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On the possibility of a long subglacial river under the north Greenland ice sheet

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Abstract. Does a long subglacial river with a source deep in the interior of the Greenland ice sheet, drain into the sea at the Petermann Glacier grounding line? Basal topographic data shows a segmented valley extending from Petermann Fjord into the centre of Greenland, however the locations of radar scan lines, used to create the bedrock topography data, indicate that valley discontinuity is due to data interpolation. Simulations where the valley is opened are used to investigate effects on basal water and ice sheet sliding. The simulations indicate that the opening of this valley results in an uninterrupted water pathway from the interior along the valley that alters ice sheet sliding in the Petermann catchment and in areas of west Greenland. Along its length, the path of the valley progresses gradually down an ice surface slope causing a lowering of ice overburden pressure that could enable water flow along its path. The fact that the valley base appears to be relatively flat and follows a path along the interior ice divide that intersects the east and west basal hydrological basins, is presented as evidence that its present day form developed as a consequence of the overlying ice sheet rather than prior to ice sheet inception. Though considerable uncertainty remains, the results are consistent with a present day active long subglacial river system. The results raise issues concerning the need to better observe, understand, and simulate the complicated basal hydrology of the Greenland and other ice sheets.

1 Introduction

The surface of the Greenland ice sheet holds visual clues to the topography of the bedrock, which in the interior can be below 2 to over 3 km of ice. Ekholm et al. (1998) found two, roughly 75 km long, elongated depressions in the surface of the ice that were connected by a "more than 100 km long, gently curving trench". Ice penetrating radar returns from the depressions were not "mirrorlike", which was considered a possible indication that subglacial water was being transported in the trench northward through a basal hydrological system. With improved topographic data Bamber et al. (2013) identified a "paleofluvial mega-canyon" that extends from central Greenland all the way to Petermann Fjord (Figure 1). The Ekholm et al. features are interior sections of this "canyon". While the feature was referred to as paleofluvial, they also suggested that the valley could have water flowing through sections of it today. In situ observations of water in the valley have not been obtained to date, nor are there current plans to acquire them. Since this "trench" or "canyon" takes a variety of cross-sectional forms along its length, in this article we will simply refer to it as a "valley".

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The BedMachine v3 basal topographic dataset (Morlighem et al., 2017) shows that the valley appears to be blocked by topographic rises at many points along its route (Figure 1b,c). However, based on the locations of the radar data lines that were used to generate this dataset and the limited extent of the valley bed elevation derived by mass conservation (two example regions are shown in Figure 1c and d), it is clear that these rises occur only in regions where data was not obtained. Therefore, we infer that these rises are due to kriging interpolation and that there is currently no evidence to suggest that this valley is filled (see Appendix A for further detail on BedMachine error estimates). This proposition is corroborated by the continuity of the signature of the valley on the surface of the ice across regions where the bed topography data has rises along the valley. This poses the question; are these rises damming subglacial water flow along this conduit in ice sheet model simulations?

If it is assumed that the valley is open, then the elevation of the bottom of the valley can be roughly determined by the points along its route where data was obtained. In (Figure 1b) The gaps in the valley where the valley base elevation rises above $-100\,\mathrm{m}$ occur where no data has been obtained and interpolation has smoothed out the valley. In fact, when you consider the smoothing effects of interpolation, a roughly level incised valley will only be resolved correctly exactly at the points where the data was obtained and everywhere else it will be shallower than it should be. Taking this into account a rough assessment is that the valley has a base that varies between $-250\,\mathrm{m}$ and $-500\,\mathrm{m}$ along its length from Interior to Petermann in Figure 1b.

The goal of our research is to investigate the impact of a continuous subglacial valley on the flow of basal water using a state-of-the-art ice sheet model. In addition, the effects on ice sheet sliding resulting from opening this valley in the basal topographic data are investigated.

2 Model and Methods

1) SICOPOLIS

The simulations use the SImulation COde for POLythermal Ice Sheets (SICOPOLIS, www.sicopolis.net) version 5-dev, a polythermal ice sheet model originally created by Greve (1995, 1997). We employ the shallow-ice approximation for grounded ice and the one-layer melting-CTS enthalpy method by Greve and Blatter (2016) for solving the thermodynamics. All results presented in this paper represent the output for the year 1990. To obtain a suitable 1990 ice condition, a series of simulations are run for a spin-up period over the last glacial/interglacial cycle (134 ka). The set-up is detailed in depth in Greve (2019, section 3.1) and summarized here.

2) Spin-up

The spin-up runs from 134 to 9 ka ago with a horizontal resolution of 10 km, and from 9 ka ago until today (1990) with 5 km resolution. The main forcing is the surface temperature anomaly derived from the δ^{18}O record of the NGRIP ice core (Nielsen et al., 2018), modified by a surface temperature anomaly derived for the GISP2 site for the final 4 ka (Kobashi et al., 2011). Since it is very difficult to reproduce the 1990 observed ice sheet geometry in a freely evolving simulation, corrections in thickness are made using the surface mass balance to reach an ice sheet shape consistent with observations.

3) Basal hydrology

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In contrast to the simulations by Greve (2019), we use a basal sliding law that incorporates basal hydrology. The hydrology model is coupled to the ice dynamics using a modified version of a Weertman-type sliding law proposed by Kleiner and Humbert (2014) with the parameters determined by Calov et al. (2018). The flux and storage of water in the subglacial hydrology model is governed by both the water pressure and the "elevation potential" which when considered together is known as the hydraulic potential (Shreve, 1972; Le Brocq et al., 2006). The basal melt rates from SICOPOLIS are used as the water input for the routing scheme and there is no basal water source from ice sheet surface melting. It is assumed that the water is moving in a layer only a few mm thick as a distributed water film where the water pressure and ice overburden pressure are in equilibrium. As such there are no subglacial rivers in the current formulation and so the results here are presented with respect to potential water routing changes rather than simulated subglacial river development. Nonetheless, where the film is thickest along an uninterrupted path to an ocean entry is considered to be a sign of a basal environment with an increased likelihood of consisting of some form of river system. The flux-routing method requires that local sinks and flat areas are removed, and this is done using a Priority-Flood algorithm which fills depressions and adds a small gradient, using the method of Barnes et al. (2014) to a depth of 10 m to account for subglacial lakes.

4) Bedrock modifications

The bedrock topographic data is altered in a way to ensure that the valley is open from Interior to Petermann in Figure 2b at an average depth of around $-400\,\mathrm{m}$. The case with the standard topography is referred to as "Control" and the case with the open valley is as "Valley".

The topographic modification is done using a flow-oriented interpolation scheme. Given the BedMachine topography and flight lines (Figure 1), a polygon of the rough location of the valley is drawn using a geographic information system (GIS). A buffer of 200 kilometres around this polygon is created and measurements falling into this buffer, but not into the valley polygon, are used to interpolate a new BedMachine, with the same procedure and parameters as Morlighem et al. (2017). The polygon outlining the valley is converted to a centerline, and then described through a spline. Data points situated within the valley polygon are converted from their Cartesian coordinates (x,y) towards this flow-oriented coordinate (s,n) system. This is a common reference frame, where s describes the distance of the thalweg and n is the normal in right-hand direction, and outperforms ordinary kriging and others when used in with an anistropic adjustment (Merwade et al., 2006). Here the anisotropic factor was 10, with a nugget of 20 metres, a sill of 30 metres and a range of 300 metres. Further details on the procedure are detailed in Legleiter and Kyriakidis (2008).

