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Glacier geoengineering to address sea-level rise: A geotechnical approach

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Abstract

It is remarkable that the high-end sea level rise threat over the next few hundred years comes almost entirely from only a handful of ice streams and large glaciers. These occupy a few percent of ice sheets' coastline. Accordingly, spatially limited interventions at source may provide globally-equitable mitigation from rising seas. Ice streams control draining of ice sheets; glacier retreat or acceleration serves to greatly increase potential sea level rise. While various climatic geoengineering approaches have been considered, serious consideration of geotechnical approaches has been limited − particularly regarding glaciers. This study summarises novel and extant geotechnical techniques for glacier restraint, identifying candidates for further research. These include draining or freezing the bed; altering surface albedo; creating obstacles: retaining snow; stiffening shear margins with ice; blocking warm sea water entry; thickening ice shelves (increasing buttressing, and strengthening fractured shelves against disintegration); as well as using regional climate engineering or local cloud seeding to cool the glacier or add snow. Not all of these ideas are judged reasonable or feasible, and even fewer are likely to be found to be advisable after further consideration. By describing and evaluating the potential and risks of a large menu of responses − even apparently hopeless ones − we can increase the chances of finding one that works in times of need.

Keywords: Climate intervention; Targeted geoengineering; Antarctica; Greenland; Glaciers; Sea level rise

1. Introduction

Glaciers are slow moving, river-like bodies of ice. The continental ice sheets of Greenland and Antarctica are drained through huge, relatively fast-flowing glaciers or ice streams (Rignot et al., 2011; Joughin et al., 2010), flowing at rates of 1−5 km per year; while the majority of the ice sheets move around 10 m per year. Just five dynamically thinning glaciers in Greenland, < 1% by area, contributed >12% of recent net ice loss (McMillan et al., 2016).

The flow of ice streams is largely through sliding over the bed, and their rate of slip is determined by basal drag which is often locally concentrated in 'sticky spots' (Alley, 1993), related to bed roughness and geometry − either directly as a bedrock bump, or by affecting subglacial water and till. Ice streams may change dramatically: for example, Kamb ice stream on Antarctica's Siple coast became stagnant during the last few centuries (Retzlaff and Bentley, 1993). The driving mechanism is bed water supply, varying because of water piracy between adjacent ice streams or basal freeze-on

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(Tulaczyk et al., 2000; Anandakrishnan et al., 1997; Bougamont et al., 2011).

Climate change affects ice sheets: warmer air can hold more water, increasing snowfall. Warmer ice is softer, and would accelerate – although warming the ice at depth would take millennia. More importantly, warming oceans melt ice rapidly where it starts to float. Where bedrock deepens away from the coast, acceleration because of warming will be dramatic, as warm seawater advances under the ice – ice flow increases due to a deepening grounding line (Alley et al., 2015; Joughin et al., 2012; Turner et al., 2017). Of global sea level rise, 3.4 m is held in unstable marine basins in West Antarctica, 19 m is in basins in East Antarctica that are potentially unstable (Fretwell et al., 2013), and only 60 cm is outside the Greenland and Antarctic ice sheets, in ~200,000 mountain glaciers and ice caps (Grinsted, 2013). Ice stream acceleration and ice sheet collapse poses a major risk to coastal cities (Jevrejeva et al., 2016; Hinkel et al., 2014).

Geoengineering, the deliberate modification of the climate system, has been proposed to address anthropogenic global warming (AGW), possibly complementing a full range of strategies to keep temperatures below thresholds, e.g. 1.5 °C above pre-industrial (Jones et al., 2018). Commonly discussed forms of geoengineering are climatic; seeking to address impacts (like sea-level rise) by controlling global average surface temperature (MacMartin and Kravitz, 2019). These approaches are controversial (Sugiyama et al., 2020), partly due to concerns that changes will be contested (Buck, 2018), causing political strife. Even if carbon emissions fall (per the Paris Agreement – Kitous and Keramidas, 2015), approaching zero sometime after 2050, the oceans will likely continue rising anyway, and the risk remains of dramatic centennial sea level rise (Jevrejeva et al., 2016).

Consequently, this study considers potential geotechnical geoengineering solutions. These do not act on climate; instead aiming to reduce glacier flow rates, restrain the ice sheets which they drain, and/or reduce ablation. The limited literature on ice sheet conservation focuses on blocking warm ocean waters accessing ice shelf cavities (Moore et al., 2018; Wolovick and Moore 2018; Hunt and Byers, 2019) and increasing snow fall (Frieler et al., 2016; Feldmann et al., 2019). Here, we expand the options available to address sea level rise, outlining new potential interventions. We are not endorsing or recommending any of these techniques; the best of them come with serious unresolved scientific and engineering challenges, while the worst of them may have fatal flaws which preclude further development. We are merely expanding the overall menu of societal options and briefly outlining the major issues involved with each one.

We provide a background to ice sheets and sea level rise, with a focus on ice/ocean interactions. For an overview on solar radiation management geoengineering's effect on ice sheets, see analyses of Greenland (Moore et al., 2019), and Antarctica (McCusker et al., 2015), though see caveats in Irvine et al. (2018). Subsequently, we consider intervention possibilities, difficulties and potential. Finally, we discuss risk and precaution in real-world applications.

2. An elementary primer on ice sheets and sea level

Glaciers have an accumulation zone at high altitude, and (in warmer climes) an ablation zone at low altitude. Glaciers are characterised by termination (calving or melting) and base condition (warm – i.e. wet; or cold – i.e. dry). Calving terminations are hard to deal with numerically (Moore et al., 2013; Åström et al., 2014). A bed lubricated with water may be due to frictional melting of the glacier moving on the bed, geothermal heat, or surface infiltration. Dry-based glaciers flow slowly, through internal deformation of the ice only (Moore et al., 2013). Dry glaciers react very slowly to changing climate conditions – on timescales, of millennia (Cuffey and Paterson, 2010). Glaciers terminating on land are relatively slow-responding – because the heat capacity of air is relatively low; melt rates are typically ~1 m per year. Glaciers terminating in water can have basal melt rates of order 100 m per year as the ocean transfers heat more efficiently (Turner et al., 2017). Fast-flow is controlled by sliding across their beds and is governed by the availability, pressure, and distribution of subglacial water.

Subglacial water systems can be classified as either efficient or inefficient. Efficient systems move large volumes of water through conduits governed by equilibrium between outward melting and inward creep closure (Rothlisberger, 1972; Weertman, 1972; Nye, 1976; Rothlisberger and Lang, 1987; Hewitt, 2013). Efficient systems cover a small fraction of the basal area, running at lower pressure than the ice overburden, increasing basal drag and decreasing ice sliding. Efficient systems dominate where surface melt is high – summer, Greenland ablation zone & downstream Antarctica (Schroeder et al., 2013).

Inefficient hydrology systems involve either poorly-connected subglacial water pockets, or water in saturated subglacial sediments (Blankenship et al., 1986; Walder, 1986; Kamb, 1987; Tulaczyk et al., 2000a, 2000b; Hewitt, 2013). In these systems water covers a large fraction of the basal area and is close to the ice pressure – reducing drag and promoting rapid sliding. In both efficient and inefficient systems, water exits at the ice margin or grounding line.

Runoff from the surface and basal melt emerges at the grounding line as cold and fresh subglacial discharge, which is buoyant. This rises along the ice face as a buoyant plume (Jenkins, 1991, 2011). Turbulent mixing entrains warm salty ocean water. This reduces plume buoyancy but increases heat content, which controls the melt rate. Increased subglacial discharge and ocean heat content increase melting at the ice–ocean interface, while also pushing calved tabular icebergs away from fragmenting ice shelf ocean margins.

Marine terminating glaciers are held by buttressing. Bedrock high-points contact the floating ice base, acting as pinning points. Greenland fjords walls provide buttressing without pinning beneath. Sea ice and icebergs forming a melange reduce calving rates (Amundson et al., 2010; Robel

2017), but do not substantially restrain ice flux except in narrow fjords (Guo et al., 2019; Pollard et al., 2018).

Glaciers terminating in the ocean and resting on bedrock that deepens inland are subject to Marine Ice Sheet Instability (MISI), a positive feedback between ice flux and grounding line depth (Hughes, 1973; Weertman, 1974; Thomas and Bentley, 1978; Schoof, 2007). As the grounding line retreats into a deepening basin, ice flux across the grounding line increases, thus increasing the rate of dynamic thinning and accelerating retreat. The MISI is mechanical, driven by bedrock geometry; but ocean coupling will accelerate instability. The retreating grounding line and thinning shelf will increase sub-shelf cavity volume, allowing more vigorous circulation and increased basal melt (Jacobs et al., 2011).

Greenland surface melting leads to about $^1/_2$ present mass loss, the rest being by calving (van den Broeke et al., 2009). Water runs through the ice mass in a sub-glacial drainage system that is well-organised, channelized and efficiently drains to the ocean. Efficient drainage means surface melt increases do not accelerate ice flow. Nevertheless, meltwater induces further melting, via albedo. In Antarctica, essentially no surface melt is supplied to the bed, which is lubricated by water formed *in situ* and well-dispersed across the ice stream. Reducing production of water at the bed will have the effect of reducing lubrication and slowing the ice stream.

Modifying either ice streams' basal conditions, or increasing buttressing will tend to restrain flow. Increasing snowfall on ice sheets, or otherwise adjusting surface mass balance will accumulate ice, but will thicken and steepen the ice, leading to increasing flow within a few centuries, thus acting against accumulation (Winkelmann et al., 2012).

