwards north of Ghent towards Zeebrugge and Ostend (De Clercq et al., 2018). This incised valley was shaped at the end of the penultimate glaciation, presumably due to GIA-related forebulge uplift and subsequent depression, and had evolved further over an estuary into a seaway during the Eemian. Due to the subsiding forebulge formed by Saalian ice sheets, the Eemian transgression lagged behind eustatic sea level rise (De Clercq et al., 2018). This phenomenon is often observed in the southern North Sea during interglacials, however, these transgressions will have proceeded quite fast (Cohen et al., 2017).

Even though the North Sea was not yet covered by ice during MIS 5, large fluctuations in temperature and sea level at the start of the last glacial period have broadly impacted the inundation-emersion history of the region. During the MIS 5c and MIS 5a interstadial peaks (in Northern Europe referred to as Brørup and Odderade), the eustatic sea level was only located 20 to 25 m below the current sea level (Lambeck et al., 2002). An example of this situation can be observed by looking at the topography plots from 85 ka, a time slice which is very close to the MIS 5a peak (Fig. 22). During these interstadials, the Axial Channel and large parts of the Southern Bight were located below sea level. When taking into account that the barrier between the Southern Bight and the North Sea more northwards is covered by a non-negligible Holocene sedimentary package, it can be inferred that the Southern Bight region was indeed connected to the inundated part of the North Sea. During these relatively warm, interstadial periods, the rivers will have drained quite similar to their late Eemian pathways, but will have incised and prograded towards the lower-located sea level and started building up delta systems such as the Rhine-Meuse delta (e.g. Busschers et al., 2007; De Clercq et al., 2018; Peeters et al., 2015).

Contrarily, sea level was located 55 to 60 m lower than today during the MIS 5d and MIS 5b stadial peaks, in Northern Europe referred to as Herning and Rederstall (Lambeck et al., 2002). An example situation can be seen on the 95 ka plot (Fig. 22), representing the MIS 5b sea level lowstand. The entire Southern Bight region was located above sea level, and the coastline ran in a SW-NE direction north of Dogger Bank. The Outer Silver Pit was the only isolated basin located below sea level and could potentially have been a lake depending on the exact drainage pattern. The Rhine, Meuse and Scheldt rivers were probably directed towards the low-lying Axial Channel, from where drainage towards the Dover Strait and the Atlantic Ocean was possible. Even if the Elbe Paleovalley did not yet exist, drainage of the Elbe and Weser will most likely have been directed towards the north along a similar pathway. An alternative hypothesis could involve a northward drainage until encountering the Oyster Ground, a

depression having been a potential target for lake formation. Water from the Oyster Ground could then have been routed towards the western part of the North Sea via the Outer Silver Pit. Whether a lake could have formed at the Oyster Ground depends on the amount of water discharge at that time and on whether or not barriers preventing water outflow towards the west and north were present. If such an outflow was present, the Oyster Ground might not have been filled by lake water, but could have been a plain traversed by the lowermost part of the Elbe-Weser drainage system. During these stadials it was probably moistier than during the peak glacial moments at MIS 4 and MIS 2, but on the other hand, meltwater discharge from the limited FIS would not yet have been directed so far south. As far as known no evidence of this alternative drainage route has been found.

At the MIS 5 - MIS 4 transition around 75 ka, the study area started to experience its first glacial influence, according to some of the ice sheet reconstructions. This likely will have caused an increase in the flow rate of rivers draining meltwater from these ice sheets, while not yet affecting the course of the rivers, as GIA effects are still rather limited. The eventual coalescence of the BISS and FIS over the North Sea will have had the largest influence on river drainage. All used ice sheet models place this coalescence sometime during MIS 4, although there is still disagreement on the exact timing. UMISM indicates a short period of coalescence around 70 ka, after which the North Sea ice bridge disappeared for over 5 kyr. Slightly later, ICE-6G_C indicates a merged BISS-FIS, followed by the re-appearance of the UMISM ice bridge. Lastly, PaleoMIST 1.0 only indicates coalescence of the two ice sheets around 60 ka.

In any case, regardless of its exact timing, the coalescence of the ice sheets over the North Sea region will have prevented a northward drainage of rivers like Elbe and Weser, while the supply of meltwater from the FIS will already have increased drainage rates. The ice blockage might have caused the formation of a proglacial lake with the ice margin as its northern boundary (Fig. 29). GIA will have further deepened the potential lake basin, with the forebulge being situated southeast of the potential lake. The confluence of the Elbe and Weser, as well as other FIS meltwater drainage routes will have been the major tributaries of this proglacial lake. Once the complete lake basin was filled with water, overtopping will have resulted in drainage of the excessive water towards the south. This is expected to have occurred via the topographically depressed Axial Channel. Together with water supplied by rivers such as Rhine, Meuse and Scheldt, the water will have been directed towards the Dover Strait and further to the Atlantic Ocean. During MIS 4, the Rhine and Meuse were characterised by lateral reworking and migration and by the re-occupation and broadening of ancient braided channels (Busschers et al., 2007). Especially the course of the Rhine became very widened at the IJssel basin, before its deflection towards the south, after which it connected with the Meuse before entering the Axial Channel. This course of the Rhine is, however, quite inconsistent with the paleotopography as a more (north)westward direction would be expected based on the topographic gradient, so no definite conclusions on the Rhine pathway can be drawn.

The existence of a proglacial lake during MIS 4 was hypothesised as well by De Clercq (2018), stating that the overtopping of the lake may have resulted in a GLOF or a cascade of GLOFs towards the Axial Channel. More proof of such catastrophic events is, however, needed to confirm this hypothesis, as the intensity of lake drainage towards the south will have largely depended on the speed of water supply towards the lake, the stability of the topographic saddle where the overtopping occurred, as well as on the exact paleotopography of the Axial Channel during MIS 4. The maximum volume of the suggested proglacial lake is calculated based on the paleotopography as shown in Fig. 29, but without backstripping of the Holocene and uppermost Pleistocene sediments and thus resulting in a slightly deeper basin and thus in a slightly higher lake volume. The volume is estimated to 1130 km³, which is 3.5 times smaller than the MIS 2 lake volume calculated by Bigg et al. (2012). The PaleoMIST 1.0 based reconstruction was used, as PaleoMIST 1.0 is the most recent available reconstruction, and was provided at the highest spatial resolution.

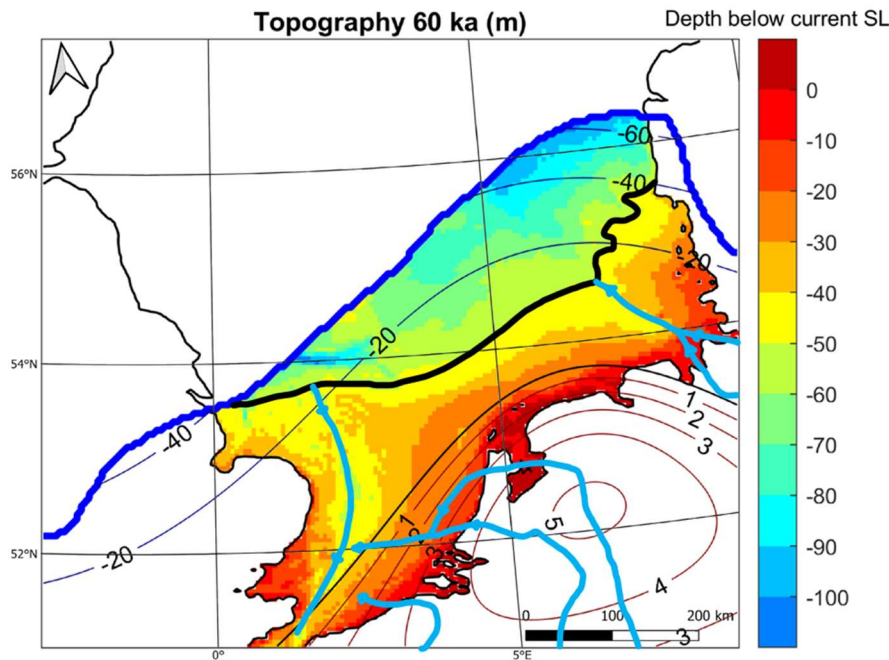


Figure 29: Assumed hydrological situation of the southern North Sea basin at 60 ka. Southern proglacial lake limit (bold black line) based on paleotopography and on the known existence of Holocene sediments, main river pathways depicted in blue. Courses of Scheldt river and Meuse and Rhine rivers based on De Clercq et al. (2018) and Busschers et al. (2007) respectively, courses of Weser and Elbe rivers as well as southward drainage of the assumed proglacial lake based on paleotopography.

After the MIS 4 peak, retreat of the ice sheets might have resulted in a temporary enlargement of the lake basin, counteracted by the subsequent attenuation of GIA-related depression. The lake will only partly have drained towards the south (i.e. by overtopping), as a complete southward drainage would have required an inversion of the topography. Once the BIIS and FIS likely became unzipped (although this is not in agreement with the ICE-6G_C reconstruction), the remaining lake will have emptied towards the north into the North Sea.

At the beginning of MIS 3, the Rhine would have partly avulsed and merged with the Meuse in the central Netherlands (Busschers et al., 2007). The newer course was shorter and had a higher gradient. The avulsion is explained to be the result of aggrading channel belts and an increase in sediment supply (Busschers et al., 2007). However, at this moment the forebulge had already partly moved in a northward direction as a result of retreating ice sheets. The Rhine was thus completely located on its southern side. Even though the magnitude of forebulge uplift will only have been a couple of meters, the downward slope might have been an extra trigger to favour a more southerly located course, as was also suggested by Busschers et al. (2007). The pathways of the Scheldt, Meuse, Weser and Elbe are expected to have been similar as during MIS 4, however, with the Weser and Elbe not draining into a proglacial lake but again into the North Sea. The northern course of the Rhine was abandoned at around 35 ka (Busschers et al., 2007).

At the MIS 3 - MIS 2 transition at ca. 30 ka, the North Sea region was likely ice-free. However, both the BIIS and FIS started to expand again, so meltwater originating from these ice sheets once again started to become more directed towards the study area. It is around this timeframe that the Elbe Paleovalley likely started to develop. Rivers such as the Weser and Elbe together with FIS meltwater drained towards the still inundated Norwegian Channel via this newly formed valley, which is shown to have been occupied by a braided river system (Özmaral et al., 2022). Simultaneously, the Scheldt, Meuse and Rhine rivers still drained southwards via the Axial Channel, with the latter by then only having occupied its southernmost pathway.

At the onset of MIS 2, the BIIS and FIS merged, once again preventing a northward drainage of FIS meltwater and northwestern German rivers towards the Norwegian Channel. A deflection of this

northward drainage towards the western part of the southern North Sea and further to the south via the Axial Channel has already been hypothesised by amongst others Ménot et al. (2006) and Toucanne et al. (2010). While Toucanne et al. (2010) suggested this southwest drainage direction to have existed from 30 to 18 ka based on the assumed period of ice sheet merging, Ménot et al. (2006) suggested this drainage direction to have changed again only at 17 ka. Both publications link this change in drainage direction to the existence of one or several proglacial lakes along these pathways south of the ice margin. Based on paleogeographical reconstructions for southwest Scandinavia by Houmark-Nielsen and Kjær (2003) depicting the inundation of the Norwegian Channel since 24 ka, Hijma et al. (2012) suggested that the Elbe became rerouted towards the north since then. This northward drainage already at 24 ka cannot be deduced from the paleotopography maps from this study. These maps suggest a northward rerouting to have not occurred before 20 ka.

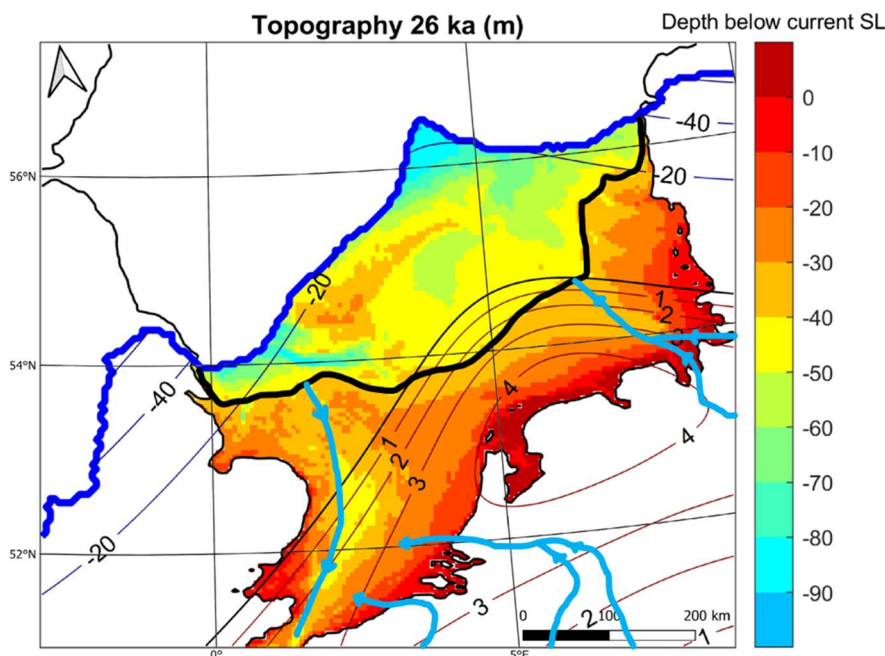
Based on amongst others neodymium isotopes, Toucanne et al. (2015) found evidence for major meltwater injections towards the English Channel since the coalescence of the BILS and FIS until their unzipping, as well as the deposition of sediments into the Bay of Biscay in the Atlantic Ocean. This is explained as being the result of short-lived episodes of substantial ice melting throughout the period of ice sheet merging. Moreover, García-Moreno (2017) found a series of erosional features at the Dover Strait, which were formed during its last known phase of valley incision, during the LGM. These features were most likely created between 30 and 18 ka and are evidence of the occurrence of a megaflood. This evidence, together with the suggestions from Ménot et al. (2006) and Toucanne et al. (2010) and the paleotopography maps from this study strongly indicate that the southwestward drainage originating from the southeastern part of the North Sea basin was accompanied by the formation of one or several proglacial lakes that drained via the Axial Channel towards the Dover Strait and further into the Atlantic Ocean via the Channel River. The timing of multiple GLOF drainage phases into the Atlantic Ocean as observed by Toucanne et al. (2015) will have depended on the timing of breaching or overtopping of this lake into the Axial Channel, which in turn was the result of the interplay between ice sheet melting phases and the GIA-depending storage capacity of the lake.

