

## Comments to editor corrections – Tom Jordan on behalf of all authors

We would like to thank Kenny Matsuoka for his supportive and helpful editorial comments. Our responses are in blue text.

Editor Decision: Publish subject to technical corrections (07 Aug 2018) by Kenny Matsuoka

Comments to the Author:

The revised manuscript addressed reviewers' concerns adequately and is much stronger than the original version. Below, I provide a few specific comments to improve this manuscript further. Because all of them are minor, I am happy to accept this manuscript subject to technical corrections. If the authors intend to revise the manuscript more substantially or in a different way from the guidance below, please contact me prior to the submission of the final manuscript.

- The 6-dB criterion used in this manuscript is supported by the phase diagram presented in Fig. 5. The authors show that this criterion satisfies (and is slightly stricter than) a classical criterion of the 10-dB reflectivity contrast in most cases. They also emphasize that the map shown in this study is a conservative estimate. In this sense, the paper is complete. However, ideally speaking, it is better if the sensitivity of geographical distribution of thawed bed is discussed for slightly different criteria. If the authors used slightly different criteria (e.g. variability > 5 dB) in their initial work and get similar spatial patterns, please mention it in this manuscript. If such results are not available, please ignore this comment.

This is a helpful comment. We would like to highlight that Fig 3c - the spatial distribution of  $\sigma_{[R]}$  - effectively contains this information (e.g. the light blue and dark blue regions indicate  $\sigma_{[R]} < 4$  dB). Moreover, the plot demonstrates that the spatially contiguous regions of low  $\sigma_{[R]}$  (i.e. no evidence for water) are largely preserved from the 6 dB threshold. We believe that this provides additional evidence for the robustness of threshold choice (in the sense that similar spatial patterns are present).

We have therefore added a sentence on this when we introduce the  $\sigma_{[R]}$  threshold (Sect. 2.6).

- Please reconsider to call "large" bed echoes as "good quality" bed echoes. I understand authors' intention to use a non-technical term but smaller bed echoes can reflect just the nature of the dry bed and can be in a good quality. I think that "bed echoes with high signal-to-noise ratio (S/N)" is a clearer way to call it, and it should be defined in such a way at least first time it appears.

Thanks for the guidance on this – we have now properly defined use of 'good' and 'poor' quality in Sect 2.1 in terms of SNR.

- slush "/" is over-used in my opinion. In many cases, they can be replaced with "and", which makes sentences clearer.

We have edited this throughout where we deemed appropriate (see marked version of the MS).

- L204-209: Spatial resolution of GISM can be mentioned earlier, at the line 204 (in the marked manuscript). The modelled ice thickness is different from the observed thickness so that the vertical variation of ice temperature is normalized to the modeled ice thickness, and then re-scaled to the observed ice thickness, I think. Statements at lines 206-208 are unclear for me, because it mentioned both this vertical scaling and probably the horizontal resolution adjustments between the 5-km-resolution model and 1-km-resolution DEM.

This interpretation is correct and thanks for pointing out the ambiguity. We have switched the paragraph ordering and added the spatial resolution of GISM earlier in the paragraph. We have also added the extra sentence: *In the calculation of <N> the temperature field was firstly interpolated to a 1 km resolution then vertically scaled using the 1 km representation of the Bamber et al. 2013 ice thickness.*

- L203: these -> those?

We could not see 'these' on L 203. We checked other usages of 'these' and could not find an error.

- L391: You said "initial results" at L369 and that only green and red data are analyzed further in Fig. 6 caption. Please mention this point in the main text as well for completeness.

This already in the main text (L 388 of the previous marked-up version). To better highlight what initial results mean we have added 'pre- sensitivity analysis' to L369. We have also added an extra sentence when introducing the results, stating that the 'red circles in results correspond to the set of red and green points in the sensitivity analysis'. Hopefully this is now clearer than before.

- L396: original paper? Replace with the full reference.

We have replaced with Jordan et al. 2016.

- L469-470: incomplete sentence?