The most important factor controlling sea level rise uncertainty (Jevrejeva et al., 2016) are glaciers and ice streams susceptible to MISI. MISI may already be underway on Pine Island Glacier, Thwaites Glacier, and their smaller neighbors (Joughin et al., 2014; Favier et al., 2014; Rignot et al., 2014). Others might become important in the multi-centennial timescale, but only a handful are of immediate concern (DeConto and Pollard, 2016); these are located in the Amundsen Sea sector of Antarctica. There is also much to be gained by observing and understanding relatively smaller systems, e.g. in Greenland (Fig. 1).

3. Interventions

3.1. Albedo management

Attempts have already been made to preserve glaciers and sea ice by using insulating or reflecting materials (Field et al., 2018; Messenger, 2010). These may be useful on glaciers draining ice sheets. Where warm air contacts the glacier surface in the ablation zone these may prevent surface darkening, from dust or melt ponds.

In the Greenland ablation zone, drainage (via surface drainage channels and then moulins) causes surface accumulation of dark material (dust and black carbon). Ice covered by thick debris (mountain glaciers), is insulated from solar heating and warmed air, and melt rates may be very low. Polar debris cover is much thinner and tends to accelerate melting, by lowering albedo.

Surface melt ponds are darker than surrounding ice and snow, absorbing much more energy from the sun. Draining ponds into moulins would negate this. Lake drainage would need to be gradual, as sudden lake drainage overwhelms the basal hydrology system and causes a surge in ice flow (Das et al., 2008). Regulating drainage rates at many separate lakes and monitoring basal hydrologic response would present major difficulties in implementing this intervention.

Alternatively, pond water may be used for snow making, when air temperature is low enough. Unlike earlier ideas that intended snow making to add bulk mass to the ice sheet (Feldmann et al., 2019), our goal is to adjust albedo. Only thin layers need be applied; water and energy requirements are far lower than when adding mass to compensate for sea level rise. These machines are efficient in air temperatures $< \sim -5$ °C.

Albedo modification has been investigated for Arctic sea ice. The 'Ice 911' proposal considers addition of reflective, sand-like material (Field et al., 2018). This may be beneficial, outperforming artificial snow in warmer conditions. Achieving the desired mono-layer is challenging even for flat sea ice, where icebreaker ships may be positioned rather arbitrarily. Glaciers are difficult and dangerous environments for machinery or pedestrians, due to crevasses, severe weather conditions, difficulties of search & rescue, and remoteness. Aircraft spreading has some similarities with crop-spraying where economic coverage is the primary concern, but is likely to be prohibitively costly when extended over vast areas of Greenland.

Instead of covering dark material, a conceivable alternative is removal. Polluted snow is darker; this is amplified during the melt season, as dark particles concentrate on the surface. Specialist machines (perhaps based on ploughs, bulldozers or dragline excavators) might be developed to scrape the dirty surface layer of relatively stable glaciers into piled windrows, powered by smokeless fuel such as hydrogen to avoid black carbon emissions exacerbating the problem. This could usefully create packed snow berms as windbreaks.

Where a mountain glacier is thinly covered in rubble, albedo may dominate thermal insulation. Removing rubble may be a locally-effective intervention – but such alpine glaciers contribute little to sea level rise. Nevertheless, this may provide a local ecosystem service.

The dark region of Greenland is extensive. There is no leverage from working only the fast outlet glaciers, because surface melt does not affect their flow. Any surface albedo modifications would be extensive, and costly. Dark material is confined to the surface decimetres, but this is rough, with surface crevassing, sastrugi (wind-formed features), and melt ponds making complex topography. Removal of the dirty top layer would expose the next layer to summer melting and renewed surface concentration of dark contaminants in that layer. Continued deposition of airborne black carbon aerosols from industrial pollution and potentially-increasing boreal

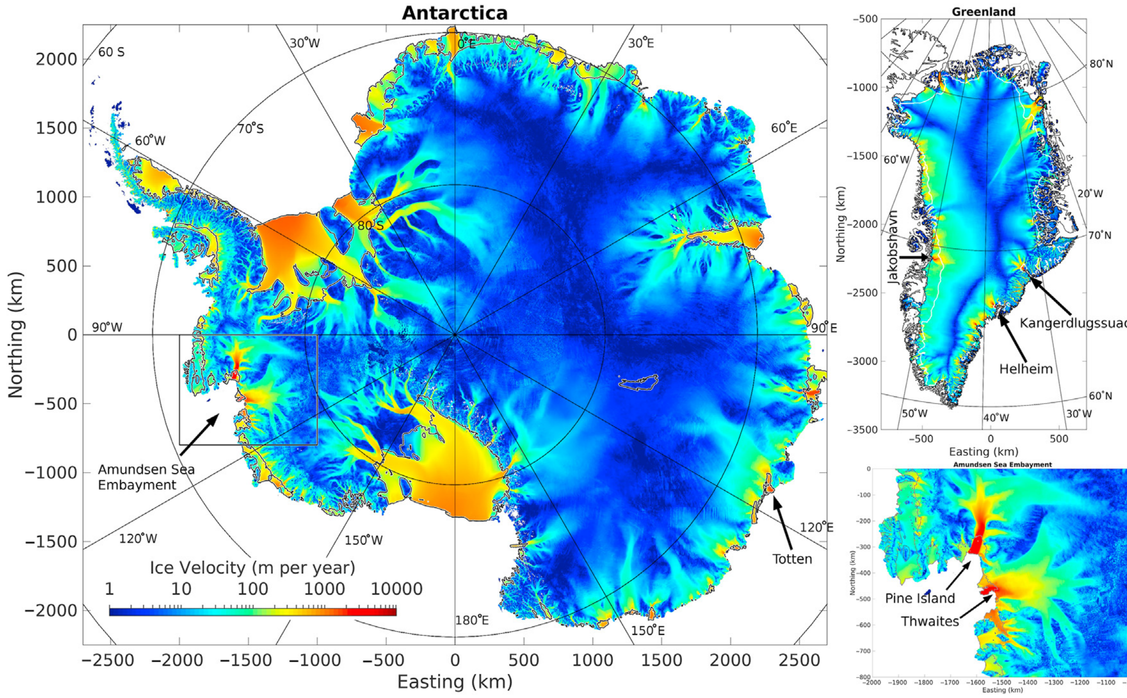

Fig. 1. The Antarctic and Greenland ice sheets showing their surface ice velocities, and an enlargement of the Amundsen Sea embayment with the Pine Island Thwaites glaciers, that are rapidly losing mass to the ocean. Particular outlet glaciers are marked that might present opportunities for engineering, for local community benefits, engineering testing, or large sea level rise commitments.

forest wildfires will continue the accumulation of dark contaminants.

In summary, such albedo modification techniques do not appear practical at the largest scale. Deployment would be required over a large fraction of the ablation zone, in order to have a meaningful influence on sea level rise. Operations on smaller, rubble-covered mountain glaciers may help local preservation, but the energy and cost requirements of dealing with the entire ablation zone are prohibitive. Albedo modification techniques may find a niche in areas around the equilibrium line where snow loss and albedo reduction were just beginning, but these techniques do not seem to be feasible at large scales.

3.2. Anchoring techniques

The gravitational driving forces in large glaciers and ice streams are much too large for ice sheet movement to be more than infinitesimally slowed by mechanical anchoring or stiffening with steel cables or chains, by bedrock blasting to roughen the glacier bed, or by somehow installing artificial protrusions, such as stub piles embedded in subglacial till or drilled into bedrock.

Some degree of ice shelf stabilization might be possible, with the objectives of maintaining buttressing of inland glaciers, preventing marine ice-cliff instability (DeConto and Pollard, 2016), and reducing the draft of released icebergs that could endanger any artificial subsea barriers installed to block deep warm seawater intrusion. The horizontal resistive stress due to buoyancy of a free-floating ice shelf 500 m thick is about 250 kPa, much too large for any reasonable tensile reinforcement of the shelf surface to exert more than minor inhibition of crevasse fracturing and rift propagation. However, tensile reinforcement by anchored steel cables or buried chains could delay final rupture of tabular iceberg rifts by mooring emergent icebergs in place. Large drifting icebergs cannot plausibly be restrained, but it could be possible to prevent them from initial drifting until much later in the process of ice shelf thinning and decay.

Iceberg drift away from a shelf margin is driven by wind and ocean currents. These include outward-flowing buoyant plumes of subglacial meltwater: these are often also associated with polynyas, which form as surface water and floating ice is pushed away from the shelf margin. These forces might be restrained by cable networks with self-embedding plough devices for load transfer to the ice, and post-tensioning by ice shelf spreading and weighted cable hanging over the active shelf edge. With icebergs moored to the parent shelf, rifts will tend to heal by winter freezing. This has taken place naturally in the case of the Brunt ice shelf, in which a flotilla of mutually jammed icebergs has been frozen together in a reconstituted ice shelf. In suitable weather conditions, pumping water onto the surface may speed this fusion.

Deployment of tensioning cables across rough and treacherous ice shelf surfaces would be difficult. Cables hanging over the shelf edge will eventually drop to the seabed, from which they might later be salvaged for recycling or reuse. Investigation of stresses, fracture modes and stability management of the new calving front would require much further research. Any such intervention would certainly be extremely expensive in terms of material, installation and continuing operational costs.

Splitting of floating shelves into icebergs is not fully understood. Fatigue by flexing from tides and waves (Pritchard and Vaughan 2007), and damage from pre-existing fractures contributes to rapid shelf disintegration (Liu et al., 2015), especially when thinned by melting from below. It is aided by freeze-thaw hydrofracturing by surface meltwater (Scambos et al., 2003), the draining of which could be a useful intervention − again, depending on weather conditions.

Tensile reinforcements to suppress the breakup (e.g. as described above), would ideally have a designed preload to create compression across the potential split, while withstanding loads from wind and ocean currents, as well as flexing from tides and low frequency ocean swell. It is challenging to locate future fractures, and to put sufficient reinforcements in place.