The maximum proglacial lake storage capacity is quantified as was done for the 60 ka time slice. However, the BRITICE-CHRONO based paleotopographic reconstructions were used for the LGM, as this is the most recent and best constrained available LGM reconstruction. The maximum lake extents and the main drainage pathways are shown for 26, 23 and 21 ka (Fig. 30). The first-mentioned time slice represents the first moment of ice sheet merging according to BRITICE-CHRONO, and gives thus also the first possibility for a proglacial lake to have formed. After 21 ka, the ice sheets unzipped again and the existence of a proglacial lake was no longer possible. 23 ka represents an intermediate time slice. Rivers such as the Weser and Elbe, together with other meltwater-transporting pathways are assumed to have been the main inputs into the proglacial lake, while catastrophic or more steady overtopping occurred north of the Axial Channel, draining proglacial lake water as well as water from rivers such as Rhine, Meuse and Scheldt towards the Atlantic Ocean. At 26 ka, the maximum lake volume is estimated to have been around 965 km³, with the maximum lake level roughly being situated 40 m below the current sea-level. An increase in lake level would have resulted in overtopping towards the Axial Channel and drainage of the excess lake water. On Fig. 30A, Dogger Bank appears as an island within the lake, however, the fact that at that location a noteworthy Holocene sedimentary unit is deposited has to be taken into account. Whether the available basin became indeed completely filled with water at 26 ka depends on the amount of water supply. Filling of the lake would have started in the north, as this was the former drainage direction of the Elbe and Weser rivers and was also the deepest part of the basin. The further filling of the lake occurred according to the paleotopography by first inundating the deepest depressions, and outflow towards the south was only possible once the entire basin was filled and overtopping occurred. During the gradual filling, possibly several semi-separated basins could have existed, although at least some sort of connection between them will have been necessary.

By 23 ka, the ice margins had moved far enough south to cover Dogger Bank. GIA-related depression had increased in the area, and the maximum elevation of the lake level was ca. 50 m below the current sea-level, as the point of breaching had become more depressed as well. The maximum lake volume at

23 ka is calculated to be roughly 765 km³. Although the area was more depressed than at 26 ka, the maximum lake volume had decreased due to the southward movement of the ice margin.

By 21 ka, shortly before unzipping of the BIIS and FIS, the ice margin had already retreated substantially towards the north. This enlarged the basin remarkably, as this northern part of the basin was also strongly depressed due to the former ice sheet. The basin could at the time be filled to roughly 60 m below present-day sea level, as the barrier north of the Axial Channel had become further depressed. The lake volume at 21 ka is estimated to 3040 km³, a value quite close to the volume estimation from Bigg et al. (2012), although the calculation from this study is not based on the current bathymetry of almost the complete southern North Sea, but on the deeply depressed paleotopography of its northern part. This volume can be compared to that of the Glacial Lake Missoula, which drained into the Pacific Ocean by several GLOFs between 15 ka and 13 ka, discharging an estimated volume of 2184 km³ (Clarke et al., 1984; O'Connor and Baker, 1992). It is likely that the available basin at 21 ka indeed became completely filled with water. The ice sheet was rapidly retreating, releasing large volumes of meltwater directly into the lake from the southern north sea ice margin as well as from other melting regions of the ice sheet directed towards the lake via proglacial rivers. The existence of the lake itself may have enhanced the retreat of the ice margin as well. If there would have been no lake, the ice would have been grounded onto the emerged North Sea basin. However, the existence of a large body of water in front of the ice sheet may have induced flotation of the ice, and at least will have caused calving of ice shelves. At the timing when the ice sheets started to unzip, the excess of meltwater that did not anymore fit the lake basin had drained towards the south and had left behind a glacial outwash plain. When the two ice sheets detached, a deeply depressed opening towards the north will have been the favorable pathway to discharge the remainder of the lake water as a GLOF. The timing of this event is still unresolved. The BRITICE-CHRONO reconstruction indicates an ice-free corridor towards the north at 20 ka, but Hjelstuen et al. (2018) suggest this GLOF to have occurred at ca. 18.7 ka, amongst others based on ice extent reconstructions from Sejrup et al. (2016). Emptying of the proglacial lake and retreat of the BIIS and FIS enabled the re-occupation of the Elbe Paleovalley and its connection with the Norwegian Channel, until its drowning during the Early Holocene (e.g. Hoebe et al., 2024; Özmaral et al., 2022).



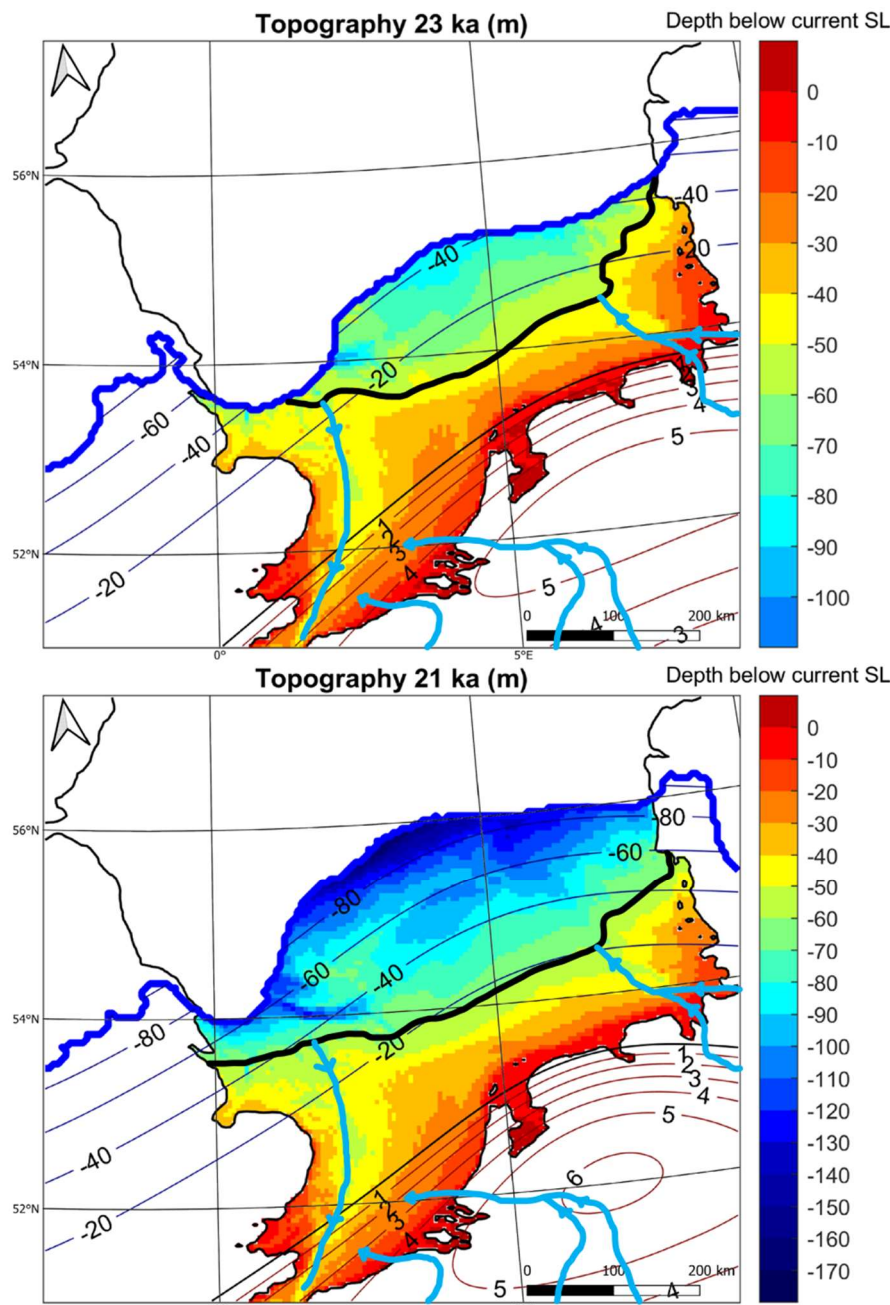


Figure 30: Assumed hydrological situation of the southern North Sea basin **A)** at 26 ka, **B)** at 23 ka and **C)** at 21 ka. Southern proglacial lake limit (bold black line) based on paleotopography and on the known existence of Holocene sediments, main river pathways depicted in blue. Courses of Scheldt river and Meuse and Rhine rivers based on De Clercq et al. (2018) and Busschers et al. (2007) respectively, courses of Weser and Elbe rivers based on the position of the Elbe Paleovalley (Özmaral et al., 2022), southward drainage of the assumed proglacial lake based on paleotopography.

Busschers et al. (2007) already suggested that the course of the Rhine at the onset of deglaciation was located quite close to the maximal zone of uplift. This can also be seen on Fig. 30C, which indicates that the most uplifted part of the Rhine and Meuse courses was the zone right before their deflection towards the west. Changes in inclination can have altered the amount of incision and aggradation of the river systems, irrespective of changes in sedimentological patterns (Busschers et al., 2007). Lateral migration of the rivers due to this forebulge-related uplift cannot be excluded as well.

At the end of the last glacial period, the Scheldt river changed its course from a westward direction north of Ghent to a continuation to Antwerp and from there further northwestward towards the North Sea. This has often been explained as the result of the formation of the east-west oriented Maldegem-Stekene

coversand ridge (e.g. De Moor and Heyse, 1978), which started roughly between 17.5 and 15.5 ka (Derese et al., 2010) and would have blocked the former drainage route via the Scheldt Paleovalley. De Clercq et al. (2018) suggested the possibility that the LGM related GIA might have impacted this late-glacial Scheldt course change, implying that the formation of the coversand ridge was not its only trigger. According to the GIA output models from this study, the upwards forebulge slope in Flanders during the timing of the pathway change runs from west to east or from south to north, depending on the model. Given the fact that the Scheldt river started diverting towards the east and northeast, the river thus deflected along the forebulge isolines or even towards more uplifted areas. A conclusion attributing deflection of the Scheldt river towards Antwerp as a result of GIA-related uplift can thus not be drawn. Then again, it cannot be excluded that the exact configuration of the forebulge might have favoured the formation of the Maldegem-Stekene sand ridge along an east-west orientation.

6.4 Further research

In this study, a eustatic (i.e. global) sea level curve was used to generate paleotopography maps, so that an estimation of the paleo-shoreline could be made. However, as the local sea level curve differs from the eustatic sea level, hydro-isostatic loading is not taken into account. When a high resolution local sea level curve for the North Sea basin is available that is well calibrated based on local observations of the sea level at a specific location at a specific moment (so-called sea level index points), the water load can be taken into account as well to calculate a more accurate isostatic response. This would then facilitate interpretations during sea level highstands. However, it should be taken into account that the local sea level curve will differ between different locations within the North Sea basin as well.

Hoebe et al. (2024) reconstructed the inundation of the southern North Sea basin during the Early Holocene. Amongst others, they relied on local sea level curves produced for the Holocene based on GIA modelling. Basin background subsidence was taken into account as well. When more data is available for the complete last glacial period, it would be interesting to also rely on a local rather than on a eustatic sea level curve and to take basin background subsidence into account. A more complete and higher resolution record of the thickness distribution of Holocene sediments as well as older glacial sediments in the study area is needed to improve reconstructions as well.

Several ice sheet reconstructions are available for (a part of) the last glacial cycle. There is, however, need for well-constrained and high resolution publicly available ice sheet reconstructions from previous glaciations that could be used for GIA calculations further back in time. These can, for example, be used to verify how isostatic effects may have affected past river course changes, such as the formation of the Scheldt Paleovalley at the end of the penultimate glaciation (De Clercq et al., 2018). There is also the need to better constrain the formation and extent of proglacial lakes during MIS 6 and MIS 12 and how their emptying resulted in the breaching and two-stage opening of the Dover Strait (e.g. Gupta et al., 2017). Nevertheless, reconstructions from previous glaciations and even from the older stages of the last glacial period become less constrained as they have to deal with several uncertainties. First of all, data on ice margins becomes more scarce when going further back in time, as the signal has partially been removed by more recent ice advances. Secondly, it is not completely resolved how exactly the ice extent can be translated into thickness variations as these are only modelled and not observed on the field. Ice thickness models can only be calibrated using now existing ice sheets such as the ones on Antarctica and Greenland, but whether the ice sheet dynamics are the same for all ice sheets is not guaranteed. This and future studies confining the hypothesised location of former proglacial lakes can aid the search for lacustrine sediments in the North Sea basin, which in turn can be used to better constrain paleotopographical reconstructions.