Thanks for spotting this (we mistakenly deleted some text from the TCD paper). The full paragraph (same as TCD paper) reads: *A summary of local GHF estimates using borehole temperature profiles and thermomechanical model inversions (Weertman,1968; Dahl-Jensen et al., 1998, 2003; Greve, 2005; Buchardt and Dahl-Jensen, 2007; Petrunin et al., 2013) are provided by Rezvanbehbahani et al. (2017); Martos et al. (2018), and demonstrate general consistency between Fig. 8(c) and local estimates at GRIP, NEEM, NorthGRIP and Camp Century. Local estimates of GHF at Dye 3 ( $\sim 20\text{-}25\text{ mW m}^2$ ) are significantly lower than all three GHF models.*

- L515: spelling (space), f' "ast.

Done

- L593: remove "the".

Done

- L598 – L604: so what's your point?

Thanks – we agree this now seems out of context as we had deleted some text from the TCD paper. The point we are making is that the comparison with ice velocity is likely to be limited in utility, since the water desitbrution is largely from pre-surface melt data. We have made this clearer by changing the final sentence to: *It is therefore essential to re-emphasise that the basal water predictions generally correspond to the winter storage (pre- surface melt) configuration and may, therefore, be of limited utility in understanding spatio-temporal patterns related to ice dynamics.*

- Fig. 1C: deep ice core sites are shown in blue in this figure but in yellow in Figs. 6-10. It is better if you show the core sites all in yellow.

A good spot. We have changed the core sites to yellow.

- Figure 2a second panel from the top: the threshold red bar is displayed for  $\sigma_R = 10$  dB, not 6 dB.

We have double-checked this point and the threshold line is correct (i.e. it is consistent with  $\sigma_R = 6$  dB). To make this clearer/avoid confusion we have added to the caption that the dashed pink line refers only the *right axis*.

Thank you for submitting the valuable work to The Cryosphere.

Kenny Matsuoka  
TC/TCD Editor

#### **Additional changes**

We have now added the full reference for Martos et al. 2018 which was recently published in GRL.

# A constraint upon the basal water distribution and thermal state of the Greenland Ice Sheet from radar bed-echoes

Thomas M. Jordan<sup>1,2</sup>, Christopher N. Williams<sup>1,3</sup>, Dustin M. Schroeder<sup>2,4</sup>,  
Yasmina M. Martos<sup>5,6</sup>, Michael A. Cooper<sup>1</sup>, Martin J. Siegert<sup>7</sup>, John D. Paden<sup>8</sup>,  
Philippe Huybrechts<sup>9</sup>, and Jonathan L. Bamber<sup>1</sup>

<sup>1</sup>Bristol Glaciology Centre, School of Geographical Sciences, University of Bristol, Bristol, UK.

<sup>2</sup>Department of Geophysics, Stanford University, Stanford, CA, USA.

<sup>3</sup>Now at British Geological Survey, Nottingham, UK.

<sup>4</sup>Department of Electrical Engineering, Stanford University, Stanford, CA, USA.

<sup>5</sup>Department of Astronomy, University of Maryland, College Park, MD, USA.

<sup>6</sup>NASA Goddard Space Flight Center, Greenbelt, MD, USA.

<sup>7</sup>Grantham Institute and Department of Earth Science and Engineering, Imperial College, London, UK.

<sup>8</sup>Center for Remote Sensing of Ice Sheets, University of Kansas, Lawrence, USA.

<sup>9</sup>Earth System Science and Departement Geografie, Vrije Universiteit Brussel, Brussels, Belgium.

**Abstract.** There is widespread, but often indirect, evidence that a significant fraction of the bed beneath the Greenland Ice Sheet is thawed (at or above the pressure melting point for ice). This includes the beds of major outlet glaciers and their tributaries and a large area around the North-GRIP borehole in the ice-sheet interior. The ice-sheet scale distribution of basal water is, however, 5 poorly constrained by existing observations. In principle, airborne radio-echo sounding (RES) enables the detection of basal water from bed-echo reflectivity, but unambiguous mapping is limited by uncertainty in signal attenuation within the ice. Here we introduce a new, RES diagnostic for basal water that is associated with wet-dry transitions in bed material: bed-echo reflectivity variability. This technique acts as a form of edge detector and is a sufficient, but not necessary, criteria for basal 10 water. However, the technique has the advantage of being attenuation-insensitive and suited to data combination enabling combined analysis of over a decade of Operation IceBridge survey data.