Ice streams are often flanked by narrow and highly fractured shear margins, across which velocities increase from near-zero to full speed. As ice streams accelerate, shear margins become more chaotic, damaged, and provides less buttressing (Alley et al., 2019; Guo et al., 2019).

Strengthening shear margins might add buttressing. Forces on grounded ice flows are much too large to control with steel − but, particularly where streams float (forming ice shelves), it might be feasible to stiffen shear margins with ice. Stream margins are strengthened naturally in cases where they merge into shelves and conditions allow accumulation of basal ‘marine ice’ underneath them (Holland et al., 2009). Marine ice infills basal crevasses with saline ice which is more ductile than freshwater ice, and strengthens suture zones between streams (Kulessa et al., 2014). Infilling floating shear margins might work, naturally penetrating the crevassed zone and freezing rapidly in cold ice. This speed may be attractive, although slower freezing and fresher marine ice accumulation might be important for mechanical strengthening, since the brine trapped affects its rheology (Khazendar et al., 2009; Moore et al., 1994).

3.3. Bed drying

Warm-based glaciers tend to flow fast because they have liquid water lubricating their beds. That bed may be hard rock, but more likely is soft sedimentary material.

The Kamb ice stream in Antarctica switched off (Retzlaff and Bentley, 1993) when its water supply was lost − either to the adjacent ice stream or to freezing (Alley et al., 1994; Tulaczyk et al., 2000; Anandakrishnan et al., 1997; Bougamont et al., 2011). Hence natural basal drying is known to slow ice streams.

Bed water may be organized in various ways: from dispersed pockets, to dendritic channels. The more connected the drainage system, the less the water introduced affects sliding; a concentrated, efficient drainage system occupies a small area and has lower pressure than the ice overburden (Rothlisberger, 1972; Nye, 1976; Rothlisberger and Lang, 1987; Hewitt, 2013). Rapid pulses of water into an efficient system (days or less) can overwhelm its capacity, causing pressure to increase and water spreading, temporarily increasing sliding (Das et al., 2008; Schoof, 2010).

In Greenland, basal water volume is large, due to surface melt; this promotes efficient drainage. In Antarctica, water comes from basal melt via plastic deformation, compression melt, geothermal flux, and bed friction (friction is dominant in faster flows). In the upper reaches of Antarctica's fast outlets (e.g. Pine Island and Thwaites, Fig. 2), drainage is inefficient and water dispersed. As ice streams approach the coast, and larger volumes of melt are available, drainage systems become more organized. Radar shows Thwaites Glacier's drainage becomes channelized <50 km from the grounding line (Schroeder et al., 2013). To slow glaciers subject to rapid MISI, reducing water in the bed at elevations above the threshold for channelized flow is likely to be desirable. However, modelling downstream impacts is critical, to ensure that this intervention does not impede channel formation downstream. Once water is channelized there is little impact on glacier speed. The objective would be to either drain distributed systems (i.e. drain subglacial sediment); or convert distributed drainage to channelized, which does not seem possible.

Removing basal water requires challenging engineering. The oil industry offers relevant techniques. Simple drilling (rotary bits, hot water drilling) allows creation of routes from bed to surface, from which water can be extracted. Directional drilling is possible, enabling a branched network of boreholes from a single surface well.

Such extraction may not be simple; the base will typically be close to freezing point, while higher layers are colder. The water temperature must be raised to prevent freezing and borehole blockage. Electrical heat tracing along borehole casings is needed, with enough heating power to compensate unavoidable conductive heat flow into surrounding ice. Heat loss into the surrounding ice will be greatest during initial operation of a drainage well. Filament-wound fiberglass pipe can be used as well casings, with advantages of lower thermal conductivity and much lower weight than steel. A braided insulation sleeve of robust thermoplastic material could somewhat reduce heat tracing power consumption, and could help manage stresses in the casing from deformation of the glacier.

The sediment bed is a sensitive system. Porosity and permeability depend on pore water pressure. When pressurised, porosity and permeability are high, as sediment grains are water-separated. Reduction in pressure collapses water pockets, severely reducing bed permeability (Tulaczyk et al., 2000) especially in tills with high clay content. Further reduction in suction pressure to drive adequate water flux will greatly increase pumping power requirements, while increasing the risk of blockage by ice formation. Removing water from a wide area will be more difficult than pumping it from one hole, unless careful adjustment of pressures and water flow are achieved. It may be impracticable to extract water over ~30 m basal distances in a distributed drainage system. A network of boreholes is probably needed − unless

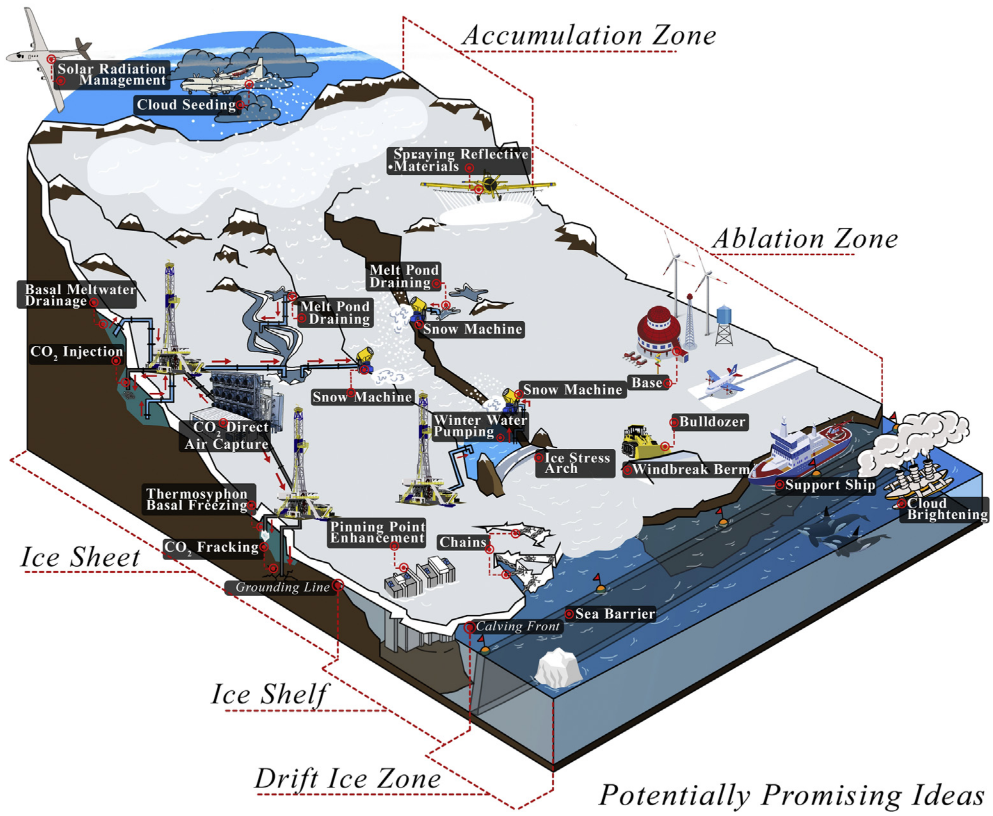

Fig. 2. Schematic representation of glacier intervention engineering schemes. Note that in this cartoon the ice area relative to the interventions is about 1 million times smaller than in reality, and it would be unlikely to utilize more than one method on any particular glacier. Albedo: reflective materials, draining melt ponds, snow making machines; cloud seeding. Bed drying & binding:melt removal; thermosyphon base freezing; enhanced Oil Recovery analogues; CO_2 hydrate formation; CO_2 fracking-chilling. Ice shelf buttressing: enhancing pinning points; thickening/strengthening ice with pumps, snow making machines, thermosyphons and wind breaks; draining shelf melt; tensile reinforcement. Environmental modification: cloud brightening, underwater berms/sheets, manipulating ocean/air currents, regional solar radiation management.

good surveys of basal water pockets can be made (e.g. with ice-penetrating radar, seismic surveys, or down-hole cameras). Optimal borehole locations could be determined by ice dynamics simulations.

Reservoir engineering expertise from the petroleum industry may be helpfully applied to subglacial drainage challenges. Enhanced oil recovery (EOR) techniques based on nitrogen or CO_2 injection into failing wells would be affected by the energy cost of compressing air or nitrogen, or capturing and liquefying atmospheric CO_2. This would greatly exceed the pumping power for lifting the water displaced by that fluid injection. However, such approaches may be required to dry porous beds, where pressures are limited by the packing of loose bed material. Were such techniques used, a delicate balance would need to be maintained: over-pressurization would result in local jacking of the glacier, removing friction; while under-pressurization would cause a drop in porosity, blocking flow towards the borehole.

Water extraction would not need to remove all bed water. Pine Island Glacier produces about 50 m^3 s^{-1} of melt over its whole area, or around 10 mm per year (Joughin et al., 2009); disposal would be difficult. Stress transfer means that increasing basal friction in a small area would slow the glacier. This is especially true in areas with both low driving stress and drag; the addition of localized sticky spots could substantially increase integrated drag. Conservatively assuming that 0.1 m^3 s^{-1} represents a quantity of water that could reasonably be disposed of at the surface, and taking 10 mm per year as an average basal melt rate, that suggests capacity to create roughly 300 km^2 of sticky spots. Freezing this much water at the surface would release 33 MW of latent heat; if we conservatively assume that 1 W m^{-2} represents a reasonable area-averaged heat flow that could be dispersed at the surface, then this would correspond to spreading the water over an area of 33 km^2, or a square just under 6 km on a side. A vastly reduced surface area would be required in the polar night, where temperatures are often tens of Celsius below zero. It is impossible to know how much water could actually be removed without far more detailed design and engineering analysis, but changing the amount of water that could be removed would change the amount of sticky area that could be created (proportionally, under the simplistic assumptions that melt rate is spatially constant, and complete and even drying can be achieved). Coupled ice dynamic and basal hydrology modelling would be required to optimise the distribution of sticky spots across the ice stream. The resulting water could be frozen easily in winter near drilling sites. Spraying water would aid cooling, and it may also need to be settled/filtered, removing dark till that would exacerbate summer melting.