The choice of Earth parameters is crucial for GIA studies. The range of asthenosphere viscosity values that are used in GIA studies is quite small, and values within this acceptable range give quite similar results. However, the exact lithosphere thickness is not yet agreed upon, and probably varies significantly between different regions depending on their geological past. Choosing different values within the range of values that is applied in different studies can largely influence the results, so a better identification of its thickness distribution is crucial.

7. CONCLUSIONS

In this study, the GIA related forebulge and depression in the North Sea region resulting from the weight of the British-Irish and Fennoscandian Ice Sheets was modelled based on four different ice sheet reconstructions. Although three of them also contain information on the thickness distribution of the Barents-Kara Ice Sheet, it was shown that for the region of interest taking the BIIS and the southern part of the FIS into account produces relevant and reliable results. However, in contrast to what was done in some previous GIA studies, both of these ice sheets really need to be considered.

This study has also shown that the forebulge migrates in space over time. Ice sheet advances push the region of uplift further away from the region of ice sheet nucleation, while a retreat of the ice induces a backward movement of the forebulge. Once all ice in the region of interest is removed, the remaining GIA signal will not migrate any further and simply decreases in amplitude over time towards zero vertical displacement. During the LGM, the zone of maximal forebulge uplift migrated from the northern part of the Netherlands further to the more central and southern parts of the country, while having been located much more to the northwest during less glaciated moments.

This study also highlighted the strong need for more empirical constraints of ice limits during the older parts of the last glacial period and during previous glaciations to be able to generate (better) ice sheet models that in turn can be used for GIA calculations. For certain periods characterised by glacial maxima there is already some data available, but at least during relatively less glaciated periods ice limit reconstructions are very poorly confined. Moreover, as the exact ice sheet geometry strongly influences the lateral displacement of mantle material, this further emphasises the importance of accurate ice sheet reconstructions.

The generation of better ice sheet models goes hand in hand with the creation of better Earth models. There is still some discussion on which Earth model and which parameter values should be used. In this study, it is shown that these Earth parameters are crucial for accurate GIA reconstructions. The lithosphere thickness and asthenosphere viscosity have an important impact on the migration and magnitude of the GIA response, respectively.

The GIA response can strongly impact the inundation-emersion history of proglacial areas, as this study has shown for the North Sea basin. Post-LGM glacio-isostasy is still an ongoing process today and is thus crucial in our understanding and quantification of anthropogenic sea level rise. Even though there are significant differences between ice sheet reconstructions and therefore also on the GIA response and reconstructed paleotopography during the last glacial period, some general conclusions could be drawn:

- Glacial isostatic adjustment had an important impact on the hydrology and drainage patterns of proglacial regions in the central and southern North Sea;
- Forebulge uplift has induced river course changes, such as the avulsion of the Rhine towards the south during MIS 3;
- GIA related depression was also able to enlarge the potential of basins for proglacial lake formation close to the ice margin, such as the assumed southern North Sea lake.

This study provided extra evidence for such a lake having been wedged in between the ice sheet margin in the north and the forebulge in the southeast, with a potential outlet drainage via the Axial Channel towards the Dover Strait and the Atlantic Ocean in the southwest both during MIS 4 and MIS 2. The results of this study are thus in agreement with already existing evidence of i) breaching events in the Dover Strait during MIS 2 due to lake overflow, and of ii) GLOF-generated deposits in the northern North Sea as a result of eventual lake emptying as a GLOF following ice sheet unzipping.

There is the need for more GIA calculations providing extra corroboration for similar mechanisms having occurred during MIS 6 and MIS 12 related to the opening of the Dover Strait, as well as demonstrating older river course changes having been related to glacio-isostasy. It is, however, also important to better understand how young Quaternary deposits have altered the paleotopography and how local sea level changes have affected the North Sea basin due to hydro-isostasy.

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9. APPENDIX

Table A1: Paleo-sea level elevation compared to present-day sea level inferred from Lambeck et al. (2002) and Lambeck et al. (2014) at time slices for which reconstructions are made.

Time (ka)	Meter relative to present-day sea level
8	-12
9	-25
10	-39
11	-55
12	-63
13	-71
14	-87
15	-99
16	-117
17	-123
18	-122
19	-126
20	-132
21	-134
22	-133
23	-132
24	-131
25	-130
26	-129
27	-128
28	-127
29	-126
30	-111
35	-68
40	-75
45	-57
50	-48
55	-62
60	-52
65	-73
70	-57
75	-60
80	-49
85	-28
90	-31
95	-53
100	-38
105	-37
110	-42
115	-39
120	+3

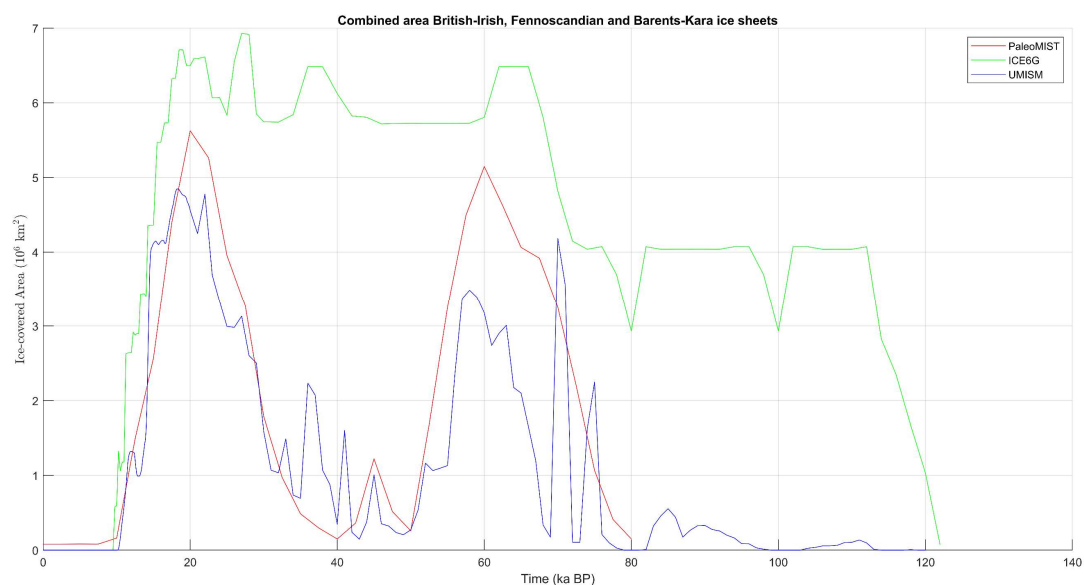


Figure A1: Evolution of areal ice extent over time for the three Eurasian ice sheet models used in this study.

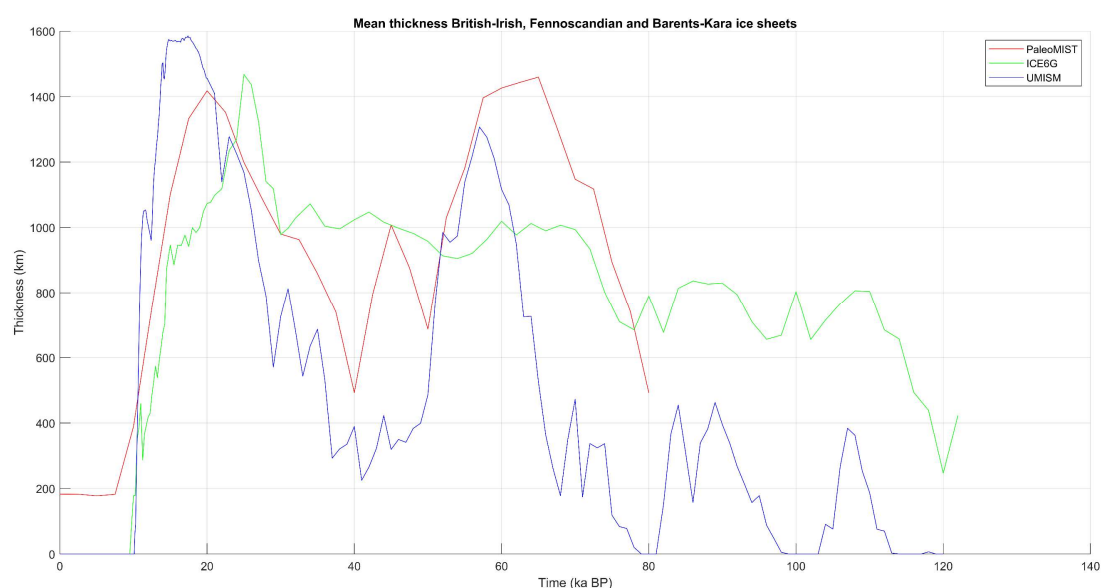


Figure A2: Evolution of average ice thickness over time for the three Eurasian ice sheet models used in this study.

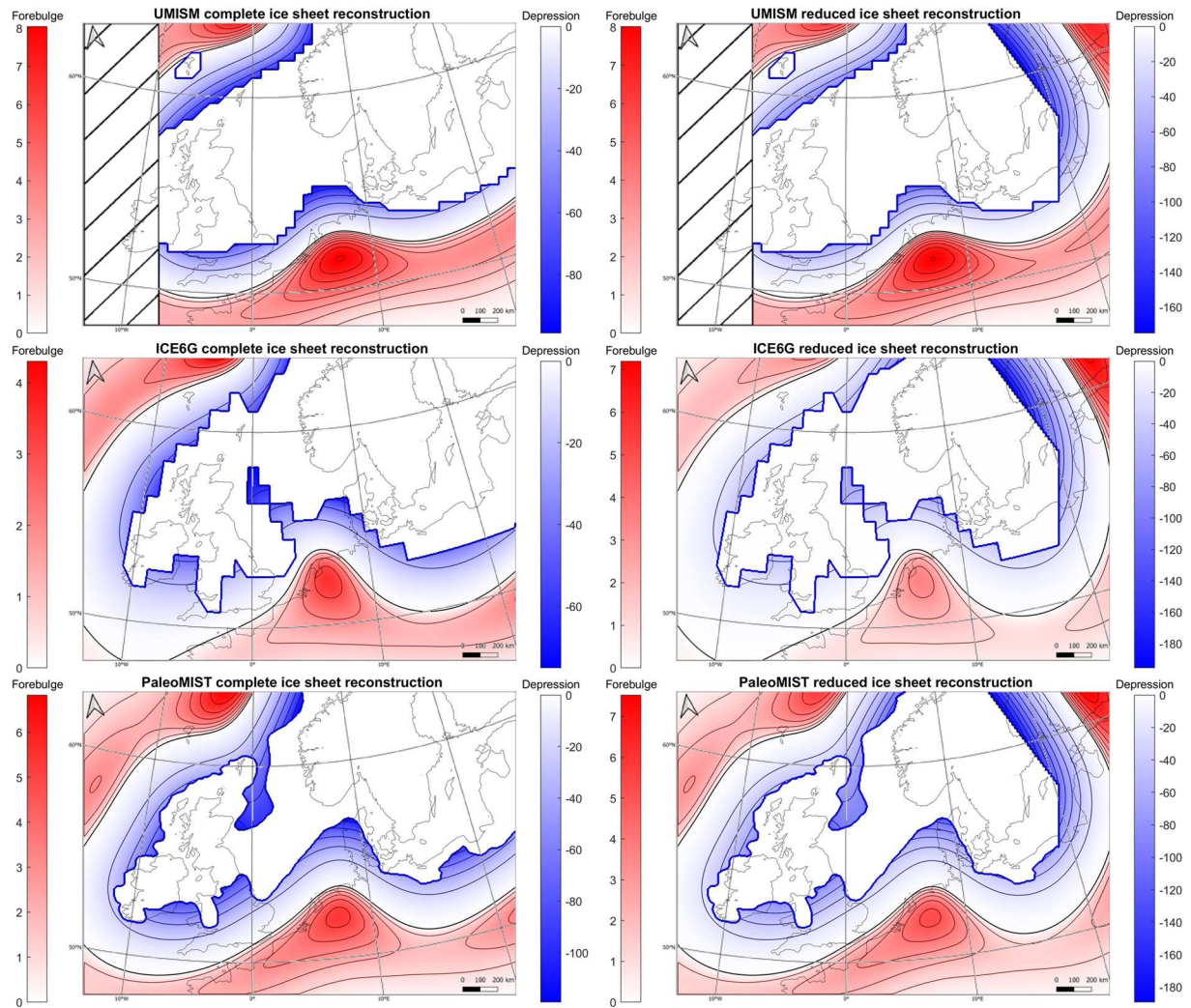


Figure A3: Comparison of vertical displacement at 20 ka based on the complete Eurasian ice sheet reconstructions (**left**) and based on the limited region used for the BRITICE-CHRONO reconstructions (**right**).