The basal water predictions are compared with existing analyses of the basal thermal state (frozen and thawed beds) and geothermal heat flux. In addition to the outlet glaciers, we demonstrate widespread water storage in the northern and eastern interior. Notably, we observe a quasi-linear 15 ‘corridor’ of basal water extending from NorthGRIP to Petermann glacier that spatially correlates with elevated heat flux predicted by a recent magnetic model. Finally, with a general aim to stimulate regional- and process-specific investigations, the basal water predictions are compared with bed topography, subglacial flow paths, and ice-sheet motion. The basal water distribution, and its relationship with the thermal state, provides a new constraint for numerical models.

## 20 1 Introduction

Basal water beneath the Greenland Ice Sheet (GrIS) influences, and is influenced by, the dynamics and thermodynamics of the overlying ice. A lubricated bed is a necessary condition for basal sliding, which can be responsible for up to about 90% of the ice surface velocity (van der Veen, 2013). Constraining the spatial distribution of basal water is important, therefore, for understanding the dynamic  
25 state of the overlying ice and its sensitivity to external forcing. A reliable estimate of the presence of basal water can also be used as a boundary condition or a constraint in numerical modelling and to evaluate model performance and is, as a consequence, an attractive objective.

The spatial distribution of basal water beneath the GrIS is known to arise from an interplay of different physical processes including: surface melt (e.g. van de Wal et al. (2008)), basal melting  
30 due to geothermal heat (e.g. Dahl-Jensen et al. (2003); Rogozhina et al. (2016)), frictional and shear heating (e.g. van der Veen (2013)), and transport processes (surface, englacial and subglacial) which redistribute water (e.g. Rennermalm et al. (2013); Chu (2014)). There are, however, limited observational constraints on the ice-sheet scale distribution of basal water, and the relationship with other glacial and subglacial properties is therefore largely unexplored and/or speculative. The lack  
35 of unambiguous information about basal water arises primarily because there are only a few existing observations of subglacial lakes (Palmer et al., 2013; Howat et al., 2015; Willis et al., 2015; Palmer et al., 2015). By contrast, the Antarctic Ice Sheet currently has over 400 identified subglacial lakes (Siegert et al., 2016)) some of which have been found to be actively draining and recharging. Instead, there is evidence that basal water beneath the GrIS exists in smaller pools (Chu et al., 2016), as wet  
40 sediment (Christianson et al., 2014), and as part of channelised networks (Livingstone et al., 2017).

The basal temperature distribution of the GrIS determines where basal water can exist, and requires basal temperatures at or above the pressure melting point (PMP) for ice (from herein ‘thawed’). Direct basal temperature measurements are, however, sparse. At the majority of the interior boreholes - Camp Century, Dye 3, GRIP, GISP2, and NEEM - basal temperatures indicate frozen beds  
45 (Weertman, 1968; Gundestrup and Hansen, 1984; Dahl-Jensen et al., 1998; Cuffey et al., 1995; MacGregor et al., 2016), with the thawed bed at NorthGRIP an exception (Andersen et al., 2004). Toward the ice-sheet margins boreholes generally indicate thawed beds (MacGregor et al., 2016). Indirect methodologies for discriminating frozen and thawed beds (ice-sheet model predictions, radiostratigraphy, and surface properties) were recently combined by MacGregor et al. (2016), to produce a  
50 frozen-thawed likelihood map for beds beneath the GrIS. Key predictions were that the central ice divides and west facing slopes generally have frozen beds, the southern and western outlet glaciers have a thawed bed, and that basal thaw extends east from NorthGRIP over a large fraction of the northeastern ice sheet.