Removing water flowing along the bed could have advantages beyond the bed. Meltwater discharge at the grounding line is many hundred meters below the surface, generating a buoyantly-rising plume, which turbulently entrains warmer seawater to wash against the ice front – causing rapid melting (Jenkins, 1991, 2011). Subsea ablation from mixed meltwater/seawater weakens the glacier from below, encouraging calving fragmentation. If the buoyant plume rises to the surface, it pushes floating ice fragments out to sea, removing any buttressing, which would otherwise prop the glacier before subsequent calving. A large fraction of total meltwater would require removal to achieve this additional benefit.

An organized drainage network is the least effective way for water to affect ice speeds. Only a tiny fraction of the glacier is then in contact with the water (rather than most of the glacier, in a distributed system), and water pressures are lower. There is nothing to be gained by disrupting or draining channelized networks; it is the drainage type that draining the bed (Moore et al., 2018) is seeking to mimic.

While subglacial water drainage could greatly reduce bed lubrication, total energy costs will be very high for drilling wells, heat tracing wells through cold ice, pumping the water to the surface, and excess pressurization for water disposal by winter freezing or snow-making. A dense field of wells over a substantial area would probably be required for stable formation of an ice stream 'sticky spot'. Designing and operating the well field will be challenging, considering hydrologically active terrain on a moving glacier and a dynamic sedimentary bed.

3.4. Basal freezing

Bed freezing equates to drying (Moore et al., 2018). Conceivably, coolant could be circulated through boreholes with winter cooling at the surface, or between the bed and cold layers at intermediate depth. This would create frozen plugs around active ends of cooling wells. There are great technical challenges. Firstly, holes need accurate drilling (near or within basal melt), including installation of impermeable sleeves. Secondly, bottom heat transfer, to take up the latent heat of freezing, is restricted by ice's low thermal conductivity. The heat transfer limitation also applies to dissipation into cold intermediate ice layers, as opposed to air cooling. Thirdly, heat transport by sensible heat coolants is impaired by thermal equalization between counter-current coolant supply and return flows – potentially necessitating separate wells. Fourth, numerous parallel boreholes would be needed to cool any substantial basal area, with costs compounded by limited life of boreholes in moving glaciers.

Heat transport by latent heat of vaporization is more effective than by sensible heat of a liquid coolant. For subglacial freezing, liquid air, liquid nitrogen or liquid CO_2 might be technically feasible direct contact refrigerants. The excessively high energy cost of air or nitrogen liquefaction would limit pragmatically possible application of those cryogens to small scale flash freezing applications. CO_2 captured from the atmosphere might conceivably be a more viable direct contact refrigerant because of the further incentive for carbon sequestration, if applied to inland glacier regions sufficiently far from unstable margins.

The low temperature and low humidity of polar air reduces the theoretical energy consumption of capturing CO_2 from ambient air. Atmospheric CO_2 capture is now being performed at pilot scale in temperate climates by adsorption on special solid adsorbents with high affinity for CO_2, followed by thermal swing regeneration of the adsorbent (Wurzbacher et al., 2012). Low temperatures over polar ice sheets would facilitate use of zeolite or activated carbon adsorbents to concentrate CO_2, followed by cryogenic liquefaction. The CO_2 will be gravitationally pressurized in delivery down the first compartment of a dual compartment well with annular casings as shown in Fig. 3. The first compartment between outer and inner casings would be pressurized above basal pressure, while the second compartment inside the inner casing would contain low pressure CO_2 being returned to the surface. A remotely actuated closed-centre switching valve at the base of the casing could sequentially open each compartment to the basal formation while closing the other compartment. This switching valve would in a first position deliver a pulse of cold liquid CO_2 at well above the lithostatic pressure into the porous subglacial till formation to act as fracking fluid, and in a second position vent flash vaporizing CO_2 back from the formation into the second compartment for mechanical recompression at the surface. The vapour pressure should be less than 2 MPa at the base of the second compartment in order to avoid CO_2 condensation in the inner casing of the vertical pipe. A refrigeration cycle would thus be provided between basal heat uptake from CO_2 vaporization in the basal area surrounding the borehole, and heat release from reliquefaction at the surface.

The switching valve sequentially opens the liquid and vapour compartments to the subglacial till, while never connecting the liquid and vapour compartments together. The switching valve might be a rotary valve whose actuator would be energized by the pressure difference between first and second compartments, and controlled by a solenoid pilot valve responding to basal pressure transducers. The injection/release cycle period would likely be hours or possibly days. Designing this cycle would be a very challenging exercise for collaboration with petroleum reservoir engineers, and requiring experimental work on fracturing frozen ice/till composites. A critical issue will be satisfactory release and return of CO_2 vapour, as a majority fraction of CO_2 liquid injected. Following typical petroleum industry fracking practices, particulate proppants may be injected with the liquid CO_2 to keep flow passages sufficiently open during the low-pressure vapour release step.

A portion of the liquid CO_2 injected into the subglacial till formation would remain there and gradually bind with basal water to form clathrate hydrates. CO_2 hydrates have a higher melting temperature at subglacial pressures than ice, and in

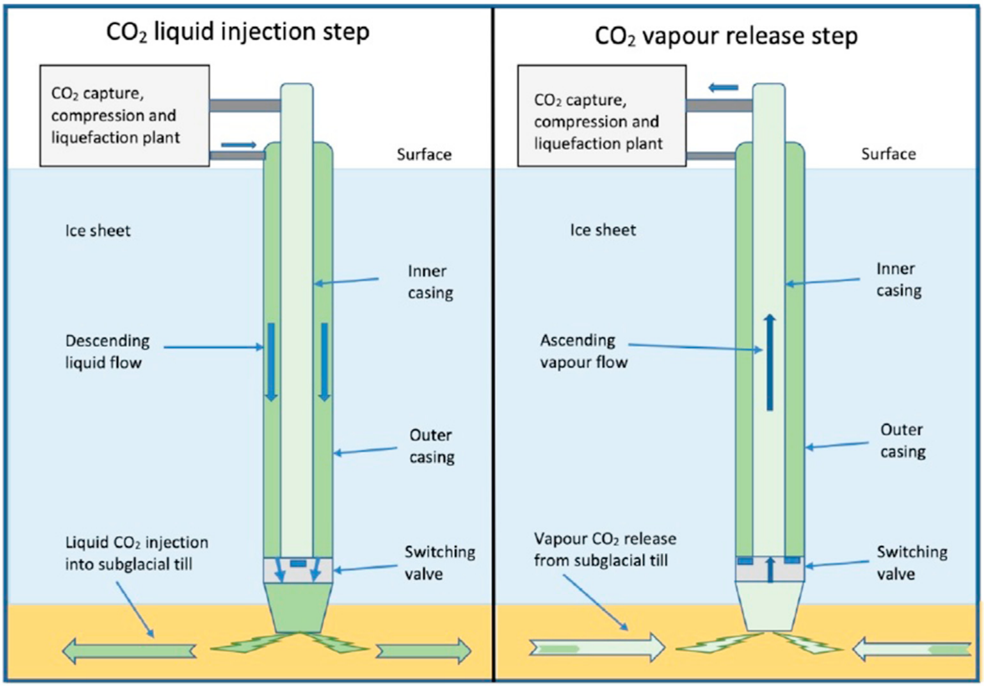

Fig. 3. Direct contact basal refrigeration by cyclic liquid CO_2 injection and vapour release.

massive forms have superior mechanical properties. Methane hydrates may have contributed to ‘sticky spots’ in palaeoglacial ice streams under the present Barents Sea (Winsborrow et al., 2016). Bed drying by hydrate formation sequesters CO_2 from the atmosphere, however exothermicity of hydrate formation would require supplemental basal refrigeration to avoid melting adjacent ice. One mole of liquid CO_2 binding with about 6.5 mol of water ice to form hydrate at the basal freezing temperature would release 12.8 kJ mol^{-1}(CO_2), enough heat to melt another 2.1 mol of ice (Anderson, 2003). As the latent heat of CO_2 vaporization at the freezing point of water is 10.3 kJ mol^{-1}, the majority of the descending liquid CO_2 must be returned to the surface as vapour in order for any net basal refrigeration to be achieved after any remaining CO_2 has converted to hydrate. The fraction of the injected CO_2 remaining in the till formation must be small to avoid compromising the basal refrigeration purpose by exothermic hydrate formation. Nevertheless, the figures above still show a net increase in solid material in the wet system, even without heat removal.

This direct contact CO_2 cyclical injection/release refrigeration concept should have a very high coefficient of performance in winter conditions, before allowing for the energy cost of the CO_2 fraction permanently deposited under the ice. The cost of CO_2 capture from the atmosphere may be supported by carbon sequestration credits, but would not apply to locations in fast-moving ice anywhere near ice sheet margins.

The above speculative concept is fraught with many challenges and uncertainties, but could provide a worthwhile multidisciplinary research topic. Difficulties include reduction of glacier bed permeability by high effective pressure when CO_2 vapour is being released to the surface, and the risk of ice plugging the well if liquid water breaks through the deeply frozen area being formed around the well. Complexity, uncertainty and costs of this approach may well be excessive.