For three regions, saddles are present along the valley. Bounding boxes over these saddlepoints are drawn that cover both the saddle and trenches within. Over these subsets a watershed algorithm called Maximally Stable Extremal Region (MSER, Donoser and Bischof (2006)) tracking is run to detect the trenches to be connected. Pixels between both trenches of the saddle were found through connecting mathematical morphologic operations. These selected pixels are then adjusted by linear interpolation of the elevation information in the trenches to make a seamless passage.

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3 Results

All results presented here represent the SICOPOLIS simulation for the year 1990 at 5 km horizontal resolution as detailed above. To examine the effect that the introduction of an uninterrupted valley has on the simulated ice sheet, an analysis of the basal water depth, basal water flux, ice sheet velocity are presented.

For north Greenland the simulated basal water depth is affected by the introduction of the valley in several ways. In Figure 3a it can be seen that the standard topography produces 3 main quasi-linear sections of deeper basal water along the valley but there are clear gaps between these areas. In contrast, when a continuous valley is introduced (Figure 3b) the basal water depth is both deeper and uninterrupted along the length of the valley all the way to Petermann Fjord. The thickest water depth (>0.01 m) along the valley route occurs where the Priority Flood algorithm has been activated to represent subglacial lakes. In most interior areas away from the valley there is little or no change in the basal water depth. In particular, there is little to no effect on the basal water pathways associated with the North East Greenland Ice Stream (NEGIS). The interior basal water changes are relatively small because the valley follows a path close to the boundary between the east and west catchments and thus has less influence on them.

To obtain a clearer picture of the changes to the basal water from the introduction of the valley, the difference in basal water between these cases (Valley – Control) is shown in Figure 3c. Doing this reveals that the increased water within the valley is surrounded predominantly by a reduction in basal water adjacent to the valley. Basal water reduction extends to some regions away from the valley, particularly to the west of the interior section. The one region of increased basal water outside of the valley is a region that extends towards the Petermann Ice Stream (NEEM zone in Figure 3d). The effect of these changes in the Petermann catchment is to redistribute the basal water into a narrower and deeper plume that can also be seen in Figure 3b.

To examine how the movement of basal water is altered by the introduction of the valley, the basal water flux is presented in Figure 4. The valley causes a shift in basal water flux in its near vicinity, with increased flux within the valley base. Water flux streamlines give an indication that water flux is generally down valley with streamlines getting "stuck" in the valley along certain sections, however confirmation that simulated water travels down the entire length of the valley will require future work. In the Petermann catchment region, the increase in water flux in the valley causes a shift downstream in the subglacial water distribution where the valley crosses a region of increased flux out of the interior (NEEM zone in Figure 4). Along this section where the valley is oriented SSW to ENE the flux just NNW of the valley is reduced in the south and then increased to the north in the region upstream of the interior part of the Petermann Ice Stream. This is consistent with the increase in basal water discussed earlier. The simulated effect of the valley is therefore to focus maximum water flux into a narrower but more elongated region that is also shifted eastward. Sensitivity tests (Appendix B) indicate that the location and magnitude of this water flux out of the valley is sensitive to the valley depth due to its consequent effect on the steepness of the valley sides.

The rule of thumb helpful for understanding the relative roles on subglacial water flow of ice overburden pressure and basal topography, is that the topographic gradient needs to be 11 times greater than, and opposing, the ice surface slope for water flowing along the bed to start accumulating (e.g., Cuffey and Paterson, 2010). Figure 5 indicates that the along-valley component of the ice surface slope is near flat over the interior section of the valley before gradually sloping down-valley all

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the way to Petermann Fjord. This is an indication that the ice overburden pressure distribution does not oppose the flow of water towards the north and should generally reinforce it for the northern half of the introduced valley. As the ice sheet surface is extremely gently sloping in the interior, the basal topography and water fluxes will have a greater influence on subglacial water routing than around the edges of the ice sheet. In this situation the basal topography needs to be either sloping downward towards the north, or near-flat, for water to flow northward. It appears it could be the latter, which, if true, suggests that this is unlikely to be paleo river valley because it is in a location, and in a form, that favours present day basal water routing.

The final results presented here concern the influence of the valley on ice sheet sliding. Figure 6 shows the ice surface velocity difference between Valley and Control to highlight the locations where Valley increases or decreases sliding. In the interior the sliding changes are relatively modest with a region of reduced sliding of 1 to $10\,\mathrm{m\,a^{-1}}$ to the west of the valley increasing towards the coast. This has two smaller regions of sliding increases of 1 to $10\,\mathrm{m\,a^{-1}}$ to the north and south. The region of decreased sliding is consistent with a region downstream of a reduction in basal water depth (Figure 4c). There is no increase in basal water depth associated with the regions of modest increased sliding so these two regions may be occurring due to mass balance to compensate for the reduced westward mass flux over the region of reduced sliding.

The more pronounced sliding changes occur in the Petermann catchment (Figure 6b) where the greatest sliding increases are 50–$60\,\mathrm{m\,a^{-1}}$ around the eastern half of the Petermann Glacier grounding line in the Valley simulation. Sliding increases to a lesser extent over both the valley and the Petermann Ice Stream leading to two separate branches of increased sliding. These pathways are consistent with the redistribution of basal water seen in Figure 3. The valley increases water flux in the region of Petermann ice stream probably due to water piracy from the valley as it bends to track more perpendicular to the ice-sheet slope.

4 Discussion

The formation of subglacial river channels has long been known to be a fundamental evolutionary property of subglacial water flow. Röthlisberger (1972) and Shreve (1972) proposed that subglacial water can form channels that cut upwards into the ice. These have come to be referred to as “R-channels”. Channeling of subglacial water occurs because the initial film of water at the ice base can become unstable due to viscous dissipation which initiates the development of R-channels. For this transition to occur the discharge has to increase beyond a threshold where you get a switch from film or cavitation to R-channels (Schoof, 2010).

The R-channel theory requires a hard bed and therefore ignores potential bed erosion from such a channel. If the channels are over a sufficiently hard bedrock and move position then this assumption should hold, however if they remain quasi-stationary, due to basal topography or persistent ice overburden pressure distribution influences, then the effects of bedrock erosion or sediment deposition should manifest. In the case where there is sufficient sediment deposition in a stationary channel an esker could develop lifting the water channel above the bedrock. In the case where there is not sufficient sediment deposition, erosion downwards into the bed will inevitably occur if the channel remains stationary. Nye (1973) suggested that channels incised upwards into the ice are more vulnerable to closure due to ice overburden pressure and ice movement and concluded

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that channels incised into the rock were expected to be much longer-lived than channels incised upwards into the ice. Nye concludes that "while there may be temporary channels incised upwards into the ice, there will be comparatively permanent channels cut downwards into the rock bed."