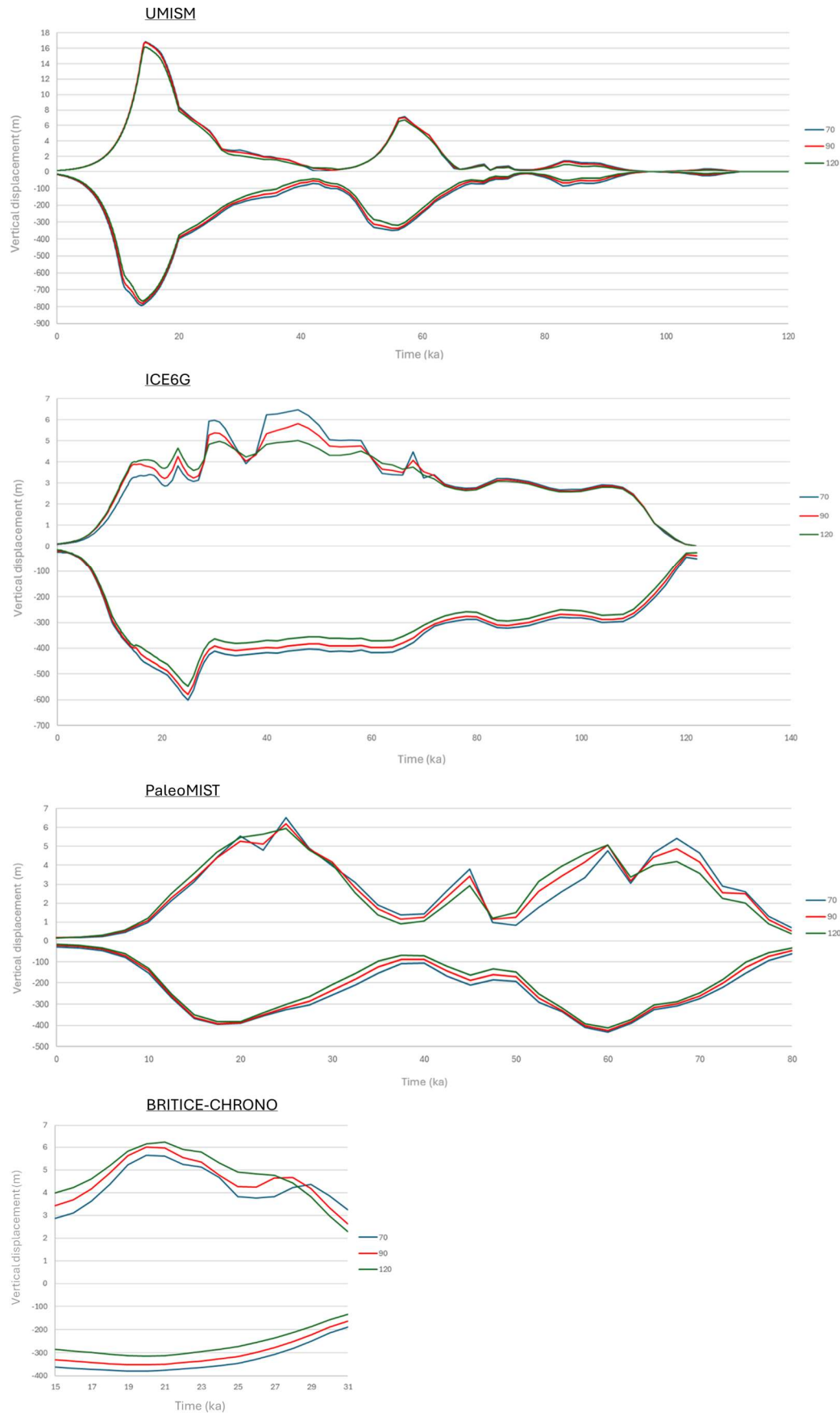
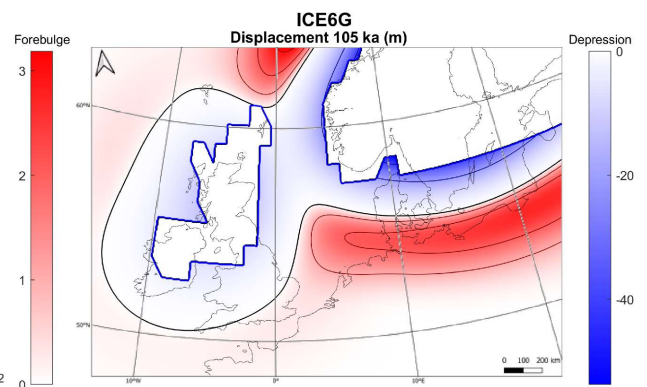
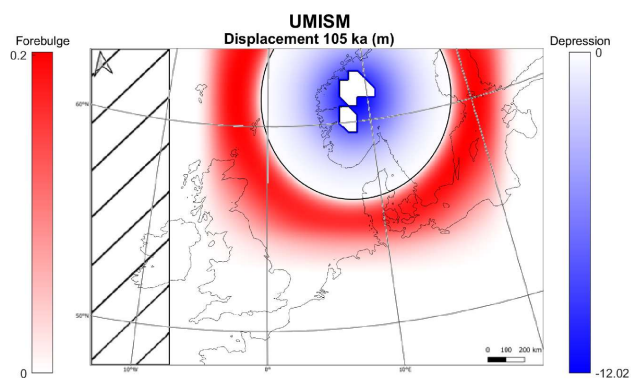
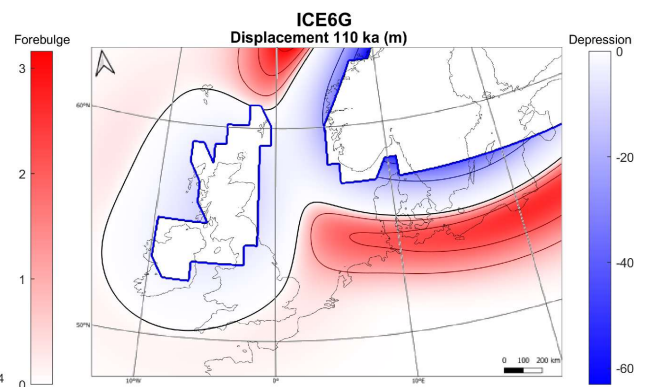
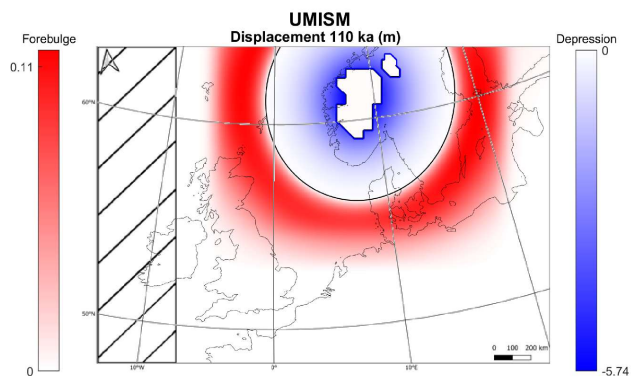
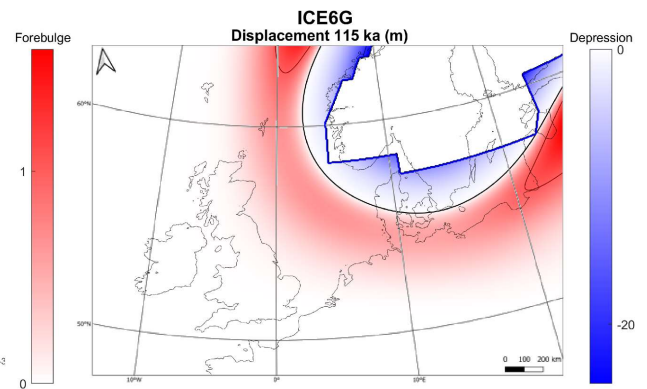
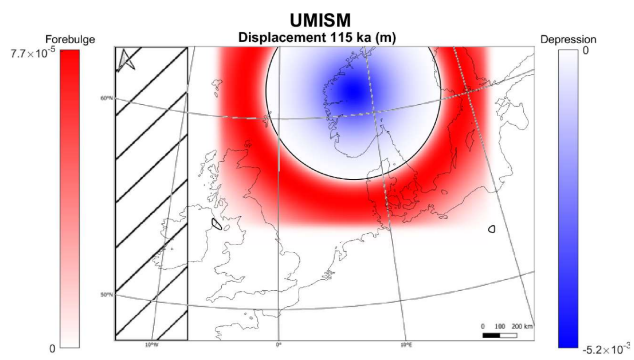
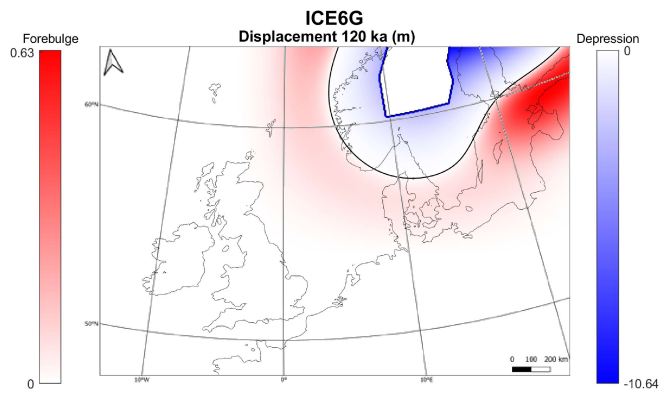
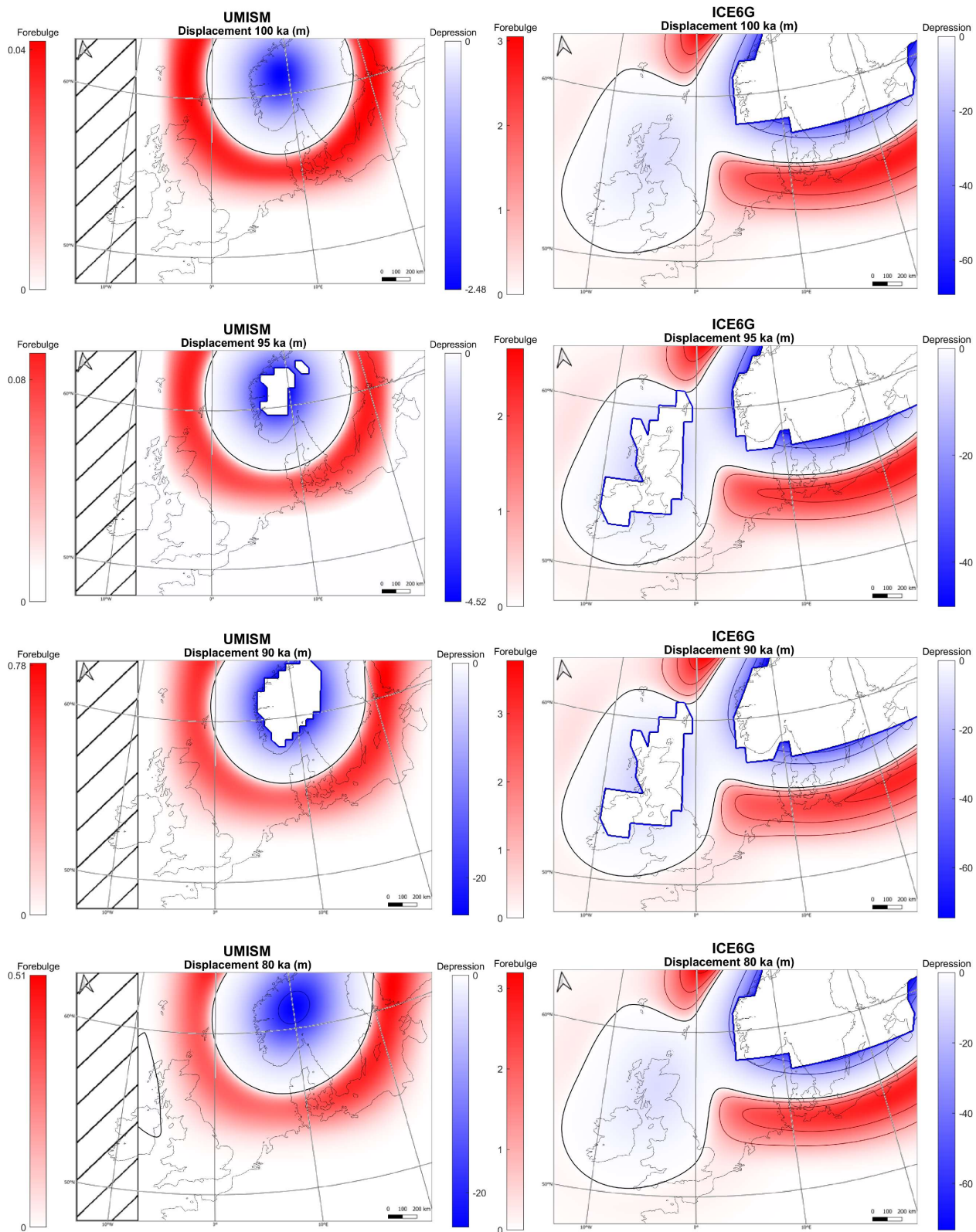