The above direct contact CO_2 refrigeration concept is closely related to closed thermosyphons that are widely used for passive winter cooling of foundations in permafrost zones (Wagner, 2014). Thermosyphons are essentially vertically oriented heat pipes lifting heat from working fluid evaporation at their warmer bottom end and rejecting heat from condensation at their colder top end. CO_2 and anhydrous ammonia are typical working fluids for thermosyphons. While the direct contact CO_2 refrigeration approach requires mechanical compression and pumping energy consumption to maintain the necessary pressure differences between injection and release steps during coolant escape, the “closed thermosyphon” using either CO_2 or NH_3 working fluid is entirely passive without any need for external power.

Thermosyphons are a device for moving latent heat, so heat exchange across the walls needs to be balanced for process stability with condensation at the colder top end and evaporation at the warmer bottom end. With condensation at lower temperature and evaporation at higher temperature, these devices are actually low grade geothermal heat engines, supporting their internal fluid circulation against frictional pressure drops.

Specialized thermosyphons might possibly be adapted for cooling glacier beds. Glaciers approaching their terminus typically have a very cold layer near their base, as a result of advection from the frigid high elevation interior. The temperature gradient is quite steep in the first 100 m above the base, while greatly flattening toward the surface. There is thus a choice between rejecting heat into the cold ice about 100 m above the base (as suggested in Fig. 4), or into the atmosphere at the surface (perhaps 1000 m above the base). While heat exchange into ice is conductivity-limited, the ice 100–200 m above the base may be colder than the summer surface temperature. Suitable thermosyphons contain an initially pressurized charge of working fluid in a hermetically sealed pipe enclosure installed in a vertical borehole. The nominal internal vapour working pressure in glacier applications must be intermediate between the vapour pressure of the working fluid evaporating at the freezing temperature of subglacial water, and the vapour pressure of the working fluid condensing at a practicable average heat rejection temperature about 20 °C colder. Higher pressure would prevent condensation in the heat exchanger. This nominal internal vapour working pressure would be about 2.5–3 MPa in CO_2 thermosyphons, and an order of magnitude smaller in NH_3 thermosyphons. Ammonia would be a suitable working fluid because of its much lower molecular weight and vapour pressure at temperatures near the freezing point of water, avoiding the steeper gravitational pressure gradient of compressed CO_2 vapour in a long vertical pipe. Nevertheless, ammonia releases would not be as environmentally benign as CO_2.

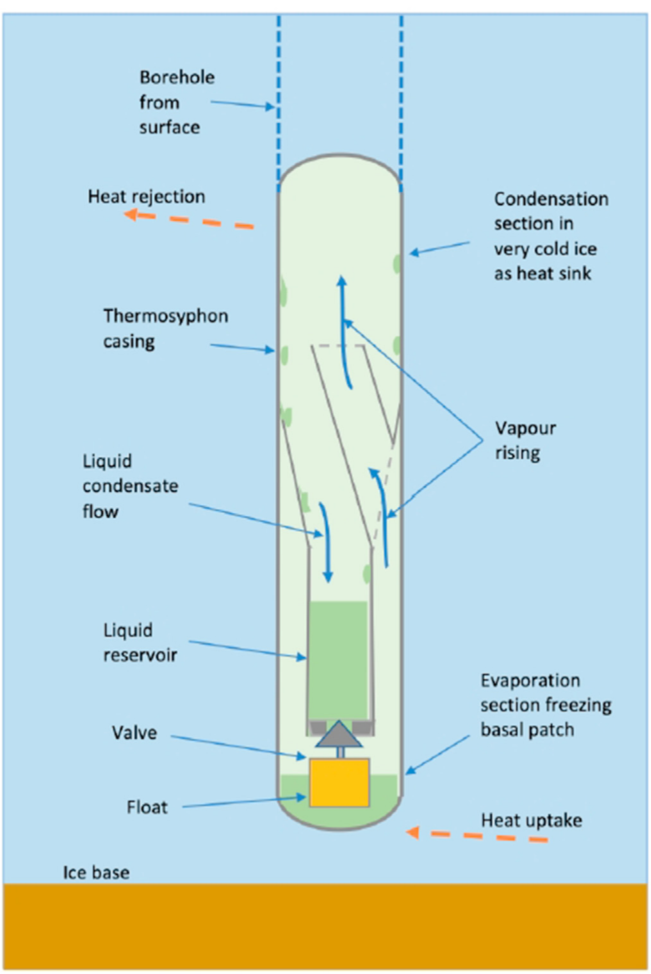

Fig. 4. Closed thermosyphon for basal refrigeration.

Conventional thermosyphons are typically no more than about 10 m tall, while glacier applications require taller devices by one or two orders of magnitude. The density of saturated liquid CO_2 at 273 K is 927 kg m^{-3}, intermediate between the densities of water and ice. Density of saturated liquid NH_3 at 273 K is 638 kg m^{-3}. The vertical gradient of boiling point elevation resulting from hydrostatic pressure will be about 0.1 K m^{-1} for CO_2 and about 0.27 K m^{-1} for NH_3. Because of the high thermal resistances faced at both ends by a thermosyphon embedded in ice, such small temperature gradients across the evaporation section may critically affect operating performance, in the direction of shutting down with excessively deep flooding. The suggestion of a float valve controlled liquid level may thus be helpful for extremely tall thermosyphons. Further study should engage qualified specialist expertise and perform detailed thermal and hydrodynamic modeling to assess feasibility of thermosyphons (with or without liquid reservoirs and float valves) as a completely passive approach to subglacial refrigeration. Careful attention should be devoted to two-phase flow conditions, internal pressure distribution and operational stability of thermosyphons at least an order of magnitude taller than any used in present applications. To avoid destructive bending and crushing, thermosyphons should not be installed in active basal fault zones. For extremely tall thermosyphons as here contemplated, liquid draining downward from the top end condensation section may be collected in a reservoir (Fig. 4) provided by a vertically elongated compartment extending between the upper condensation and lower evaporation sections of the thermosyphon, with a float actuated valve in the evaporator controlling depressurization and admission of the working fluid liquid to the evaporation section where excessive flooding or drying must be avoided. Failure of fluid control will result in instability of partial/complete failure of the device.

The low thermal conductivity of ice will greatly impair the basal cooling effectiveness of thermosyphons, though they may have elongated near-horizontal extensions of their evaporation section into the ice for greater contact area. Dense fields of thermosyphons and corresponding large numbers of boreholes are may well be prohibitively expensive.

3.5. Ice shelf buttressing

Floating ice shelves provide buttressing, resisting ice flux and mitigating MISI (Gudmundsson, 2013). This force is derived from points which contact the Earth, either at its underside where it grounds (forming rises/rumples); or at lateral margins, contacting an embayment or fjord. Even small natural pinning points have wide-ranging impact (Fürst et al., 2015).

There are two ways to increase buttressing provided by floating ice shelves. Firstly, building artificial pinning points (Wolovick and Moore, 2018). Secondly, increase thickness or structural strength of the ice shelf, so it contacts more natural points, or transmits more force from existing pinning points.

These techniques mostly require work on the ice shelf. Fringing shelves, ubiquitous around Antarctica, vary in size and accessibility. Floating glacier tongues and shelves of the Amundsen Sea are small and chaotically crevassed, because of rapid sub shelf melting driven by the warm ocean currents contacting grounding lines. Some crevasse-free regions may be available, providing a base for work.

Artificially-raised pinning points might be constructed on bedrock reefs in open water just outside present ice shelf margins, to enable glacier regrounding after ice margin advance has been stimulated by other methods such as subsea barriers (discussed in the next section) to block deep warm water ablation. Such pinning points should be strongly armoured, and might be formed as interlocking arrays of floated-in reinforced concrete caissons anchored by drilled piles. The caissons may be built by slip forming in the same way as offshore petroleum platforms, and designed to accept some crushing iceberg impact damage without loss of overall cohesive integrity. Structural durability of such artificial pinning points would need to be evaluated by future feasibility studies. Deep ice sheets would apply very high forces to such caissons, which would be subject to a variety of failure modes: being torn off moorings; snapping under bending/shear loads; surface abrasion from till; shearing of square corners; saltwater corrosion of rebar; freeze-thaw fracking of concrete; denting or crushing of hollow members; etc.

Ice shelf thickening may be applied evenly across the ice shelf, or it may be possible to organize thickening into useful shapes, such as a compressive arch (Kulessa et al., 2014) or on the shear margins. Pumping water through ice shelves via boreholes onto the surface could thicken ice in cold seasons (perhaps integrating brine rejection). A similar concept has been suggested to manage Arctic sea ice (Desch et al., 2017).

Winter cooling may be applied to the ice shelf base, using thermosyphons. The limited cooling area available (small heat exchangers, as compared to the top surface of the ice) limits use to well-chosen locations − likely with mechanical strength importance, as opposed to being a way of simply increasing ice volume. The syphon base will quickly become encased in ice, limiting heat transfer. Further investigation of such thermosyphons may explore possibilities for shedding at least some encasing ice, through use of ice-phobic coatings and perhaps cyclic mechanical impulses or water-hammer deflections generated by abrupt releases of gravitationally pressurized liquid working fluid from an internal reservoir into the evaporation section − akin to a flushing toilet. The combination of small surface area and modest heat flux (once encased) will mandate a forest of devices, probably excessively increasing costs.

About 3.1 Gt per year of snow is blown to the sea around Antarctica (Palm et al., 2019). Coastal Antarctica is subject to katabatic winds; gravity flows of cold air from the interior. These dominate stable weather patterns, except during summers. Wide flat plains ensure that much wind-blown snow ends up in the sea. Windbreaks may retain some of this snow and may fashion functional structures, e.g. arches. Windbreaks can be created by erecting fencing, or bulldozing snow. Similar techniques prevent desertification, dune movement and wind erosion in loose soils[1]. In Antarctica, structures quickly become buried − creating aerodynamically smooth features. These stop accumulating snow; windbreaks need constant maintenance.