There are other reasons to suppose that flowing water in a subglacial valley would be the favoured mode of water transport under an ice sheet. These are associated with the resistance to freezing of water in such a channel. Firstly, because the ice sheet surface will likely not have as pronounced an indentation as an incised valley in the bedrock beneath has, the ice thickness and therefore ice overburden pressure will be higher at the base of the valley than under the ice over the surrounding bedrock. This increases the likelihood of the ice at the base of a valley being at the pressure melting point.

Secondly, an incised valley under an ice sheet will tend to have higher geothermal heat flux at its base and particularly along its sides. This is because of the distortion of isotherms beneath the valley that increases the isotherm gradient and consequently also the heat flux (e.g., van der Veen et al., 2007). For example, Lees (1910) found that a depth to width ratio of 0.5 increases heat flux by around 50% while van der Veen et al. (2007) found a 100% increase in heat flux in a Jakobshavn-scale idealized simulation. Essentially the deeper and steeper the valley, the greater the heat flux increase will be at the valley bottom. In the case of a melting ice base, if the valley sides are steep enough to overcome any opposing overburden pressure forcing on water flow, then meltwater will collect at the base of the valley which can also further enhance melting there.

Thirdly, flowing water generates heat through frictional heating, increasing the temperature of the water. It also transfers heat downstream so factors such as the upstream water heat capacity and the duration of cooling will determine whether the water freezes or not. Fourthly, applicable to all basal water, is one of the odd properties of water. As water cools below $4°\mathrm{C}$ it starts to become less dense causing the coldest water to rise up. Thus, freezing occurs at the top, which in the scenario of a subglacial river would be at the base of the ice sheet. This new ice acts to insulate the liquid water below as is observed in frozen rivers and lakes. This allows liquid water to persist beneath the ice in situations where it would not if water did not have this property.

These are some factors that may enable flowing water to continue in a valley under an ice sheet even in some situations when the ice sheet is frozen to the bed outside of the valley. These factors are presented here as indicators that positive feedback processes may exist that favour the development of subglacial river channels incised into the bed.

The model results indicate that the valley follows a path down a gentle ice surface slope (Figure 5) which would imply that the ice overburden pressure lowers as the valley progresses towards Petermann Fjord. In this scenario, if an open water channel were to be maintained along a relatively flat uninterrupted valley base, the overburden pressure should propel water towards the ocean outlet. This is providing that water does not escape out the sides of the valley as appears to happen as the valley crosses the upper Petermann catchment (NEEM zone in Figure 5). If this propulsion occurs, a possibility is that it occurs sporadically through the build-up, and release, of water in reservoirs along the channel route.

The results also indicate that the course of the valley in the interior runs close to the boundary of the east and west subglacial hydrological catchments (Figure 3a,b). This catchment boundary occurs in this region because it is below the gentle northward ridge of highest surface ice (Figure 5) which forces the division between the east and west basal hydrological basins. A consequence of this positioning is that the valley enters the Petermann surface catchment at its southernmost location (Figure 1a).

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Because the gentle ice ridge is the region where hydro potential directed towards the east or west is lowest, it represents the most favourable location in the interior of northern Greenland for a hydrological pathway towards the north to develop. This may be a further indication that the path of this valley has developed as a consequence of the ice overlaying it.

The current simulations do not include rivers in the hydrology module. Rivers could funnel greater amounts of water away from, and to, particular locations leading to focused areas of suppressed and enhanced sliding. The valley could extend further southward and there is evidence of several tributaries that could increase the main valley's discharge potential. The most prominent possible tributary, shown as "Tributary" in Figure 1 projects towards a region of high basal melt associated with the interior of the NEGIS and could therefore be a significant additional source of basal meltwater.

The valley originates from under some of the thickest and highest ice in Greenland (Figure 5). The valley we have inserted in the simulations has its upper end at "Interior" but given the basal topographic basin at "Basin" (Figure 1), could it be possible that basal water is transported from Basin to Interior? Between these two regions the ice surface slope is relatively flat so water flow should be more heavily influenced more by the basal topography. In this inter-basin region the basal topography is poorly resolved and it is unknown whether the ice sheet is frozen to the base. It is therefore unclear whether a basal water connection could exist between Basin and Interior. In the simulations presented here, Basin is frozen at the ice base and so no basal water is produced there (Figure 3a,b). This is due to the geothermal heating distribution used by SICOPOLIS which is, as with all Greenland geothermal distribution estimates to date, highly uncertain due to severely limited observations at the base (e.g., Rezvanbehbahani et al., 2017). If these basins are connected hydrologically, it could significantly extend the catchment of the valley and imply a subglacial river over 1600 km long. At present there is not enough data on the bedrock heat flux or topography to know if this is the case and the fact that we are in the dark on such a potentially large feature on the Earth's surface, expresses the importance of observation campaigns that can improve our understanding of the conditions at the bed.

It could be argued that the path of such a long basal valley down an ice surface slope is, in itself, evidence of a feature that has developed over a long period as a direct consequence of the ice sheet covering it. The logic behind this reasoning can be understood by asking the question; why would a paleo-valley in the bedrock below an ice sheet follow a path that is favourable for subglacial water transport under 2 to 3 km of ice today? A paleo-river that developed when the ice sheet was much smaller, or absent, would have formed when the topography was significantly different due to bedrock isostasy. In addition, the water flow would have been governed by gravity where conditions were ice-free. This different water flow environment would make it highly unlikely to follow a path that is today favourable for water transport from the deep interior all the way to the coast under a thick ice sheet. In addition, the apparent flatness of the valley base in the interior where the ice surface is relatively flat, is, just like any other river, the ultimate erosional and depositional form of a long-term active waterway. Due to bedrock isostasy there would, again, seem to be no reason for a paleo-river system to have a relatively flat base today. One can imagine that a paleo-river valley pushed down by the weight of the ice as the ice thickened would end up today having an uneven base that ascended and descended depending on the complicated evolution of the competing pressures from the ice and crustal rock. A paleo-river valley that ended up having a very long level base in this situation would be a remarkable coincidence. In the absence of adequate direct observations, perhaps the topographic form of the base of this valley could, with further work, help us deduce the likelihood of this being an active subglacial river.

5 Conclusions

The Greenland bedrock data indicates that a subglacial valley extends from Petermann Fjord into the center of Greenland. The valley is segmented along its route in the current bed topographic datasets used in ice sheet simulations. The rises occur where data is interpolated to fill in gaps between where radar has obtained reliable data. This suggests that the valley rises are not real. Therefore, simulation tests have been completed to investigate the consequences of removing these rises. Opening up the valley in SICOPOLIS simulations causes water to be re-routed leading to ice sheet sliding changes including modest slowing in interior west Greenland and an increase in the Petermann Ice Stream region. Since the model relies on a thin-film hydrology model, it is possible that much larger quantities of water are transported in this valley that could have a greater impact on the basal water distribution. The valley progresses gradually from thicker to thinner ice causing a lowering of ice overburden pressure that could enable water flow along its path towards the sea. If this is the case, some of the basal water routed to Petermann Fjord may originate from melting of the deepest and oldest part of the ice sheet. These results raise the possibility of a long subglacial river system that is poorly realized in current ice sheet simulations. If this potential subglacial river system has formed and/or is maintained due to the presence of the ice sheet, then it is a fundamentally different system that requires a different understanding to that of a paleo-fluvial river valley. Questions remain over how much a complete depiction of this feature could affect ice sheet predictions under climate change scenarios.