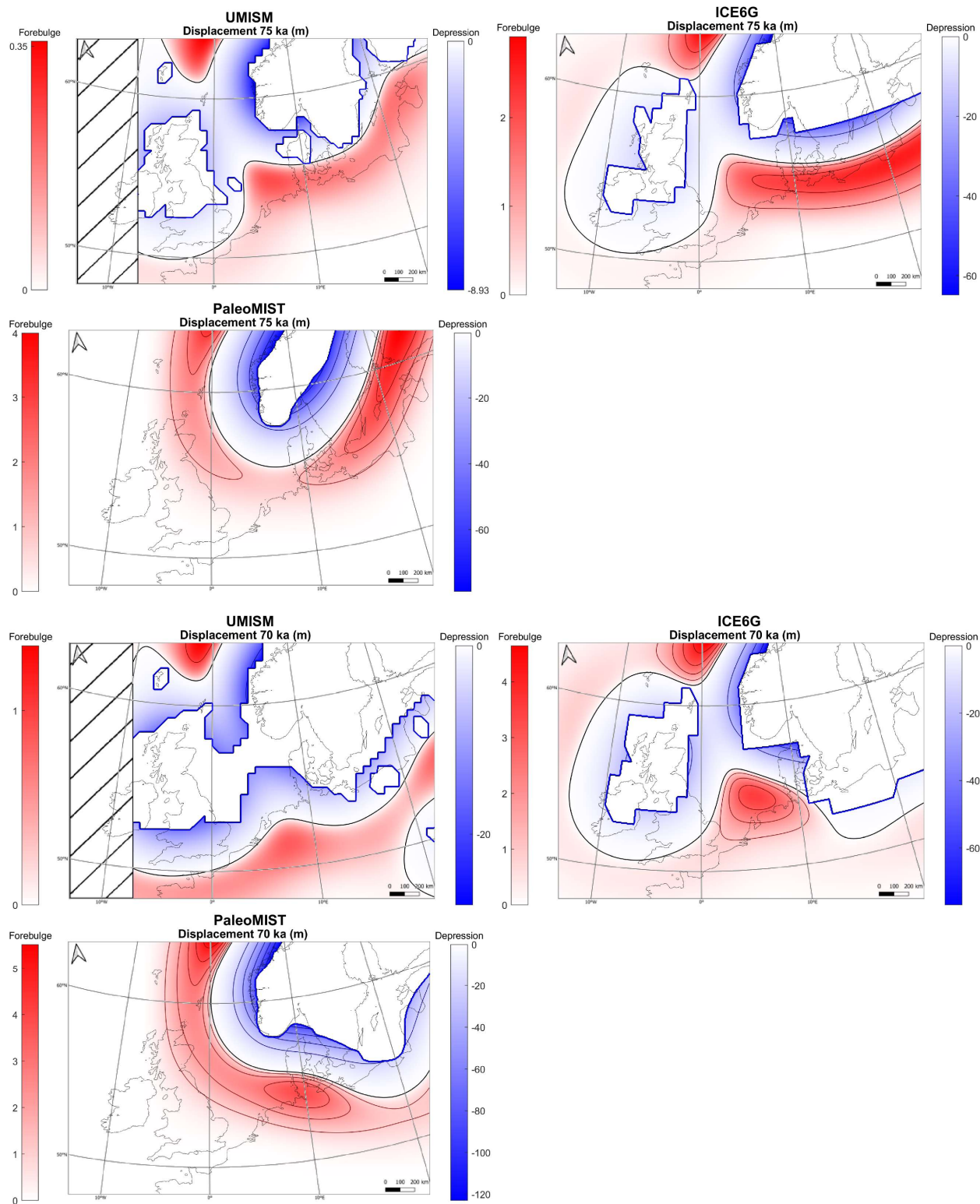
Figure A4: Evolution of maximal uplift in the region of interest and maximal depression over time for lithospheric thicknesses of 70 km, 90 km and 120 km.

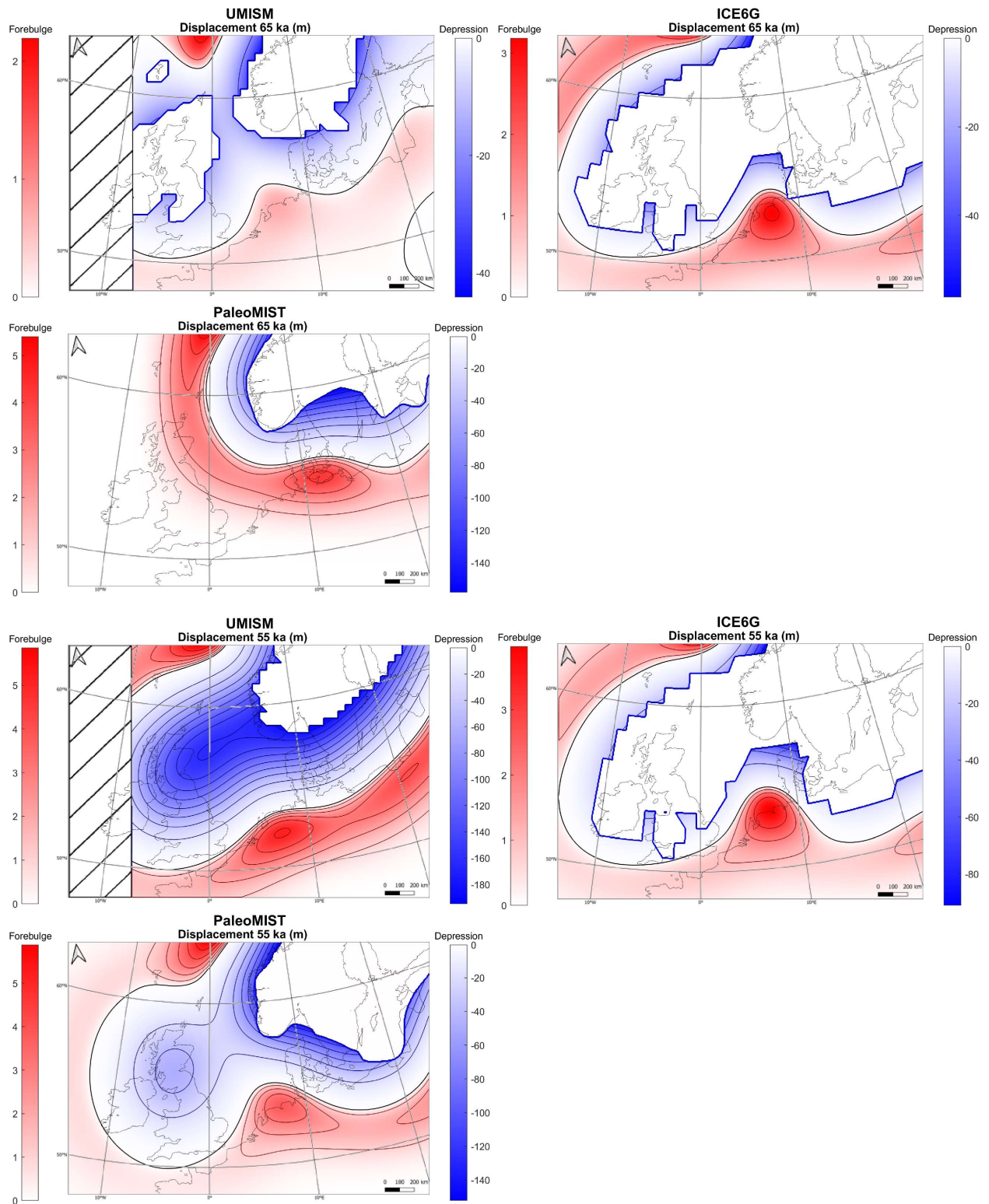


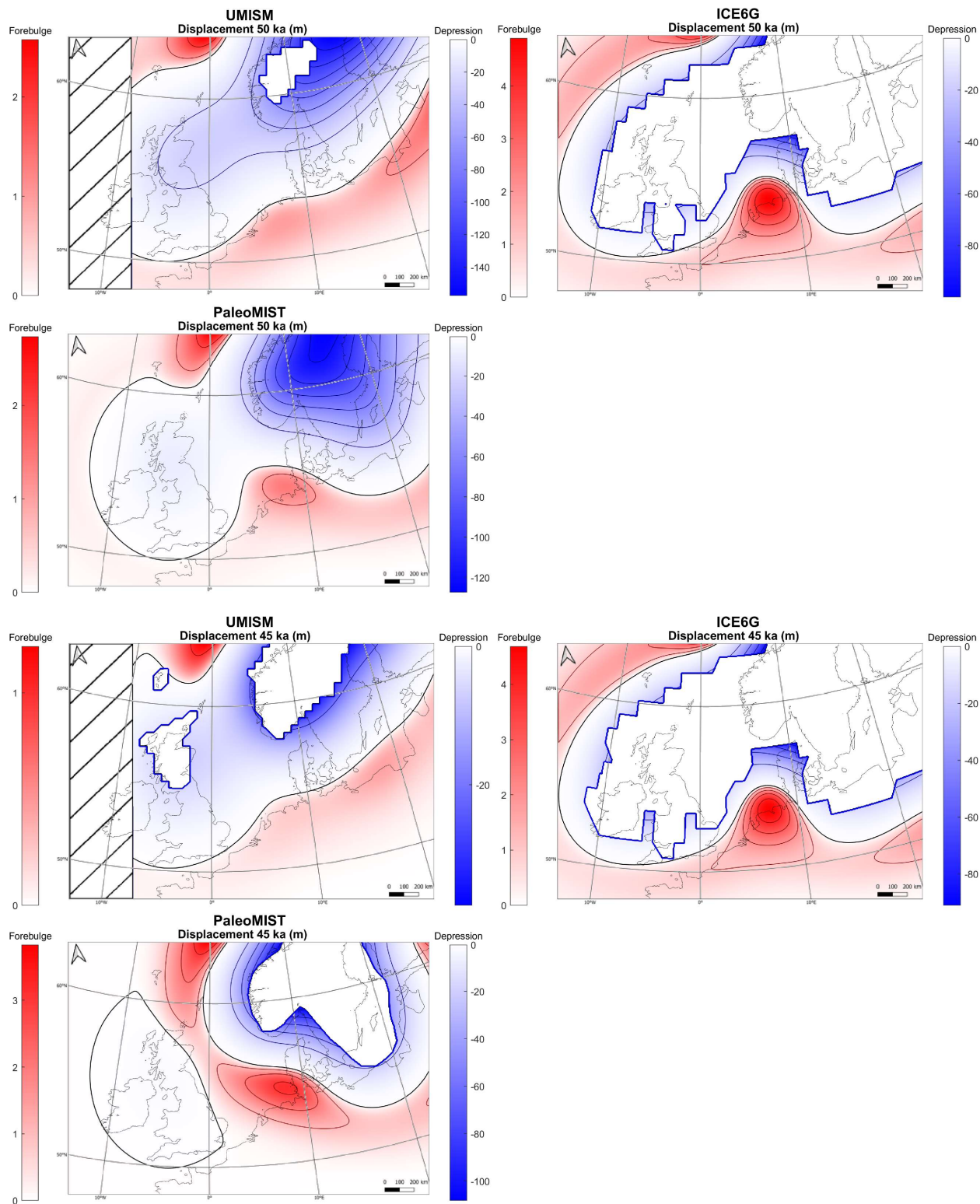
Figure A5: Evolution of maximal uplift in the region of interest and maximal depression over time for time-constants of 1000 years, 6000 years and 10 000 years.