Spatially variable geothermal heat flux (GHF) influences the basal temperature distribution (Dahl-  
55 Jensen et al., 2003; Greve, 2005; Rogozhina et al., 2016), hydrology (Rogozhina et al., 2016), and flow features (Fahnestock et al., 2001) in the interior of the GrIS. Notably, the onset of the North

East Greenland Ice Stream (NEGIS) is predicted to arise from basal melting that is attributed to locally elevated GHF (Fahnestock et al., 2001), which, in turn, has recently been attributed to the path of the Iceland hotspot track (Rogozhina et al., 2016; Martos et al., 2018). As with basal temperature, the sparsity of borehole measurements limits direct inference of GHF (which is related to the vertical gradient of basal temperature). Instead, a range of geophysical techniques including: tectonic (Pollack et al., 1993), seismic (Shapiro and Ritzwoller, 2004), and magnetic (Fox Maule et al., 2009; Martos et al., 2018) models have been used to map GHF beneath the ice sheet. These models, however, differ greatly in the predicted spatial distribution for GHF and also in the absolute values (Rogozhina et al., 2012). Due to the relationship between basal melt and GHF, basal water observations can be used to further refine and cross-validate GHF models (Siegert and Dowdeswell, 1996; Schroeder et al., 2014; Rogozhina et al., 2016).

In principle, airborne radio-echo sounding (RES) surveys provide the information and spatial coverage to infer the presence of basal water at the ice-sheet scale. Bed-echo reflectivity is the most commonly used diagnostic for this purpose (e.g. Peters et al. (2005); Jacobel et al. (2009); Oswald and Gogineni (2008)), and is based upon the prediction that, across a range of subglacial materials, wet glacier beds have a higher reflectivity than dry or frozen beds (Bogorodsky et al., 1983; Martinez et al., 2001; Peters et al., 2005). However, due to uncertainty and spatial variation in radar attenuation (an exponential function of temperature (Corr et al., 1993)) bed-echo reflectivity is subject to spatial bias and can be ambiguous when mapped over larger regions (Matsuoka, 2011; MacGregor et al., 2012; Jordan et al., 2016). In order to mitigate spatial bias from radar attenuation, bed-echo scattering properties - including the ‘specularity’ (a measure of the angular distribution of energy and, consequently, the smoothness of the bed at a radar-scale wavelength) (Schroeder et al., 2013; Young et al., 2016) and the bed-echo ‘abruptness’ (a waveform parameter) (Oswald and Gogineni, 2008, 2012) - have also been employed in basal water detection. Although attenuation-independent, use of some bed-echo scattering properties to discriminate basal water can be prone to ambiguity and can potentially lead to false-positive detection of smoother bedrock as water (Jordan et al., 2017).

In this study we introduce a RES diagnostic for basal water that is specifically tuned to be suitable for an ice-sheet scale assessment of basal thaw. The RES diagnostic, which we term ‘bed-echo reflectivity variability’, detects wet-dry/dielectric transitions in bed material and acts as a form of edge detector. The technique is demonstrated to be insensitive to radar attenuation, thus reducing the likelihood of false-positive identification of basal water at the ice-sheet scale. Moreover, it also only requires local radiometric power calibration, and thus enables different Operation IceBridge RES campaigns, using different radar system settings (e.g. peak transmit power, antenna gain), to be combined and mapped. However, a limitation of the technique is that it provides a sufficient (not necessary) condition for basal water, and the detected subset of basal water corresponds to sharp transitions/ horizontal gradients in the bed dielectric.

The primary geographical focus of the study is to compare the spatial relationship between predicted basal water and up-to-date analyses for the basal thermal state (MacGregor et al., 2016) and GHF (Shapiro and Ritzwoller, 2004; Fox Maule et al., 2009; Martos et al., 2018) beneath the GrIS. We observe new predictions for basal water predominantly in the northern and eastern ice sheet, which spatially correlate with elevated GHF recently inferred by Martos et al. (2018). We then compare basal water and bed topography (Morlighem et al., 2017), which enables us to identify likely subglacial flow paths and storage locations beneath the contemporary ice sheet. Finally, with a view toward improving our understanding of the influence of basal water upon ice-sheet motion, we compare the relationship between basal water and ice surface speed (Joughin et al., 2010, 2016).