Similarly, given the right suitable weather conditions, pumping water on top of the glacier to form an ice layer will serve to bind any snow which is there. While ice has a lower albedo than does snow, any fresh falls onto the ice surface will tend to brighten it. Ensuring that ice is laid down only in winter will minimise heating from albedo effects.

Snow control techniques could also be used on the upper reaches of glaciers, and would be more energy efficient that either pumping seawater to refreeze, or generating artificial snow.

In summary, increasing ice shelf thickness and strength are keys to stabilizing against MISI. Both reducing melting from warmer oceans and increasing stability by surface accumulation of snow might be worthwhile. Addition of water onto ice shelves however carries risks from hydrofracturing, but there is plentiful windblown snow that could usefully be restrained and sculpted.

3.6. Environmental modifications

Glaciers and ice sheets exist within a wider environment, bordered by air, rock and ocean. All can be modified, to transfer less heat into the ice, or more heat out.

Thermosyphons could passively cool basal ice or surrounding rock, though severely constrained by low thermal conductivity of that ice. Operational challenges of installation are not trivial − especially if cooling at depth. Valuable devices may be damaged or destroyed after emplacement too deeply in basal shear zones. Intervention at scale would be costly. Carefully targeted interventions may be justified.

There are already techniques which could be used to cool water contacting ice. Marine cloud brightening (MCB) postulates ships spraying fine seawater mist (Latham et al., 2012), making clouds whiter to reflect sunlight to space (Stjern et al., 2018). Similarly, marine micro bubbles brighten the ocean (Gabriel et al., 2017). Local MCB has been proposed to protect the Great Barrier Reef from warming (McDonald et al., 2019).

Various authors have suggested erection of underwater berms or suspended sheets of metal to prevent warm water reaching beneath ice sheets; modelling suggests that this is likely effective, although with challenging civil engineering (Moore et al., 2018; Wolovick and Moore, 2018; Hunt and Byers, 2018; Gurses et al., 2019). Current ongoing investigations are considering alternative approaches for installing such barriers in ice-infested waters, with suitably resilient configurations of reinforced tensile fabric. The ‘seabed anchored curtains’ would rise from the ocean floor to

[1] Scottish Natural Heritage, 2020. Dune fencing. https://www.coastalmanagement.eu/dune-fencing.

intercept the thermocline a few hundred meters below the surface, while allowing free passage of cooler water and floating ice above the thermocline. Further work is needed to determine cost-effectiveness of technically viable designs, though this approach seems quite promising.

Changing mesoscale marine currents was suggested (Hunt et al., 2019) with a focus on the Arctic. Such techniques would not necessarily be helpful in preserving local ice sheets, except possibly in special locations such as the Amundsen embayment, where the Antarctic Circumpolar Current loops around the Ross Gyre and impinges on the Amundsen Sea shelf with deep warm water entering the troughs along the continental shelf edge. Very careful simulation would be necessary, before any practical attempts.

Changing heat transfer from the atmosphere is difficult. The atmosphere absorbs ~20%−25% of solar radiation (Trenberth et al., 2009). Heat transfer from air could be locally controlled. Windbreaks and insulating sheeting are already in commercial use to preserve ski resort snow pack over the summer period. Changes to air circulation can cause local cooling but may not scale. Nevertheless, this may have some limited use in particular topographies, where crucial elements were desirable to preserve.

Falling snow matches air temperature and is difficult to change. Manipulating snow might be beneficial for other reasons. Fresh-fallen snow has a higher albedo. Cloud seeding may be beneficial, but would rely on appropriate temperatures and humidity levels; much of Antarctica has extremely low humidity. Coastal regions are the exception and cloud seeding may conceivably strengthen ice shelves, giving extra leverage to sea level mitigation.

Modifying the global radiation budget would certainly help. Boreal forest wildfires darken Greenland ice (Thomas et al., 2017). There is presently inadequate effort to control wildfires, except where settlements are endangered. Remedial silvicultural can protect boreal forest carbon stocks, while simultaneously reducing darkening of Greenland ice by black carbon fallout. Encouragement can be given to significant polluters (Russia, China) to accelerate their move away from coal.

Alternatively, using general solar radiation management may be a more achievable and realizable option than many of those presented here. While broad intervention over both polar regions would be essentially part of a globalised climate intervention (Moore et al., 2019), localised interventions could also be envisaged. These may include tropospheric aerosols, especially during heat waves. Continuous distribution would be costly, as persistence is limited. Acting only during heat waves may prove effective in Greenland, or to prevent accumulation of meltwater on ice shelves slowing hydrofracture.

In summary, modifying the non-ice environment is probably more complex in terms of side-effects than confining action to the glaciers. There are possible high-leverage locations where localized ocean currents can feasibly be diverted from MISI-sensitive glaciers. Large-scale solar radiation management is also expected to conserve ice sheets, due to general cooling; this continues to be actively researched across the world.

4. Precautionary principles

An obvious guiding principle for choosing climate interventions is no net harm. Thus, we should look critically at installing equipment needing later removal. Removal or weakening ice during construction, particularly of buttressing, should be viewed suspiciously. Temporarily removing ice, particularly floating ice, implies potential containment failure − more so than proposals that increase buttressing or thicken ice. Risky interventions include: removing ice shelves to build buttresses on the seabed; fracturing sea ice to promote increased heat exchange or precipitation.

A distinction can be drawn between destructive versus constructive approaches: ‘destructive’ being any intervention which deliberately damages a part of the ice sheet system that adds stability, such as destroying a portion of an ice shelf to build an artificial pinning point or warm-water barrier underneath it. A constructive (or preservative) version would be building the same structure with submersibles underneath the floating shelf.

Destructive interventions should not be summarily dismissed; though it seems unlikely that such approaches would be recommended after careful analysis. Because nonlinear feedbacks exist (and are uncertain), any intervention disturbing a stabilizing part of the system must be subjected to increased and extremely cautious scrutiny. Destroying shelves could trigger runaway marine ice-cliff failure (DeConto and Pollard, 2016). When in doubt, we should fear nonlinearities in the system. Implementing the destructive version thus requires more confidence in predicting ice fracture than implementing the constructive version.

In contrast, distributed hydrology systems are destabilizing. So subglacial drying would not be “destructive”, because it is not deliberately damaging a stabilizing element.

The timescale of harms and benefits must also be considered. Thickening glaciers by increasing snowfall or slowing advance will build gravitational driving stresses. Eventually, this will overcome basal friction, potentially leading to rapid acceleration and ice release if glacier retreat has been reversed to the point of substantial sea level drawdown. What timescale would such a release require? Observed behaviour of ice streams switching on centennial to millennial periods (Joughin et al., 2002; Tulaczyk et al., 2000) implies this would be far into the future; human-kind may be better able to cope than in the 21st century. Glacial geoengineering techniques should be designed with full controllability, so that interventions can readily be moderated or ended as required for overall stability.

What happens to the energy used to melt ice if it is diverted? Gurses et al. (2019) used an ocean model to explore blocking warm water entering the Amundsen and Bellingshausen Seas in West Antarctica − that is larger than interventions simulated by Wolovick and Moore (2018), conserving Thwaites glacier. They found diverted warm water would intensify melting of neighbouring ice shelves, but these ice shelves were not involved in critical buttressing on unstable glaciers (MISI). The area simulated to melt faster is the most stable parts of West Antarctica. Hence, it could be considered a success that warm water is directed to less damaging places.

5. Conclusions

This study lists techniques, some more realistic than others, to conserve ice sheets against collapse − and resulting sea level rise. These techniques all require: geo-technical analysis, ensuring efficacy; projecting engineering scrutiny, ensuring deliverability and cost-effectiveness − particularly considering the harsh, inaccessible, pristine environments involved.

There will be visual impacts and ecosystem changes, during and often after construction. The techniques listed have no guarantee of efficacy; they are not a ‘get out of jail free card’ for sea level rise.

Methods discussed include:

increasing albedo of glaciers by removing or burying dust, or by adding artificial or induced snow;

drying the glacier bed by draining sediments or freezing *in situ* − using closed-loop thermosyphons, open-loop CO_2 injection, or bedrock fracking;

thickening and thus strengthening ice shelves − by adding water in winter, anchoring emerging tabular icebergs, raising pinning points, adding artificial or induced snow, or by using windbreaks to retain natural snow;

jamming glacier flow − using cold shear margin water injection, ice shelf thickening & buttressing, and by creating marine pinning points;

altering glaciers' wider environment, using climate engineering technologies.

Inclusion on this list does not constitute recommendation, save for further analysis. Indeed, all approaches considered are somewhat problematic, and it is difficult to see how some could be made to work technically or economically.

Probably more important than technical feasibility is societal acceptance (Buck, 2018; Moore et al., 2020). Working in Greenland requires agreement by the Greenland home rule and, in Antarctica, from Antarctic Treaty voting members.

The interventions we outline require polar activities far more costly and invasive than those in recent years on the polar ice sheets. However, during the Cold War era of the 1960s, large numbers of men and associated equipment were stationed in Greenland, including substantial infrastructure on the ice sheet. The investment and commitment of the 1960s is alien to present ice sheet professionals, and most of the indigenous population. If ice sheets once again became strategic − as we suggest they must, because of the potential for catastrophic sea level rise − such infrastructure would be reinstated.

Conserving ice sheets tackles sea level rise at source, rather than mandating local adaptations. It benefits rich and poor equally, and may save small island states and coastal or low-lying areas from disappearance. Accordingly, future research is promising.

Declaration of competing interest

The authors declare no conflict of interest.