Appendix A: Error estimates

A map of error estimates from BedMachine v3 (Morlighem et al., 2017) shows the variation in error across north Greenland (Figure A1). Errors range from 2 to 600 metres with a median of 158 metres along the valley. Bed elevation is improved in the lower part of the Petermann catchment (<250 km in our profile in Figure A2g) as it is derived from mass conservation and from a dense IceBridge campaign (see Figure 1b for an outline of the mass conservation region). The kriging interpolation is applied to the rest of the interior of the ice sheet and thus most of the valley. The kriging algorithm is described in Morlighem et al. (2017) as "The variogram is modeled as a Gaussian function, with a sill of 100 m, a range of 8 km and a nugget effect of 50 m, to account for uncertainty in ice thickness measurements". The along valley profile (Figure A2g) indicates that our introduced valley is deeper in the interior than in the BedMachine data. The cross-sections along 3 flight lines across the valley (Figure A2a, b, and c) indicate the valley sides have similar slope angles on the 5 km grid to the observed. 3 more example cross-sections (Figure A2d, e, and f) in regions of high BedMachine error (away from flight lines) show the consequent failure to resolve the valley.

Appendix B: Sensitivity tests

The results from four additional simulations are presented here that test the sensitivity of the water routing to the valley base topographic elevation. 3 of the tests uses 26 linear 10 km wide idealized valleys to form an uninterrupted valley from Interior to Petermann. The valleys are created using the Matlab function "inpolygon" that sets grid point values within an along-valley rectangle to be a specified value. The lithosphere is then relaxed in a short SICOPOLIS simulation to produce an isostatically relaxed bed topography. Tests are done with inserted idealized valleys at constant maximum depths of -100 m, -300 m, and -500 m and are compared to a 4th test which uses standard SICOPOLIS topography as a control simulation. The -100 m simulation effectively removes most of the segmented valley while the -500 m case best represents the slopes of the sides of the valley.

The tests are 100 year long simulations that are initialized with the 1990 SICOPOLIS result and run with constant 1990 atmospheric forcing. The subglacial water flux for the four cases at the end of the simulation (year 100) is in Figure A3. There are large differences in water routing in and around the valley, between these cases. For a -100 m valley (Figure A3f), the northward water flux signature associated with the valley is largely eliminated. If the valley base is lowered to -300 m (Figure A3g), increased valley water flux occurs from Interior to NEEM zone where water appears to be entirely evacuated into a plume directly towards Petermann. For the -500 m case (Figure A3h) the valley water flux is continuous high from Interior to Petermann and the plume out of the valley at NEEM zone is largely eliminated. The results confirm the finding that NEEM zone is the region most prone to water leakage from the valley. The result for a valley base at -500 m suggest that the valley side slopes on the 5 km grid in this case are steep enough to overcome the northwestward directed hydropotential component due to the ice surface slope. From a modelling perspective the results highlight the need to improve the bedrock topography data in the NEEM zone region.

Author contributions. Chris Chambers initiated the study. Chris Chambers and Ralf Greve set up and carried out the numerical experiments with the SICOPOLIS model. Bas Altena performed the primary subglacial topography alteration operation, while the topography for the sensitivity tests was processed by Chris Chambers and Ralf Greve. Pierre-Marie Lefeuvre prepared data, produced the topographic cross-sections, and analyzed the radar flight lines. Chris Chambers interpreted the results and wrote the manuscript with contributions from all co-authors.

Code and data availability. SICOPOLIS is available as free and open-source software at www.sicopolis.net.

Competing interests. The authors declare no competing interests.

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Figures

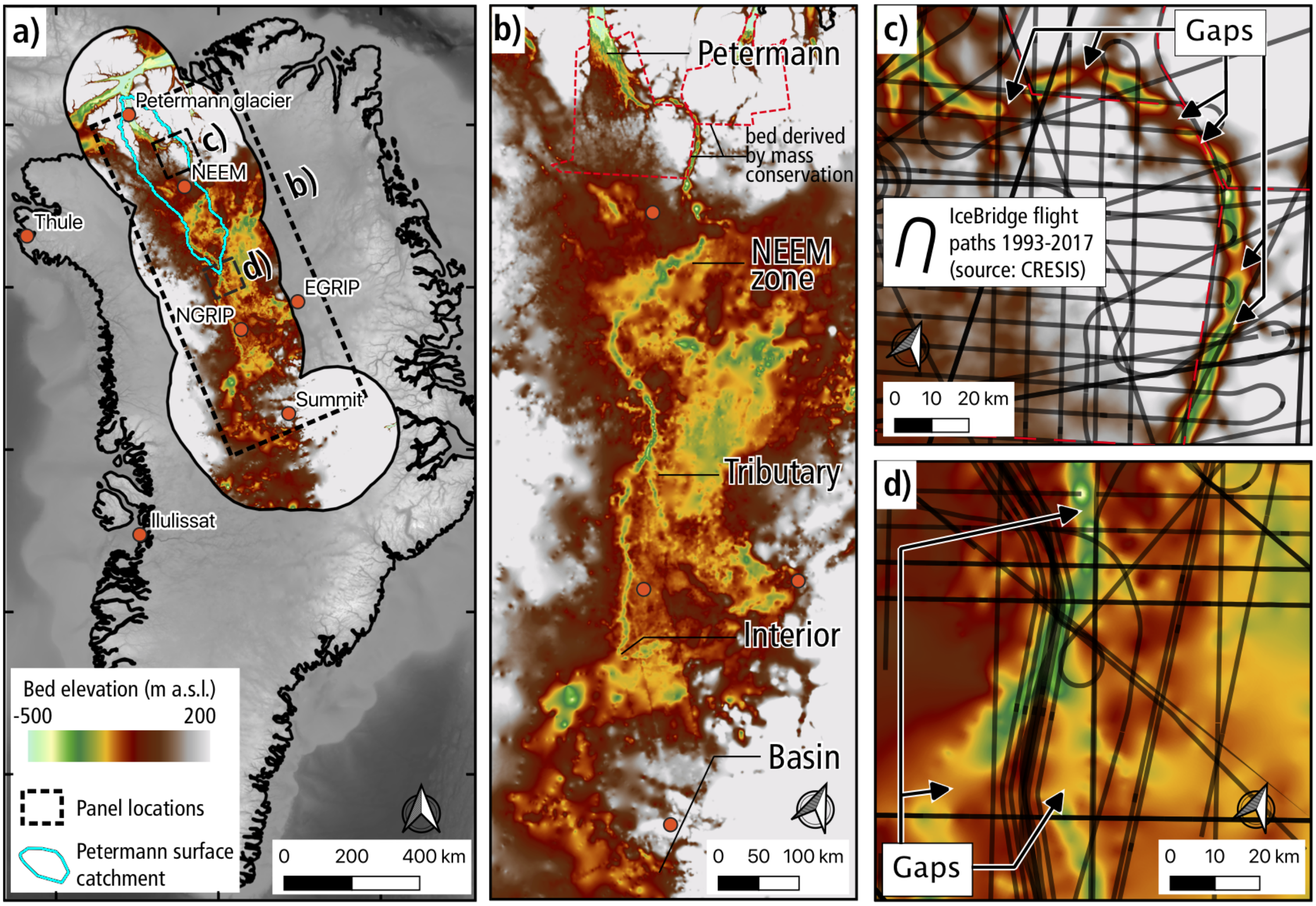

Figure 1. BedMachine v3 bed topography (Morlighem et al., 2017) between −500 to 200 metres above sea level for a) Greenland overview with boxes for b) the valley region and c) and d), two regions showing the IceBridge flight paths.