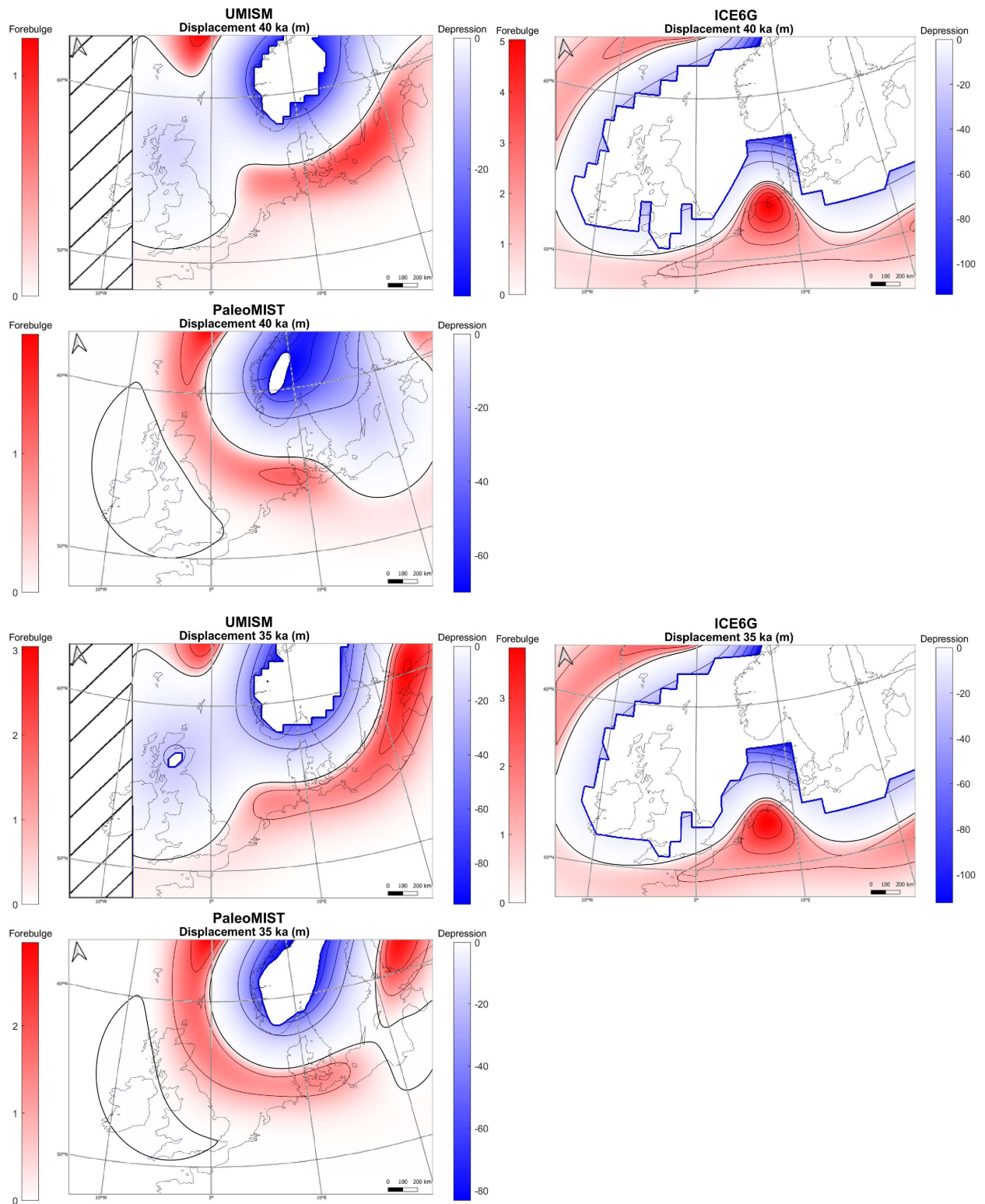


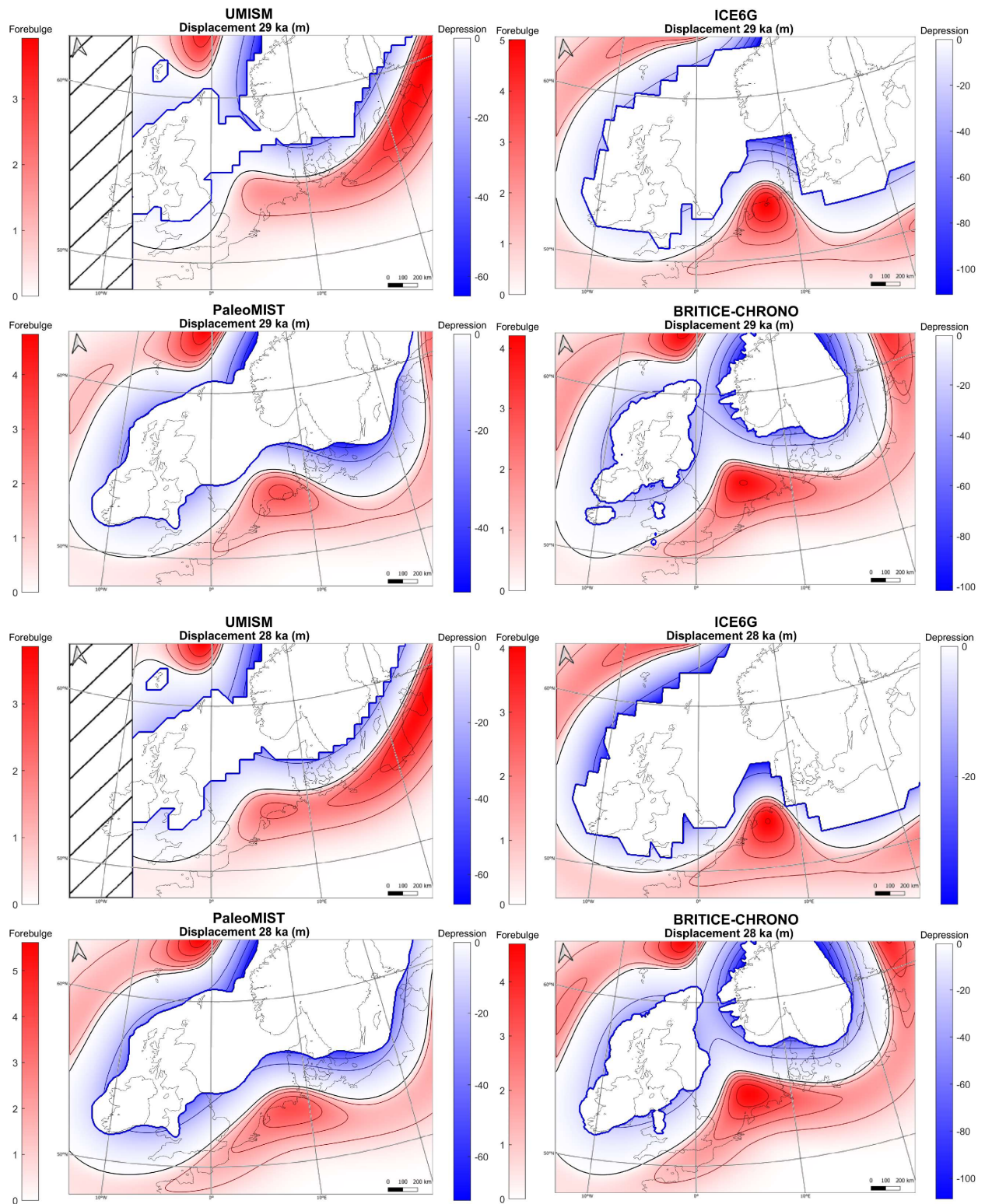


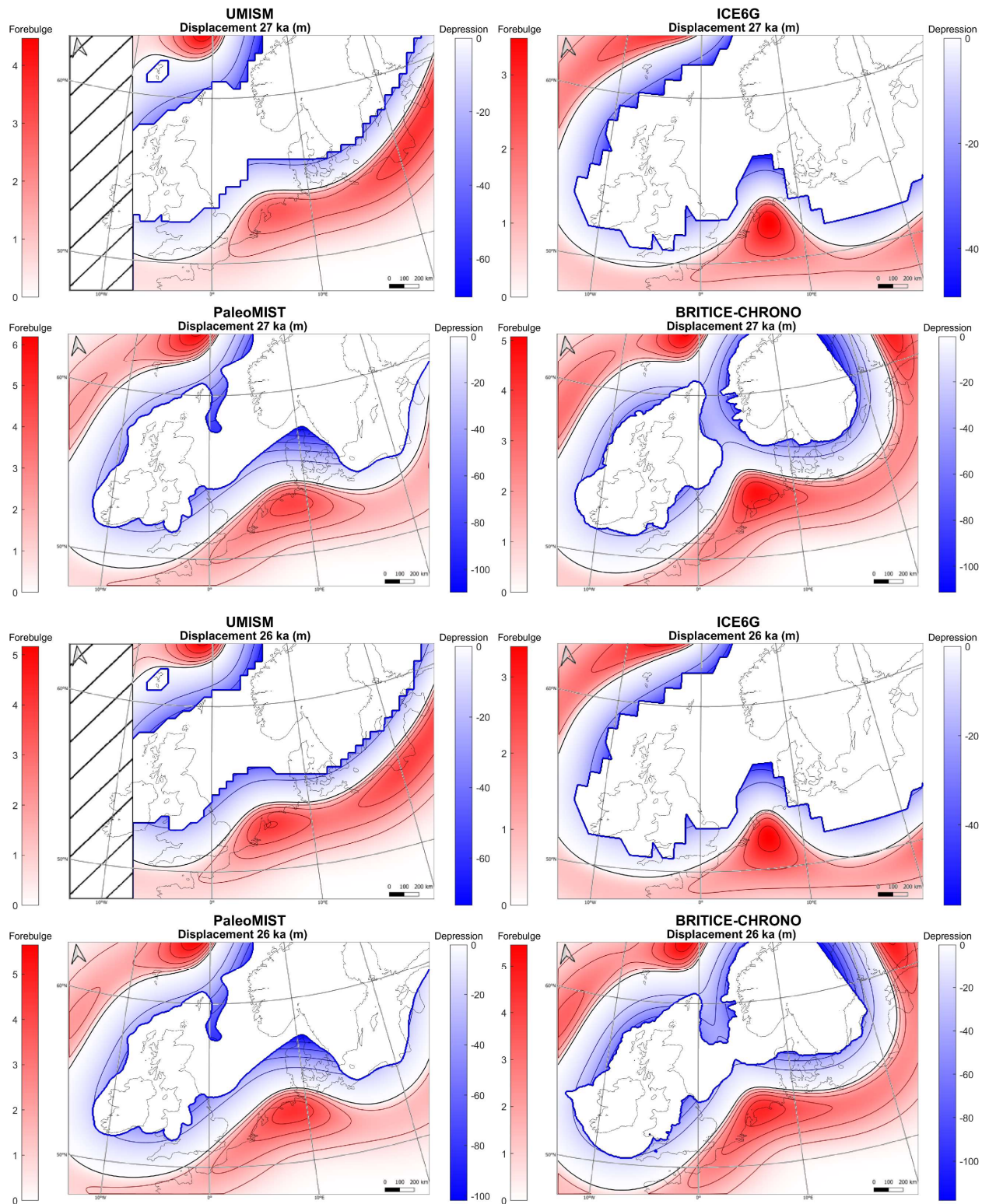


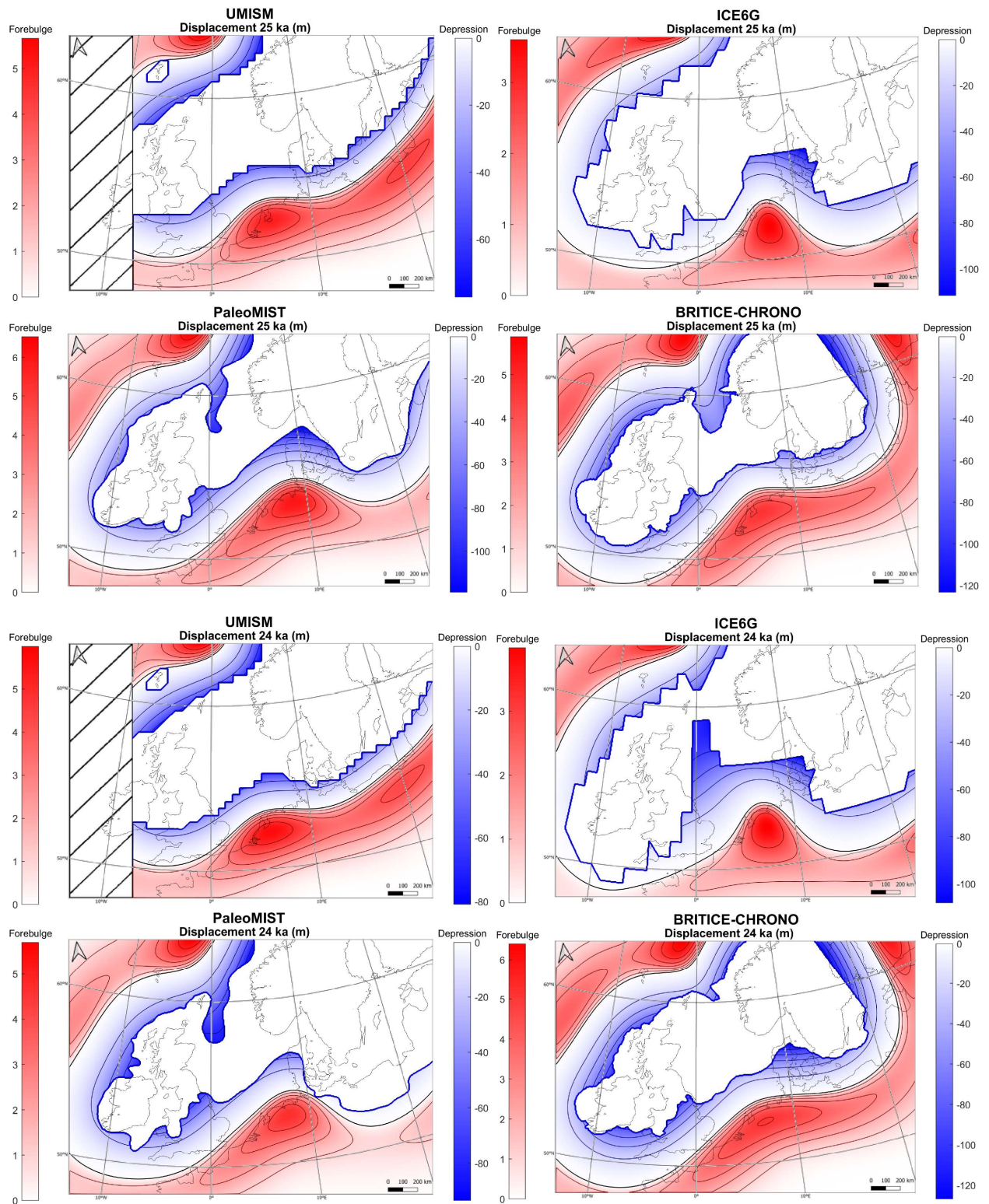


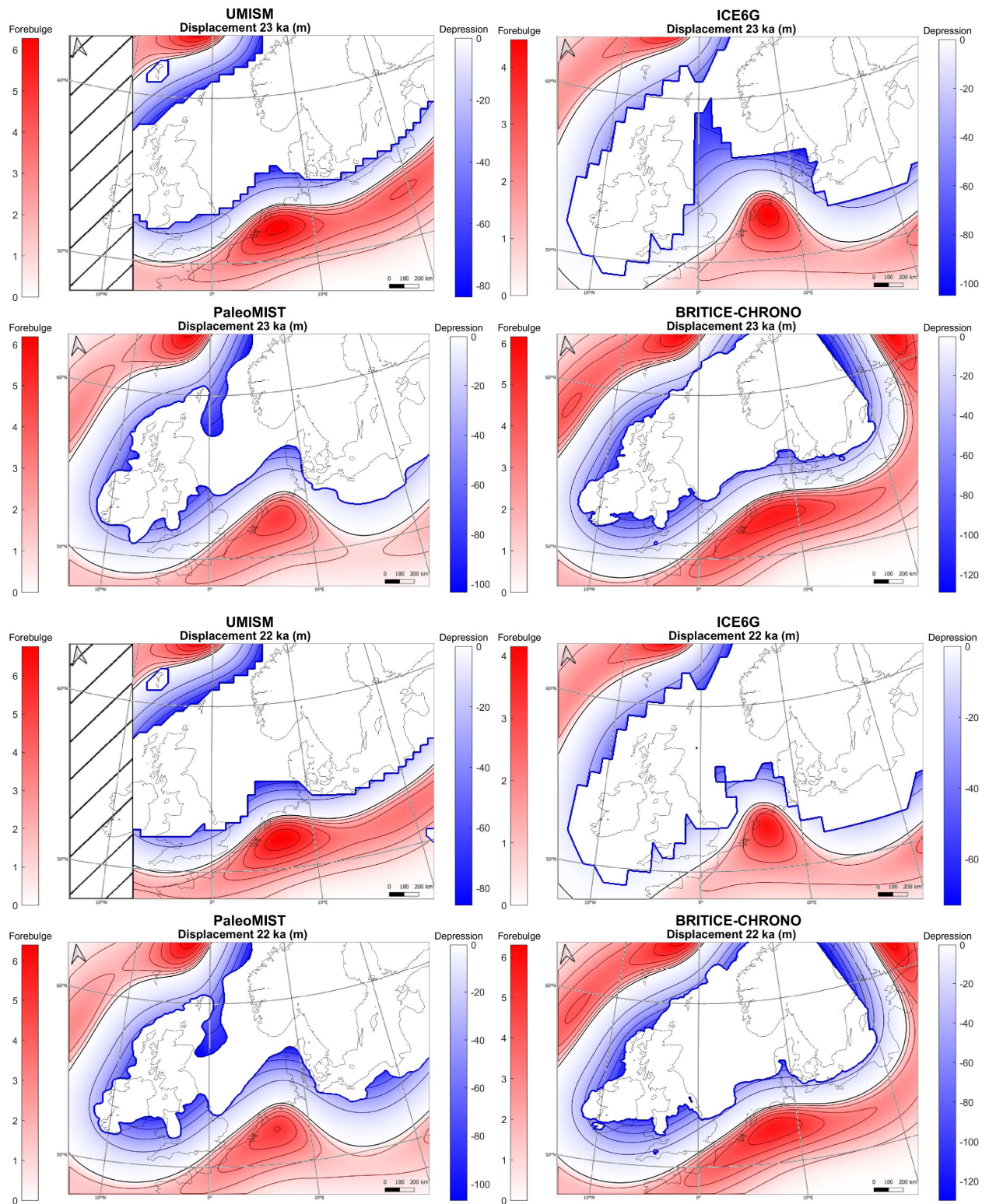


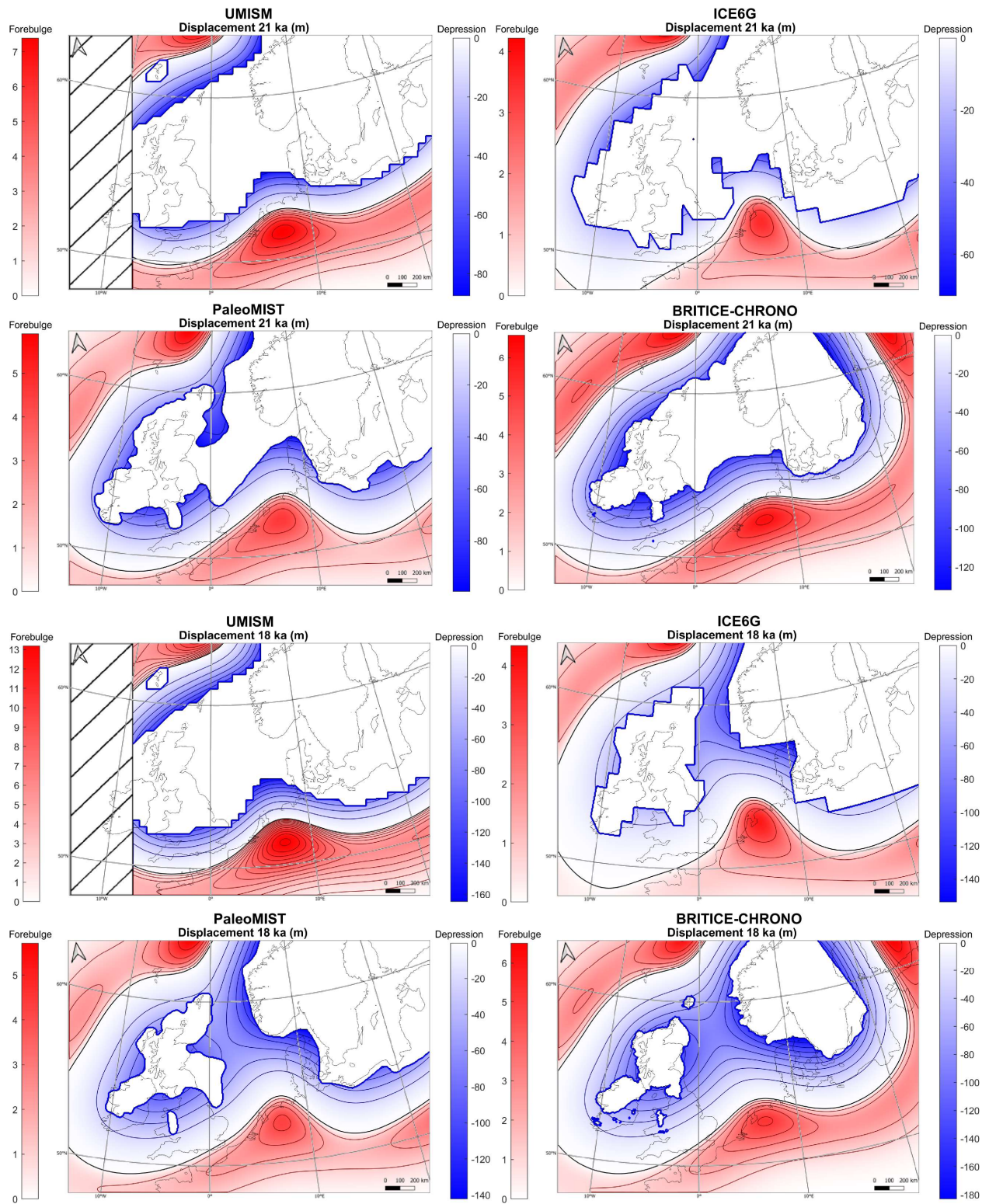


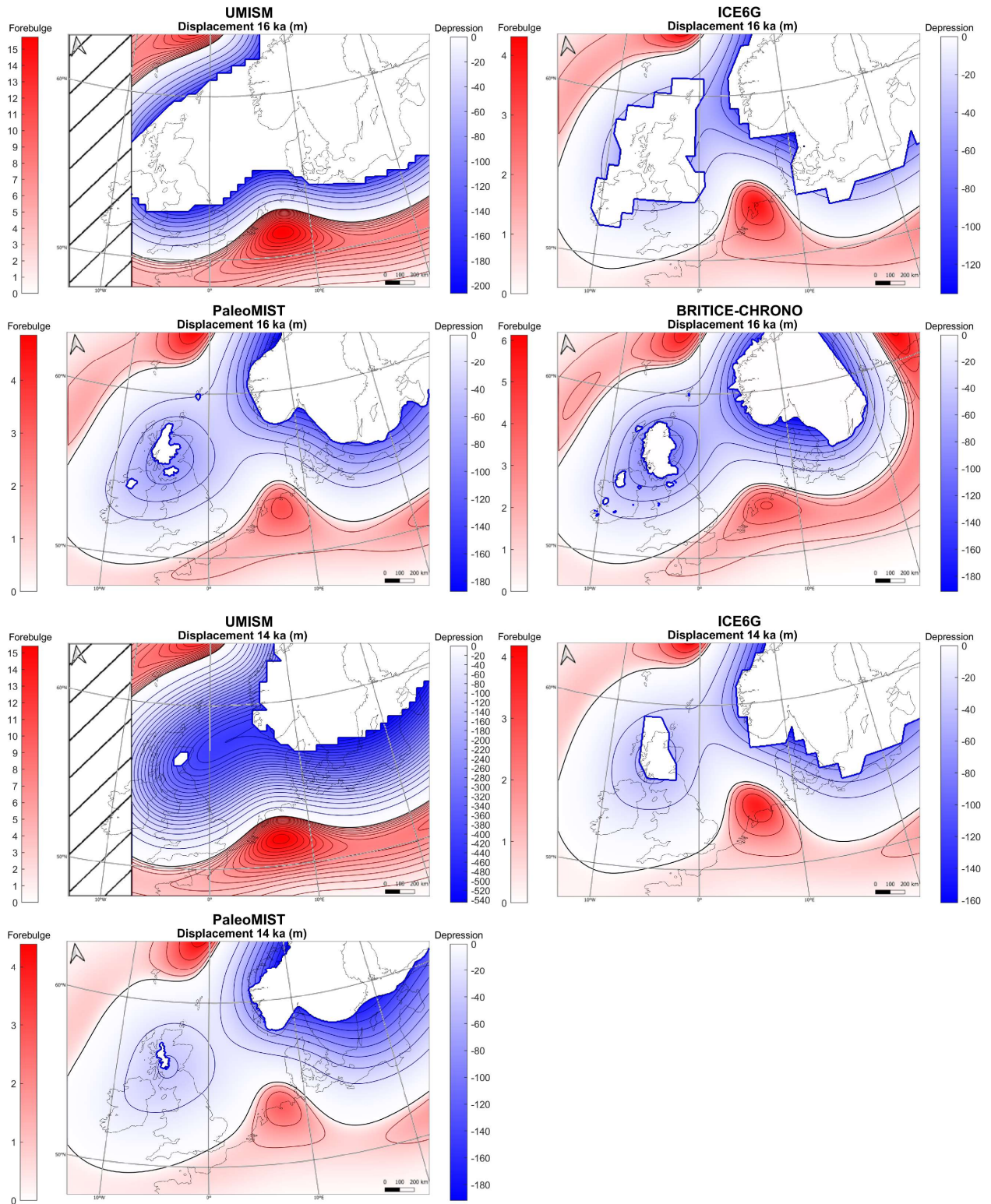


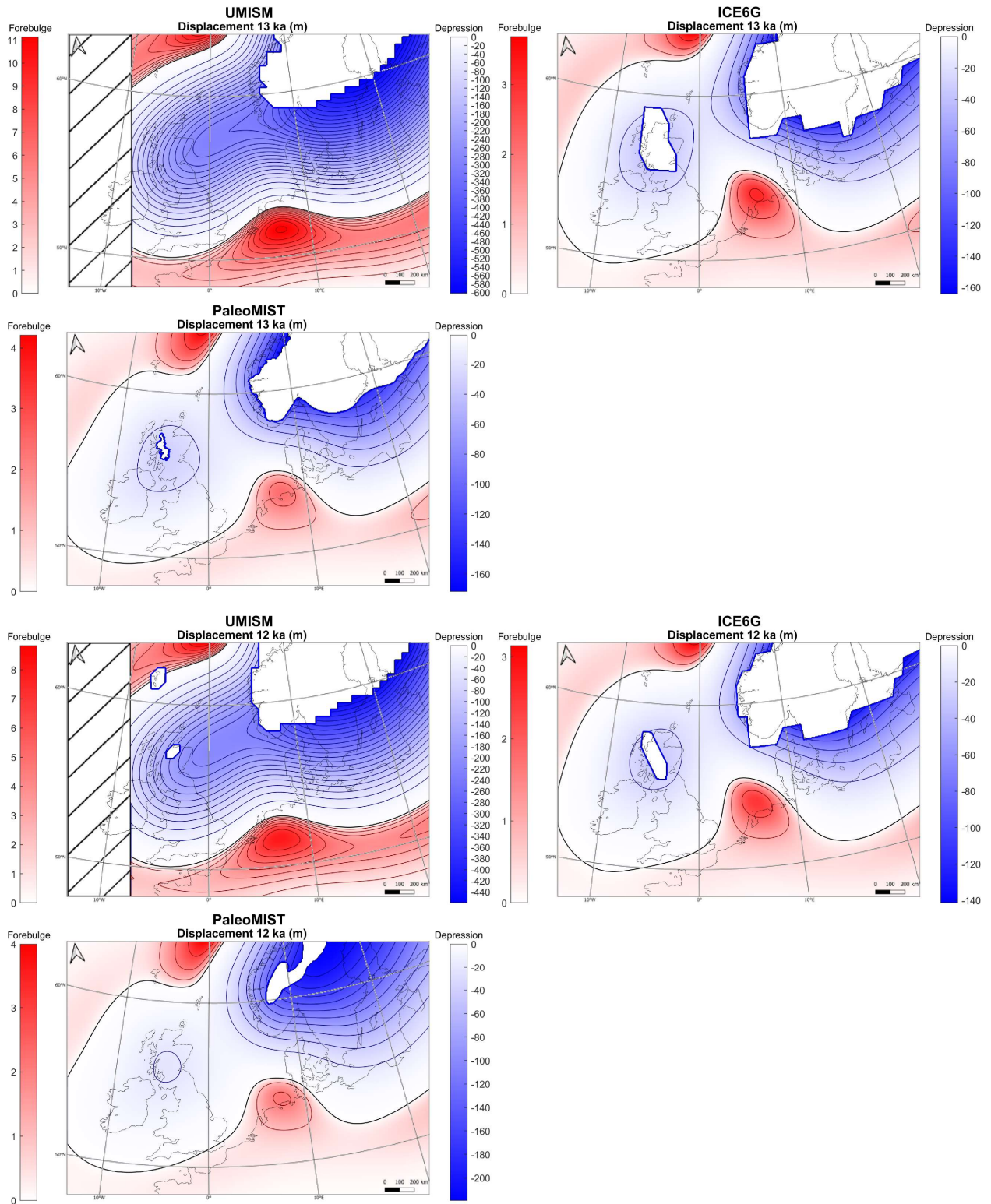