## 2 Methods

### 2.1 Radio-echo sounding data and bed-echo analysis

The airborne RES data used in this study were collected by the Center for Remote Sensing of Ice Sheets (CReSIS) over the time period 2003-2014. The data were taken using a succession of radar instruments: Advanced Coherent Radar Depth Sounder (ACORDS), Multi-Channel Radar Depth Sounder (MCRDS), Multi-Channel Coherent Radar Depth Sounder (MCoRDS), Multi-Channel Coherent Radar Depth Sounder: Version 2 (MCoRDS v2), mounted on three airborne platforms: P-3B Orion (P3), DHC-6 Twin Otter (TO) DC8, Douglas DC-8 (DC8) (Paden, 2015). The flight-track coverage, subdivided by radar system, is shown in Fig. 1(a) with the season breakdown: ACORDS 2003 P3 and 2005 TO; MCRDS 2006 TO, 2007 P3, 2008 TO and 2009 TO; MCoRDS 2010 P3 and 2010 DC8; MCoRDS v2 2011 TO, 2011 P3, 2012 P3, 2013 P3, 2014 P3. More details on the track lengths and data segmentation can be found in MacGregor et al. (2015a). The vast majority of the data were collected in the months March-May.

The various radar system details and signal processing steps are described in detail in previous works (Rodriguez-Morales et al., 2014; Gogineni et al., 2014; Paden, 2015). The centre-frequency of the radar systems is either 150 MHz (ACORDS and MCRDS) or 195 MHz (MCoRDS and MCoRDS v2), and, after accounting for pulse-shaping and windowing, the depth-range (vertical) resolution can vary from  $\sim 4.3$ -20 m in ice. The along-track (horizontal) resolution also varies between field seasons, and is typically  $\sim 30$  or 60 m. The radar-echo strength profiles (CSARP, Level 1B data) employ fixed fast-time gain where the receiver gain is constant for each recorded pulse which enables consistent interpretation of bed-echo power on a season-by-season basis. Whilst transmitted power can differ between seasons, since we consider local variability, offsets between seasons do not matter which enables straightforward inter-season data combination. Pre-2003 CReSIS data uses manual gain control and hence these seasons are not included.

The procedure to extract bed-echo power is similar to Oswald and Gogineni (2008); Jordan et al. (2016, 2017) and aggregates power over bed-echo fading (i.e. performs a depth-range integral).

Briefly, the procedure consists of the following steps. Firstly, CReSIS ice thickness (Level 2) picks  
 130 were used as initial estimate for the depth-range bin of the peak bed-echo power. Secondly, a local  
 re-tracker was used to locate the true depth-range bin for peak bed-echo power. Thirdly, the power  
 was aggregated by a discrete summation over the bed-echo envelope (both before and after the peak).  
 The summations were truncated when the power was 10 dB or less than the peak, thus to ensure that  
 the integral consisted of a dominant peak associated with a dominant reflecting facet. The chosen  
 135 signal-to-noise (SNR) threshold of 10 dB in this study is less strict than the  $\sim 17$  dB threshold used in  
 Jordan et al. (2016, 2017), and was required to extend the effective coverage to some interior regions  
 in southern Greenland. Additionally, in this study we did not apply along-track averaging of L1B  
 data that was done previously in Jordan et al. (2016, 2017). Finally, again, to ensure a suitable SNR,  
 a ‘quality control’ measure was imposed such that peak bed-echo power must be 10 dB above the  
 140 noise floor. This results in the effective coverage shown in Fig. 1(b) which shows ‘good’ (SNR > 10  
 dB) and ‘poor’ (SNR  $\leq$  10 dB) quality bed-echoes. The poor quality bed-echoes include a spatially  
 coherent coverage gap in the southern interior, high altitude data, and some marginal regions.

The rationale for use of aggregated bed-echo power (over peak power) is that it serves to reduce  
 bed-echo power variability due to roughness, and thus better enables comparison with the specu-  
 145 lar bed-echo reflectivity values that are used to infer bulk material properties (Oswald and Gogi-  
 neni, 2008). Additionally, since roughness scattering loss is frequency-dependent (MacGregor et al.,  
 2013), aggregated power serves as a pragmatic way to best combine bed-echo power measurements  
 from the 195 MHz and 150 MHz radar systems. This is supported by the observed  $\sim 1$  dB greater  
 scattering loss (estimated here by the dB difference between peak and aggregated power) for the 195  
 150 MHz systems at cross-over points.