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References

Alley, K.E., Scambos, T.A., Alley, R.B., et al., 2019. Troughs developed in ice-stream shear margins precondition ice shelves for ocean-driven breakup. Sci. Adv. 5, eaax2215.

Alley, R.B., 1993. In search of ice-stream sticky spots. J. Glaciol. 39 (133), 447−454.

Alley, R.B., Anandakrishnan, S., Bentley, C.R., et al., 1994. A water-piracy hypothesis for the stagnation of ice stream C, Antarctica. Ann. Glaciol. 20 (1), 187−194.

Alley, R.B., Anandakrishnan, S., Christianson, K., et al., 2015. Oceanic forcing of ice-sheet retreat: west Antarctica and more. Annu. Rev. Earth Planet Sci. 43 (1), 207−231. https://doi.org/10.1146/annurev-earth-060614-105344.

Amundson, J.M., Fahnestock, M., Truffer, M., et al., 2010. Ice melange dynamics and implications for terminus stability, Jakobshavn Isbræ, Greenland. J. Geophys. Res.-Earth 115 (F1). https://doi.org/10.1029/2009JF001405.

Anandakrishnan, S., Alley, R.B., 1997. Stagnation of ice stream C, West Antarctica by water piracy. Geophys. Res. Lett. 24, 265−268.

Anderson, G.K., 2003. Enthalpy of dissociation and hydration number of carbon dioxide hydrate from the clapeyron equation. J. Chem. Thermodyn. 35, 1171−1183.

Åström, J.A., Vallot, D., Schäfer, M., et al., 2014. Termini of calving glaciers as self-organized critical systems. Nat. Geosci. 7 (12), 874−878. https://doi.org/10.1038/ngeo2290.

Blankenship, D.D., Bentley, C.R., Rooney, S.T., et al., 1986. Seismic measurements reveal a saturated porous layer beneath an active Antarctic ice stream. Nature 322 (6074), 54−57. https://doi.org/10.1038/322054a0.

Bougamont, M., Price, S., Christoffersen, P., et al., 2011. Dynamic patterns of ice stream flow in a 3-D higher-order ice sheet model with plastic bed and simplified hydrology. J. Geophys. Res. 116, F04018. https://doi.org/10.1029/2011JF002025.

Buck, H.J., 2018. Perspectives on solar geoengineering from Finnish Lapland: local insights on the global imaginary of Arctic geoengineering. Geoforum 91, 78−86.

Cuffey, K.M., Paterson, W.S.B., 2010. The Physics of Glaciers, fourth ed. Butterworth-Heineman/Elsevier, Burlington.

Das, S.B., Joughin, I., Behn, M.D., et al., 2008. Fracture Propagation to the Base of the Greenland ice sheet during supraglacial lake drainage. Science 320 (5877), 778−781. https://doi.org/10.1126/science.1153360.

DeConto, R.M., Pollard, D., 2016. Contribution of Antarctica to past and future sea-level rise. Nature 531 (7596), 591−597. https://doi.org/10.1038/nature17145.

Desch, S.J., Smith, N., Groppi, C., et al., 2017. Arctic ice management. Earth’s Future 5 (1), 107−127. https://doi.org/10.1002/2016ef000410.

Favier, L., Durand, G., Cornford, S.L., et al., 2014. Retreat of Pine Island Glacier controlled by marine ice-sheet instability. Nat. Clim. Change 4, 117−121. https://doi.org/10.1038/nclimate2094.

Feldmann, J., Levermann, A., Mengel, M., 2019. Stabilizing the West Antarctic ice sheet by surface mass deposition. Science Advances 5 (7). https://doi.org/10.1126/sciadv.aaw4132 eaaw4132.

Field, L., Ivanova, D., Bhattacharyya, S., et al., 2018. Increasing Arctic sea ice albedo using localized reversible geoengineering. Earth's Future 6, 882−901. https://doi.org/10.1029/2018 EF000820.

Fretwell, P., Pritchard, H.D., Vaughan, D.G., et al., 2013. Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. Cryosphere 7 (1), 375−393. https://doi.org/10.5194/tc-7-375-2013.

Frieler, K., Mengel, M., Levermann, A., 2016. Delaying future sea-level rise by storing water in Antarctica. Earth System Dynamics 7, 203−210. https://doi.org/10.5194/esd-7-203-2016.

Fürst, J.J., Durand, G., Gillet-Chaulet, F., et al., 2015. Assimilation of Antarctic velocity observations provides evidence for uncharted pinning points. Cryosphere 9, 1427−1443. https://doi.org/10.5194/tc-9-1427-2015.

Gabriel, C.J., Robock, A., Xia, L., et al., 2017. The G4Foam experiment: global climate impacts of regional ocean albedo modification. Atmos. Chem. Phys. 17 (1), 595−613. https://doi.org/10.5194/acp-17-595-2017.

Grinsted, A., 2013. An estimate of global glacier volume. Cryosphere 7, 141−151. https://doi.org/10.5194/tc-7-141-2013.

Gudmundsson, G.H., 2013. Ice-shelf buttressing and the stability of marine ice sheets. Cryosphere 7, 647−655. https://doi.org/10.5194/tc-7-647-2013.

Guo, X., Zhao, L., Gladstone, R., et al., 2019. Simulated retreat of jakobshavn isbræ during the 21st century. Cryosphere 13, 3139−3153. https://doi.org/10.5194/tc-13-3139-2019.

Gurses, O., Kolatschek, V., Wang, Q., et al., 2019. Brief communication: a submarine wall protecting the Amundsen Sea intensifies melting of neighboring ice shelves. Cryosphere 13 (9), 2317−2324. https://doi.org/10.5194/tc-13-2317-2019.

Hewitt, I.J., 2013. Seasonal changes in ice sheet motion due to melt water lubrication. Earth Planet Sci. Lett. 371−372, 16−25. https://doi.org/10.1016/j.epsl.2013.04.022.

Hinkel, J., Lincke, D., Vafeidis, A.T., et al., 2014. Coastal flood damage and adaptation costs under 21st century sea-level rise. P. Natl. Acad. Sci. USA 111, 3292−3297. https://doi.org/10.1073/pnas.1222469111.

Holland, P.R., Corr, H.F., Vaughan, D.G., et al., 2009. Marine ice in Larsen ice shelf. Geophys. Res. Lett. 36 (11), L11604.

Hughes, T., 1973. Is the West Antarctic ice sheet disintegrating? J. Geophys. Res. 78 (33), 7884−7910. https://doi.org/10.1029/JC078i033p07884.

Hunt, J.D., Byers, E., 2018. Reducing sea level rise with submerged barriers and dams in Greenland. Mitig. Adapt. Strategies Glob. Change 24, 779. https://doi.org/10.1007/s11027-018-9831-y.

Hunt, J.D., Nascimento, A., Diuana, F.A., et al., 2020. Cooling down the world oceans and the earth by enhancing the North Atlantic Ocean current. SN Appl. Sci. 2, 15. https://doi.org/10.1007/s42452-019-1755-y.

Irvine, P.J., Keith, D.W., Moore, J.C., 2018. Brief communication: understanding solar geoengineering's potential to limit sea level rise requires attention from cryosphere experts. Cryosphere 12, 2501−2513.

Jacobs, S.S., Jenkins, A., Giulivi, C.F., et al., 2011. Stronger ocean circulation and increased melting under Pine Island Glacier ice shelf. Nat. Geosci. 4 (8), 519−523. https://doi.org/10.1038/ngeo1188.

Jenkins, A., 1991. A one-dimensional model of ice shelf-ocean interaction. J. Geophys. Res. 96 (C11), 20671−20677.

Jenkins, A., 2011. Convection-driven melting near the grounding lines of ice shelves and tidewater glaciers. J. Phys. Oceanogr. 41 (12), 2279−2294. https://doi.org/10.1175/JPO-D-11-03.1.

Jevrejeva, S., Jackson, L.P., Riva, R.E.M., et al., 2016. Coastal sea level rise with warming above 2 °C. P. Natl. Acad. Sci. USA 113 (47), 13342−13347. https://doi.org/10.1073/pnas.1605312113.

Jones, A.C., Hawcroft, M.K., Haywood, J.M., et al., 2018. Regional climate impacts of stabilizing global warming at 1.5 K using solar geoengineering. Earth's Future 6 (2), 230−251. https://doi.org/10.1002/2017EF000720.

Joughin, I., Alley, R.B., Holland, D.M., 2012. Ice-sheet response to oceanic forcing. Science 338 (6111), 1172. https://doi.org/10.1126/science.1226481.

Joughin, I., Smith, B.E., Medley, B., 2014. Marine ice sheet collapse potentially underway for the Thwaites Glacier Basin, West Antarctica. Science 344 (6185), 735−738. https://doi.org/10.1126/science.1249055.

Joughin, I., Tulaczyk, S., Bindschadler, R., et al., 2002. Changes in West Antarctic ice stream velocities: observation and analysis. J. Geophys. Res. 107 (B11), 2289. https://doi.org/10.1029/2001JB001029.

Joughin, I., Tulaczyk, S., Bamber, J.L., et al., 2009. Basal conditions for Pine Island and Thwaites glaciers, West Antarctica, determined using satellite and airborne data. J. Glaciol. 55, 245−257.

Joughin, I., Smith, B.E., Howat, I.M., et al., 2010. Greenland flow variability from ice-sheet-wide velocity mapping. J. Glaciol. 56 (197), 415−430. https://doi.org/10.3189/002214310792447734.

Kamb, B., 1987. Glacier surge mechanism based on linked cavity configuration of the basal water conduit system. J. Geophys. Res. 92 (B9), 9083−9100. https://doi.org/10.1029/JB092iB09p09083.