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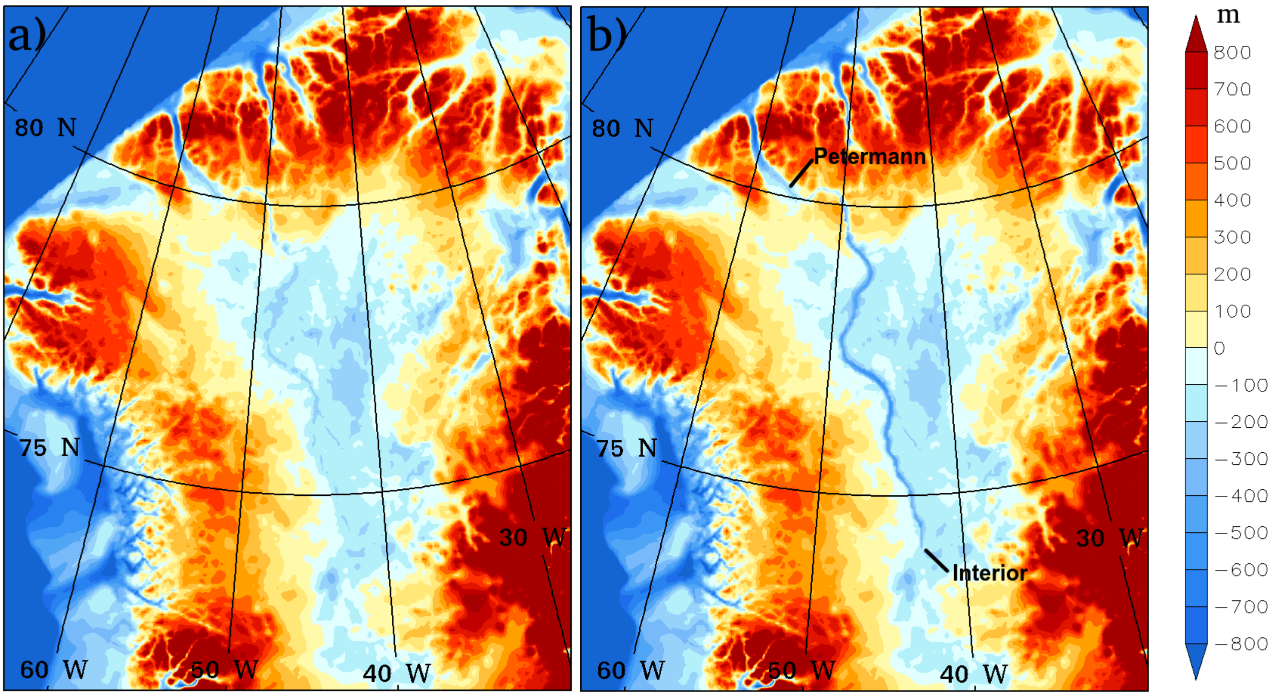

Figure 2. Basal topographic height between −800 and 800 metres above sea level for a) Control (standard SICOPOLIS input derived from BedMachine), and b) Valley (manually adjusted from Petermann to Interior).

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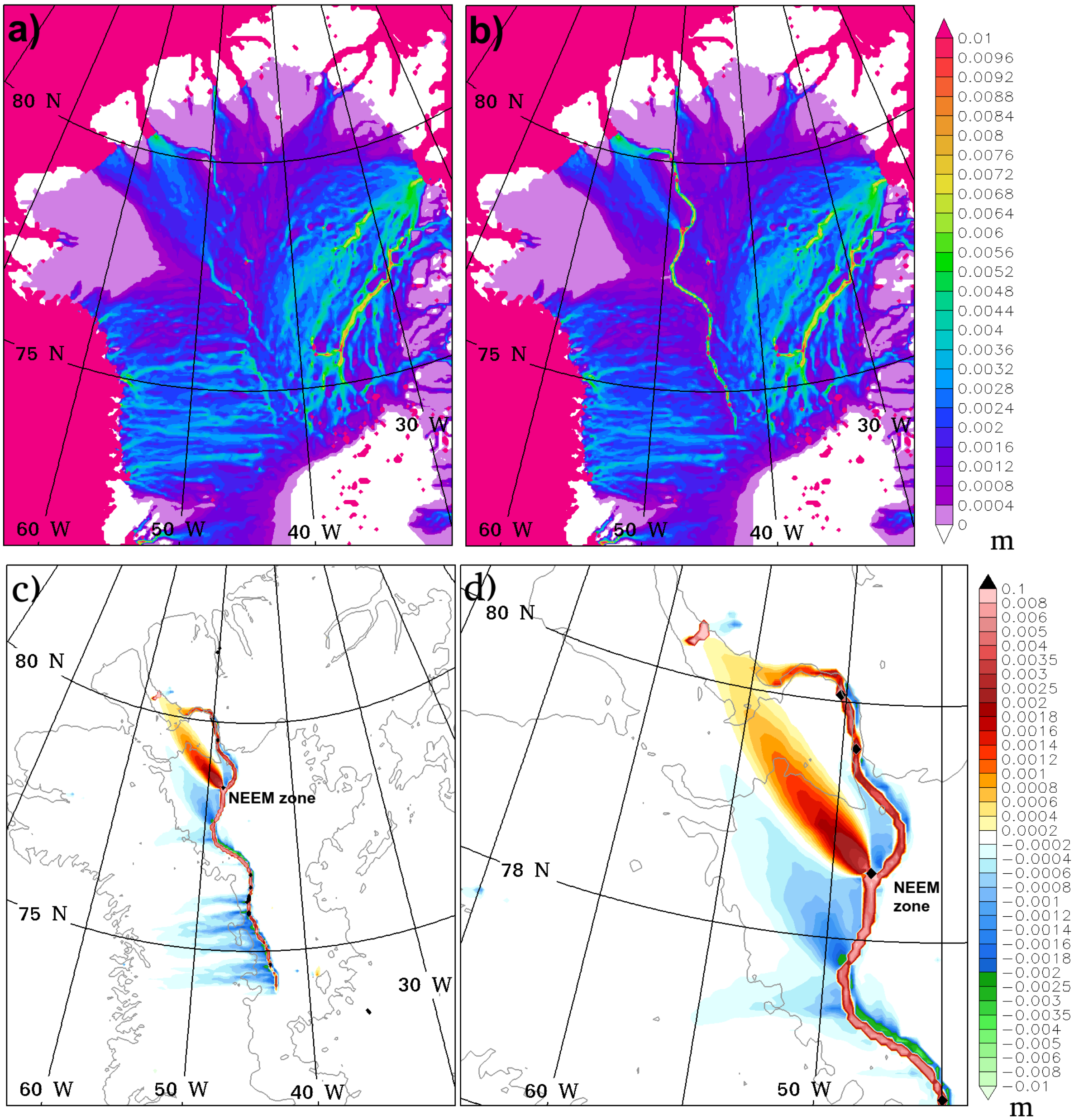

Figure 3. Basal water depth (m) for a) Control and b) Valley, from SICOPOLIS simulations for the year 1990. Basal water depth difference (m) (Valley – Control) for c) northwest Greenland and d) Petermann catchment region.