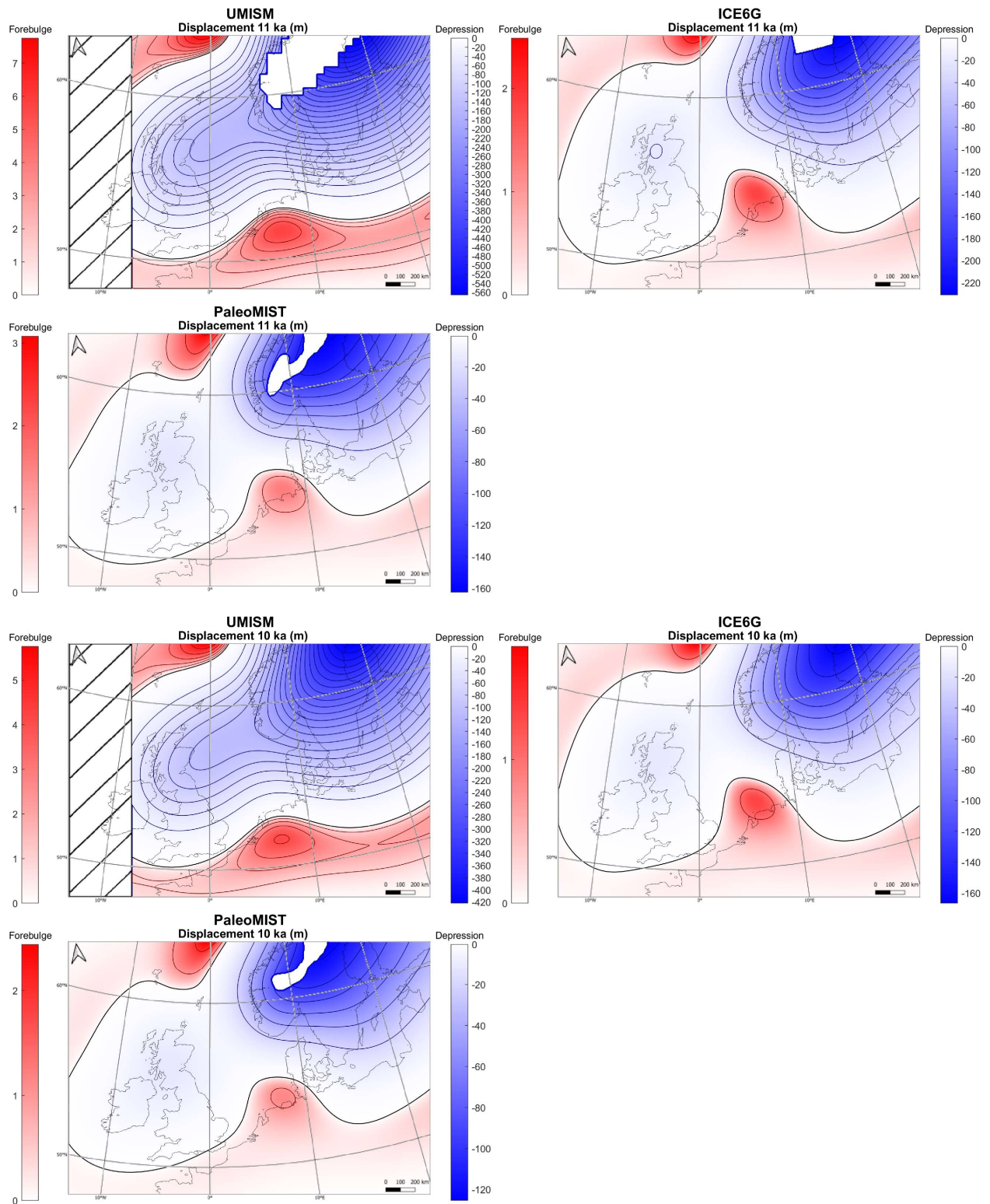












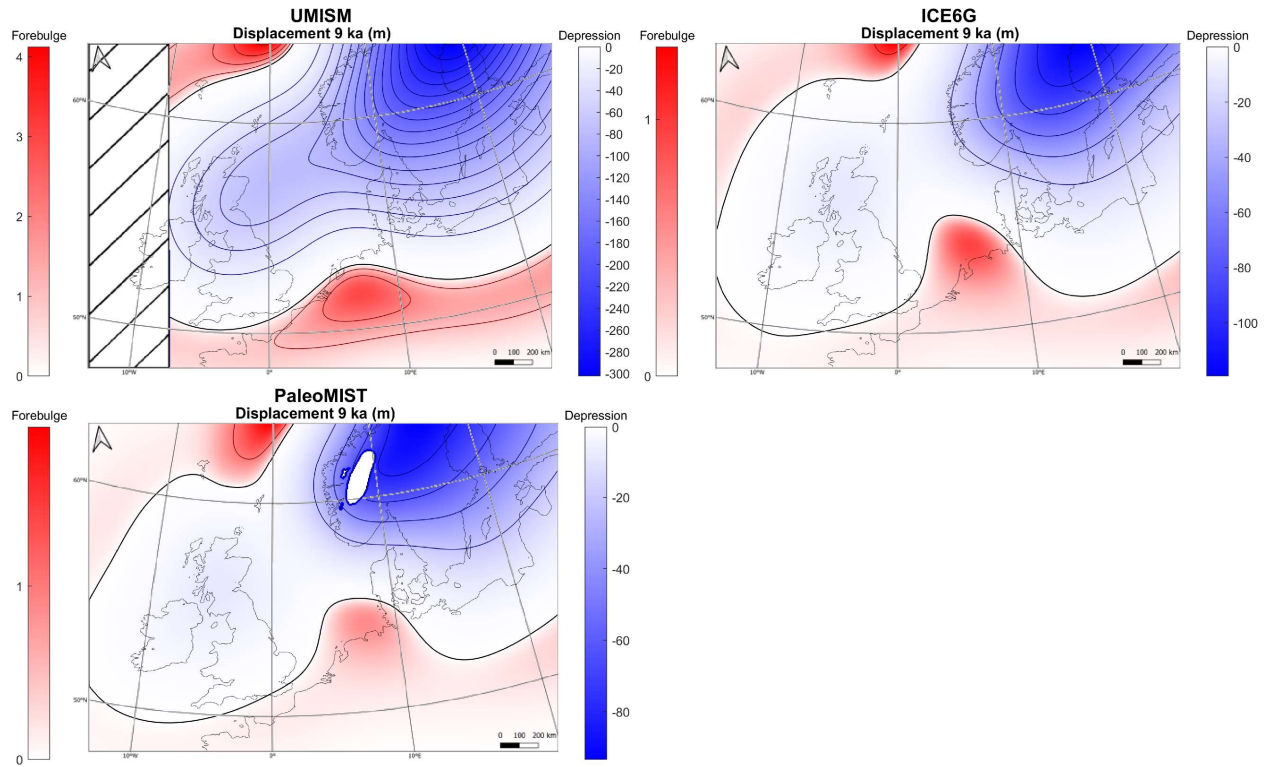


Figure A6: Reconstruction of the vertical displacement (m) based on the different ice sheet models for the time slices that are not shown in section 5.4, i.e.: 120 ka, 115 ka, 110 ka, 105 ka, 100 ka, 95 ka, 90 ka, 80 ka, 75 ka, 70 ka, 65 ka, 55 ka, 50 ka, 45 ka, 40 ka, 35 ka, 29 ka, 28 ka, 27 