Khazendar, A., Rignot, E., Larour, E., 2009. Roles of marine ice, rheology, and fracture in the flow and stability of the Brunt/Stancomb-Wills Ice Shelf. J. Geophys. Res. 114 (F4), F04007.

Kitous, A., Keramidas, K., 2015. Analysis of Scenarios Integrating the INDCs. European Commission.

Kulessa, B., Jansen, D., Luckman, A.J., et al., 2014. Marine ice regulates the future stability of a large Antarctic ice shelf. Nat. Commun. 5 (1), 3707. https://doi.org/10.1038/ncomms4707.

Latham, J., Bower, K., Choularton, T., et al., 2012. Marine cloud brightening. Phil. Trans. Roy. Soc. A 370, 4217−4262. https://doi.org/10.1098/rsta.2012.0086.

Liu, Y., Moore, J.C., Cheng, X., et al., 2015. Ocean-driven thinning enhances iceberg calving and retreat of Antarctic ice shelves. P. Natl. Acad. Sci. USA 112 (11), 3263−3268. https://doi.org/10.1073/pnas.1415137112.

MacMartin, D.G., Kravitz, B., 2019. Mission-driven research for stratospheric aerosol geoengineering. P. Natl. Acad. Sci. USA 116, 1089−1094.

McCusker, K.E., Battisti, D.S., Bitz, C.M., 2015. Inability of stratospheric sulfate aerosol injections to preserve the West Antarctic Ice Sheet. Geophys. Res. Lett. 42 (12), 4989−4997. https://doi.org/10.1002/2015GL064314.

McDonald, J., McGee, J., Brent, K., et al., 2019. Governing geoengineering research for the great barrier reef. Clim. Pol. 19 (7), 801−811. https://doi.org/10.1080/14693062.2019.1592742.

McMillan, M., Leeson, A.A., Shepherd, A., et al., 2016. A high resolution record of Greenland mass imbalance. Geophys. Res. Lett. 43, 7002−7010. https://doi.org/10.1002/2016GL069666.

Messenger, S., 2010. Melting glacier in Italy gets a giant thermal blanket. Treehugger. https://www.treehugger.com/clean-technology/melting-glacier-in-italy-gets-a-giant-thermal-blanket.html/. (Accessed 10 January 2020).

Moore, J.C., Reid, A.P., Kipfstuhl, J., 1994. Microstructure and electrical properties of marine ice and its relationship to meteoric and sea ice. J. Geophys. Res. 99, 5171−5180.

Moore, J.C., Grinsted, A., Zwinger, T., et al., 2013. Semi-empirical and process-based global sea level projections. Rev. Geophys. 51 (3), 484−522. https://doi.org/10.1002/rog.20015.

Moore, J.C., Gladstone, R., Zwinger, T., et al., 2018. Geoengineer polar glaciers to slow sea level rise. Nature 555, 303−305.

Moore, J.C., Yue, C., Zhao, L., et al., 2019. Greenland ice sheet response to stratospheric aerosol injection geoengineering. Earth's Future 7. https://doi.org/10.1029/2019EF001393.

Moore, J.C., Mettiäinen, I., Wolovick, M., et al., 2020. Targeted geoengineering: local interventions with global implications. Global Policy. https://doi.org/10.1111/1758-5899.12867.

Nye, J.F., 1976. Water flow in glaciers: jokulhlaups, tunnels, and veins. J. Glaciol. 17 (76), 181−207.

Palm, S., Yang, Y., Kayetha, V., 2019. New perspectives on blowing snow in Antarctica and implications for ice sheet mass balance. In: Kanao, M. (Ed.), Antarctica, A Key to Global Change. IntechOpen. https://doi.org/10.5772/intechopen.81319.

Pollard, D., DeConto, R.M., Alley, R.B., 2018. A continuum model (PSU-MEL1) of ice mélange and its role during retreat of the Antarctic Ice Sheet. Geosci. Model Dev. (GMD) 11, 5149−5172. https://doi.org/10.5194/gmd-11-5149-2018.

Pritchard, H.D., Vaughan, D.G., 2007. Widespread acceleration of tidewater glaciers on the Antarctic Peninsula. J. Geophys. Res. 112 (F3), F03S29.

Retzlaff, R., Bentley, C.R., 1993. Timing of stagnation of Ice Stream C, West Antarctica, from short-pulse radar studies of buried surface crevasses. J. Glaciol. 39, 553−561.

Rignot, E., Mouginot, J., Scheuchl, B., 2011. Ice flow of the antarctic ice sheet. Science 333 (6048), 1427−1430. https://doi.org/10.1126/science.1208336.

Rignot, E., Mouginot, J., Morlighem, M., et al., 2014. Widespread, rapid grounding line retreat of pine island, Thwaites, smith and kohler glaciers, west Antarctica from 1992 to 2011. Geophys. Res. Lett. 41, 3502−3509. https://doi.org/10.1002/2014GL060140.

Robel, A.A., 2017. Thinning sea ice weakens buttressing force of iceberg mélange and promotes calving. Nat. Commun. 8, 14596. https://doi.org/10.1038/ncomms14596.

Rothlisberger, H., 1972. Water pressure in intra- and subglacial channels. J. Glaciol. 11 (62), 177−203.

Rothlisberger, H., Lang, H., 1987. Glacial hydrology. in: Gurnell, A.M., Clark, M.J. (Eds.), Glacio-fluvial Sediment Transfer: an Alpine Perspective. John Wiley and Sons, New York, pp. 207−284.

Scambos, T.A., Hulbe, C.L., Fahnestock, M.A., 2003. Climate-induced ice shelf disintegration in the Antarctic Peninsula. Antarct. Res. 79, 79−92. https://doi.org/10.1029/AR079p0079.

Schoof, C., 2007. Ice sheet grounding line dynamics: steady states, stability, and hysteresis. J. Geophys. Res. Earth 112 (F3), F03S28. https://doi.org/10.1029/2006JF000664.

Schoof, C., 2010. Ice-sheet acceleration driven by melt supply variability. Nature 468 (7325), 803−806. https://doi.org/10.1038/nature09618.

Schroeder, D.M., Blankenship, D.D., Young, D.A., 2013. Evidence for a water system transition beneath Thwaites glacier, west Antarctica. P. Natl. Acad. Sci. USA 110 (30), 12225−12228. https://doi.org/10.1073/pnas.1302828110.

Stjern, C.W., Muri, H., Ahlm, L., et al., 2018. Response to marine cloud brightening in a multi-model ensemble. Atmos. Chem. Phys. 18, 621−634. https://doi.org/10.5194/acp-18-621-2018.

Sugiyama, M., Asayama, S., Kosugi, T., 2020. The North−South Divide on public perceptions of stratospheric aerosol geoengineering? A survey in six Asia-Pacific countries. Environmental Communication 1752−4040. https://doi.org/10.1080/17524032.2019.1699137.

Thomas, R.H., Bentley, C.R., 1978. A model for holocene retreat of the West Antarctic ice sheet. Quat. Res. 10 (2), 150−170.

Thomas, J.L., Polashenski, C., Soja, A.J., et al., 2017. Quantifying black carbon deposition over the Greenland ice sheet from forest fires in Canada. Geophys. Res. Lett. 44, 7965−7974. https://doi.org/10.1002/2017GL073701.

Trenberth, K., Fasullo, J., Kiehl, J., 2009. Earth's global energy budget. BAMS 311−323. https://doi.org/10.1175/2008BAMS2634.1.

Tulaczyk, S., Kamb, W.B., Engelhardt, H.F., 2000a. Basal mechanics of ice stream B, west Antarctica: I. Till mechanics. J. Geophys. Res.-Sol. Ea. 105 (B1), 463−481. https://doi.org/10.1029/1999JB900329.

Tulaczyk, S.M., Kamb, B., Engelhardt, H.F., 2000b. Basal mechanics of ice stream B, west Antarctica. II. Plastic-bed model. J. Geophys. Res. 105 (B1), 483−494.

Turner, J., Orr, A., Gudmundsson, G.H., et al., 2017. Atmosphere-ocean-ice interactions in the Amundsen Sea embayment, west Antarctica. Rev. Geophys. 55 (1), 235−276. https://doi.org/10.1002/2016RG000532.

van den Broeke, M., van den Broeke, M., Bamber, J.L., et al., 2009. Partitioning recent Greenland mass loss. Science 326, 984−986.

Wagner, A.M., 2014. Review of thermosyphon applications, report ERDC/CRREL TR-14-1. Cold Regions Research and Engineering Laboratory, Engineer Research and Development Center.

Walder, J., 1986. Hydraulics of subglacial cavities. J. Glaciol. 32, 439−445.

Weertman, J., 1972. General theory of water flow at the base of a glacier or ice sheet. Rev. Geophys. Space Phys. 10, 287−333.

Weertman, J., 1974. Stability of the junction of an ice sheet and an ice shelf. J. Glaciol. 13 (67), 3−11.

Winkelmann, R., Levermann, A., Martin, M.A., et al., 2012. Increased future ice discharge from Antarctica owing to higher snowfall. Nature 492 (7428), 239−242. https://doi.org/10.1038/nature11616.

Winsborrow, M., Andreassen, K., Hubbard, A., et al., 2016. Regulation of ice stream flow through subglacial formation of gas hydrates. Nat. Geosci. 9 https://doi.org/10.1038/ngeo2696.

Wolovick, M., Moore, J.C., 2018. Stopping the flood: could we use targeted geoengineering to mitigate sea level Rise. Cryosphere 12, 2955−2967. https://doi.org/10.5194/tc-12-2955-2018.

Wurzbacher, J.A., Gebald, C., Piatkowski, N., et al., 2012. Concurrent separation of CO_2 and H_2O from air by a temperature-vacuum swing adsorption/desorption cycle. Environ. Sci. Technol. 46 (16), 9191−9198.