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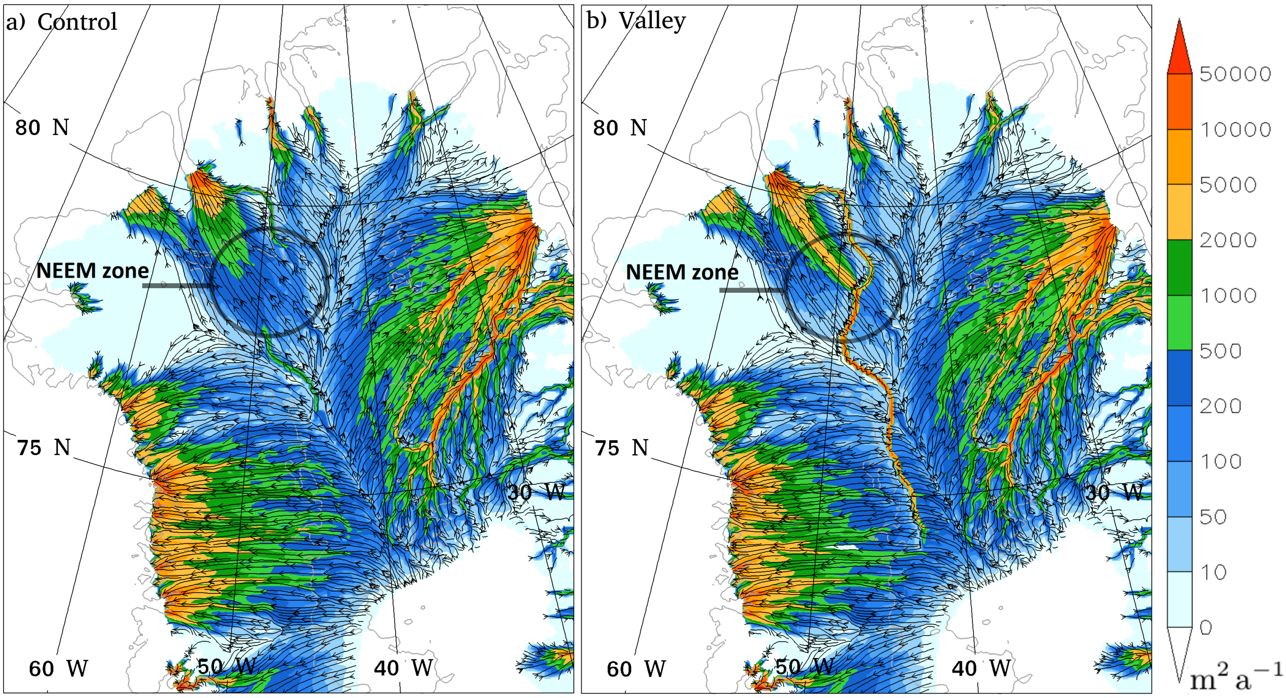

Figure 4. Basal water flux magnitude ($\mathrm{m^2\,a^{-1}}$ colours) and streamlines for north Greenland for a) Control and b) Valley. NEEM zone marks where the greatest change occurs out of the valley as discussed in the text.

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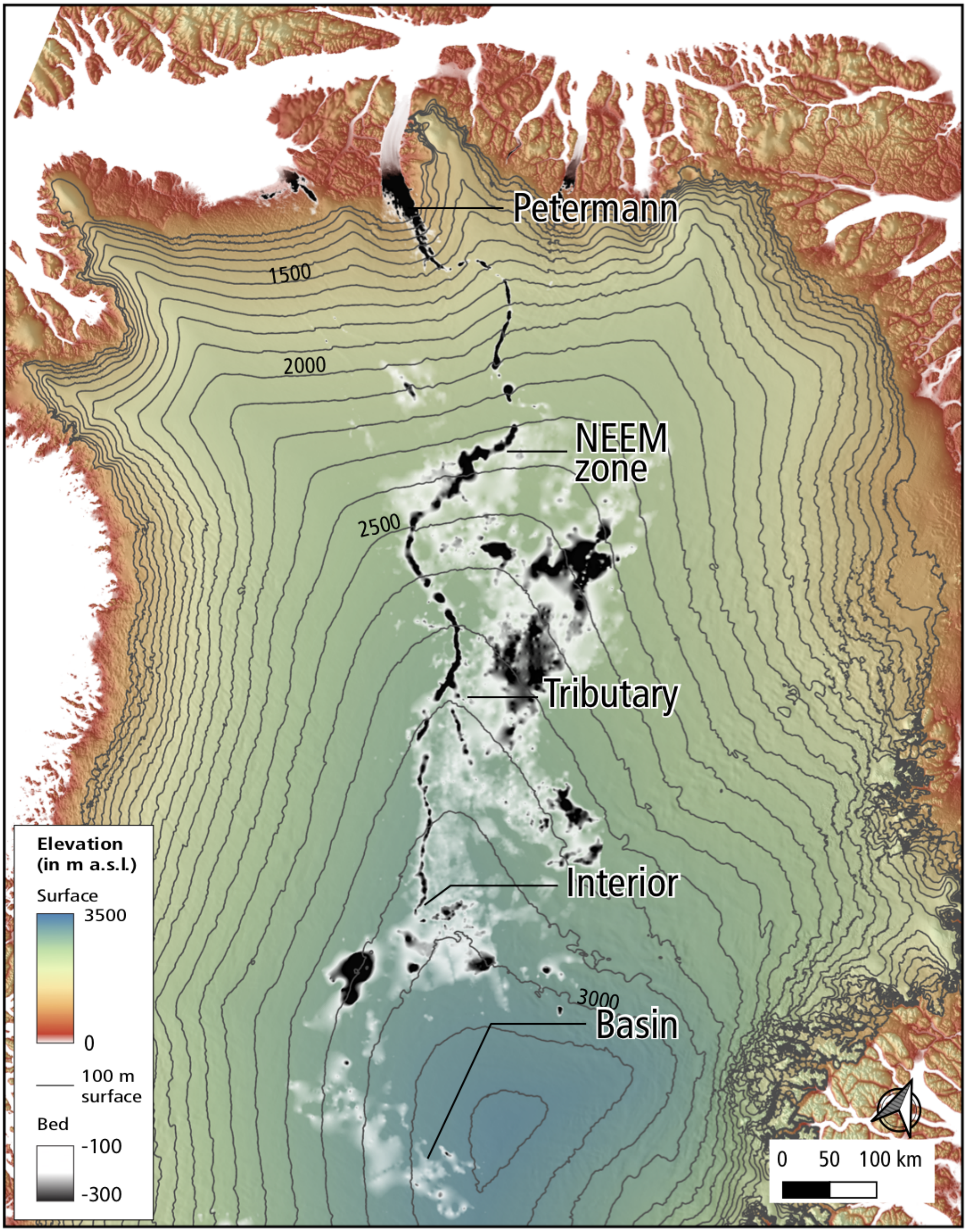

Figure 5. Surface elevation (m) with bed elevation for −100 m or lower overlayed in grey to indicate the path of the valley.