ka, 26 ka, 25 ka, 24 ka, 23 ka, 22 ka, 21 ka, 18 ka, 16 ka, 14 ka, 13 ka, 12 ka, 11 ka, 10 ka and 9 ka.

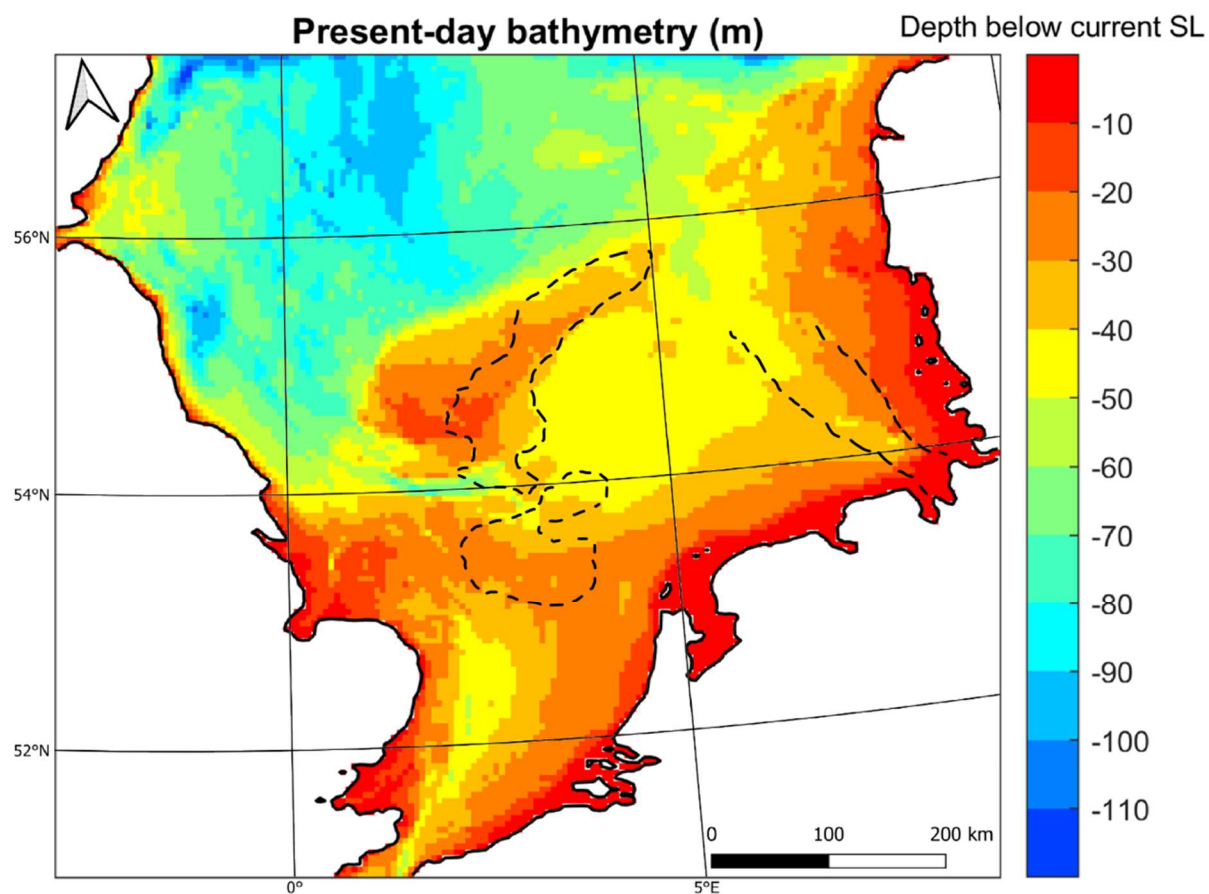


Figure A7: Current bathymetry map of the North Sea region with dotted lines indicating Holocene sediment units up to 15 to 20 m north of the Axial Channel, at the eastern side of Dogger Bank and along the Elbe Paleovalley.