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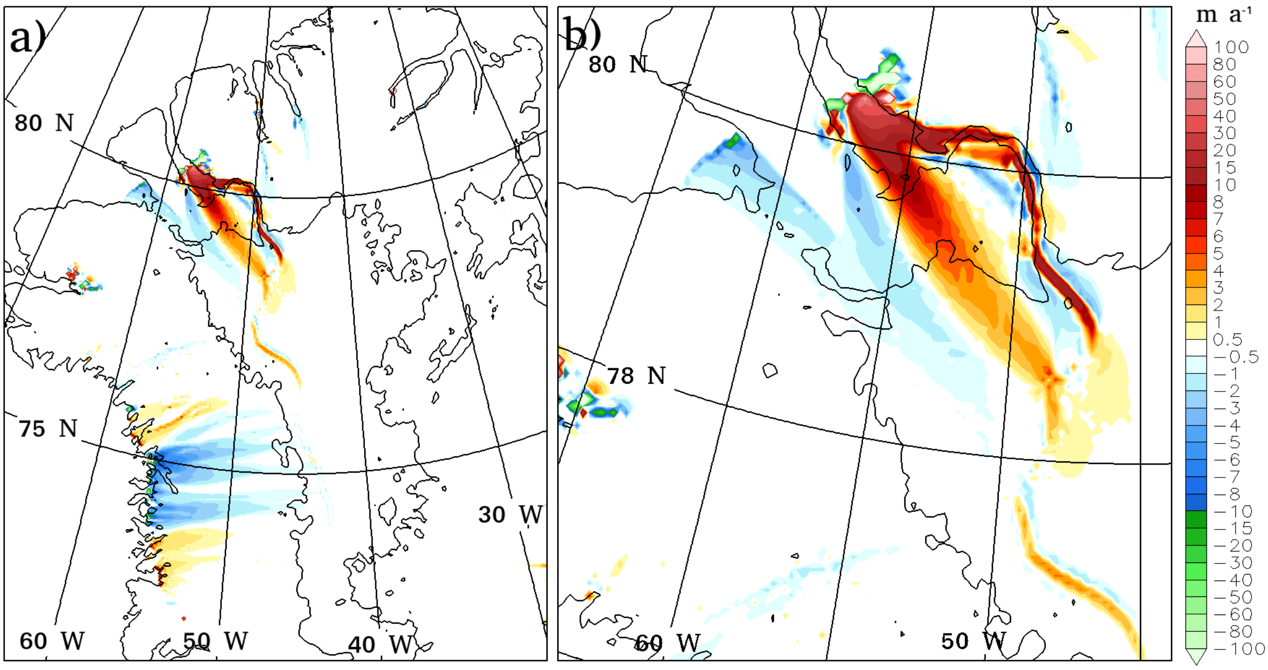

Figure 6. Surface ice velocity difference (Valley – Control) in metres per year for a) north Greenland and b) the Petermann catchment.

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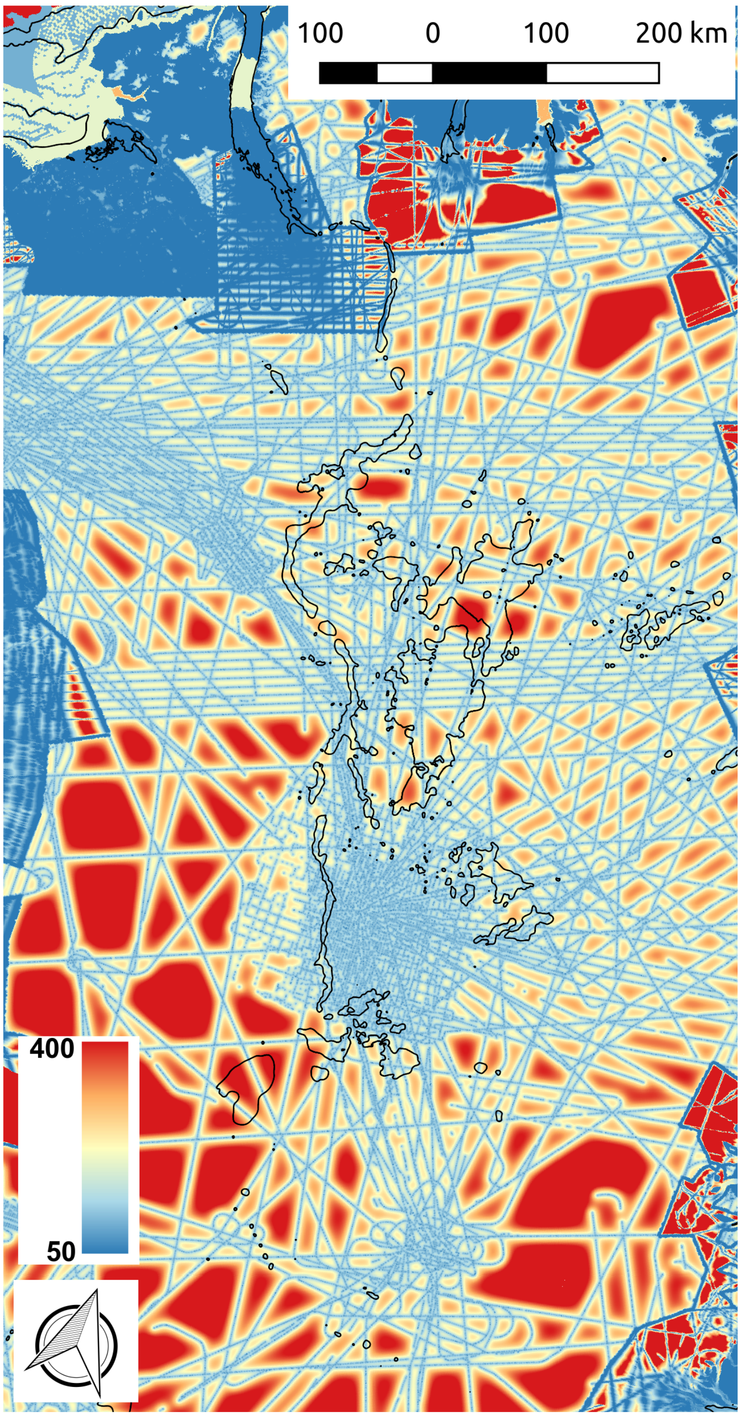

Figure A1. BedMachine basal topography error in metres for the region from Petermann to Basin. The black contour indicates -200 metre elevation.

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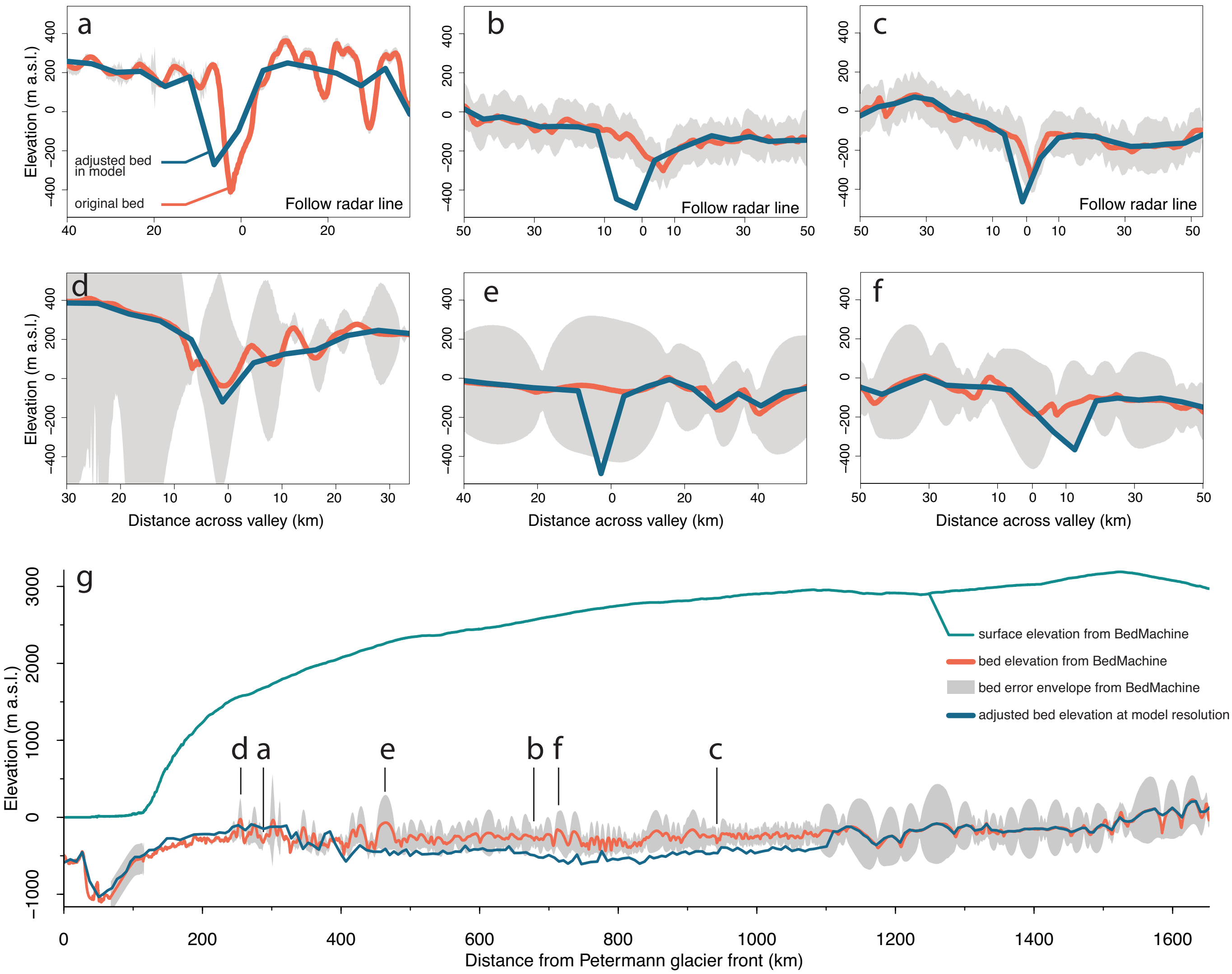

Figure A2. Across (a-f) and along (g) valley profiles from BedMachine v3 bed topography (Morlighem et al., 2017) and the adjusted bed elevation used in our model. The error envelope is derived from error estimates provided in BedMachine. Reduction in error depends on the proximity to radar data as shown in lines a-c that are parallel to flight lines or the use of mass conservation to derive bed topography which covers the region between 130 and 250 km in g).

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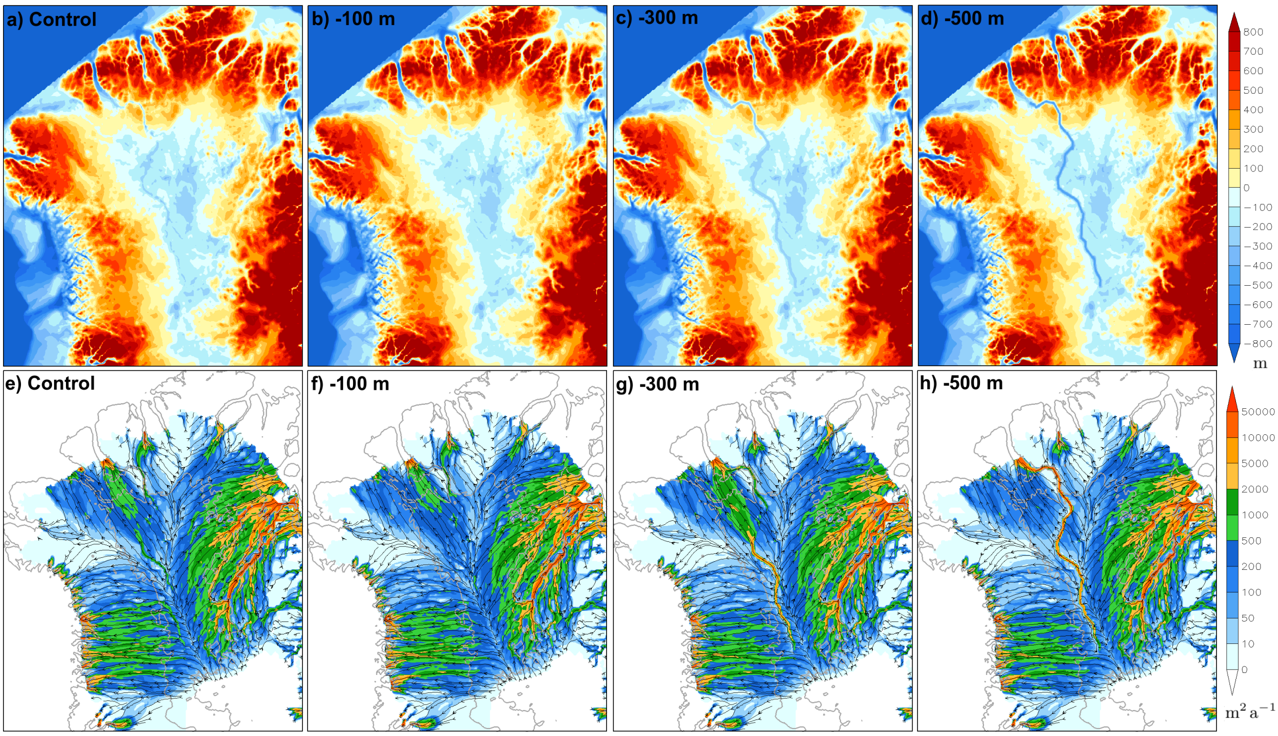

Figure A3. Sensitivity to valley depth tests. Bed topography (m) for a) Control, and for fixed valley base elevations (relative to sea level) of b) -100 m, c) -300 m, and d) -500 m. Basal water flux magnitude (m^2 a^{-1} colours) and streamlines for north Greenland for e) Control, f) -100 m, g) -300 m, and h) - 500 m.