Key landmarks of the GrIS that are used in the data description - temperature boreholes, drainage  
 basin boundaries, and major fast flow regions - are shown in Fig. 1(c).

## 2.2 Bed-echo power, attenuation correction and bed-echo reflectivity

The bulk material properties of glacier beds, including the presence of basal water, can, in principle,  
 155 be inferred from the reflectivity of the bed-echo (Bogorodsky et al., 1983; Peters et al., 2005). The  
 reflectivity,  $[R]$ , is obtained from solving the decibel form of the bed-echo power equation

$$[P] = [S] - [G] + [R] - [L] - [B], \quad (1)$$

where  $[P]$  is the bed-echo power (in this case the aggregated bed-echo power, which mitigates for  
 scattering losses from the ice-bed interface),  $[S]$  is the system performance,  $[G]$  is the geometric  
 160 correction,  $[L]$  is the attenuation loss in ice, and  $[B]$  is the loss due to birefringence (ice fabric  
 anisotropy), and the notation  $[X] = 10 \log_{10} X$  is assumed (Matsuoka et al., 2010). Rough surface  
 losses from transmission through the air-ice interface (e.g. Grima et al. (2014); Schroeder et al.  
 (2016a)) also influence the radar power budget, and are discussed in more detail in Sect. 4.6. The



geometric correction for a specular reflector can be defined by

$$165 \quad [G] = 2[2(s + h/\sqrt{\epsilon_{ice}})], \quad (2)$$

where  $s$  is the aircraft height and  $h$  is the ice thickness and  $\epsilon_{ice} = 3.15$  is the relative dielectric permittivity of ice to give the geometrically-corrected bed-echo power

$$[P_g] = [P] + [G], \quad (3)$$

(e.g. Schroeder et al. (2016a)). For the majority of the CReSIS data used (2006 TO onward)  $s$  and  $h$  are known and eq. (2) can be applied exactly. For the 2003 P3 and 2005 TO seasons only  $h$  is known and  $s = 500$  m is assumed (approximately the mean aircraft height). This approach is justifiable since in this study we are interested in local (length scale  $\sim 5$  km) power variation, where  $s$  varies slowly. It is assumed that variation in  $[S]$  and  $[B]$  is also negligible (again, approximations that are strengthened by consideration of local power variation) then eq. (1) reduces to

$$175 \quad [P_g] = [R] - [L], \quad (4)$$

where  $[P_g] = [P] + [G]$  is geometrically-corrected bed-echo power. Finally, setting  $[L] = 2 \langle N \rangle h$  gives

$$[P_g] = [R] - 2 \langle N \rangle h, \quad (5)$$

where  $\langle N \rangle$  (dB km<sup>-1</sup>) is the (one-way) depth-averaged attenuation rate (Matsuoka et al., 2010).

180 A fundamental ambiguity in bed-echo reflectivity analysis is that there are two unknowns in eq. (5):  $\langle N \rangle$  and  $[R]$ . Two approaches are typically used to determine  $\langle N \rangle$ : (i) ‘forward modelling’ using estimates of attenuation as a function of englacial temperature (e.g. MacGregor et al. (2007); Matsuoka et al. (2012a); Chu et al. (2016)), (ii) ‘empirical-determination’ using the linear regression of bed-echo power and ice thickness (e.g. Jacobel et al. (2009); Schroeder et al. (2016b)). Attenuation follows an Arrhenius (exponential) relationship with temperature and a linear dependence upon  
185 the concentration of ionic impurities: primarily hydrogen (H<sup>+</sup>), but also chlorine (Cl<sup>-</sup>), and ammonium (NH<sub>4</sub><sup>+</sup>) (Corr et al., 1993; Wolff et al., 1997; MacGregor et al., 2007, 2015b). On an ice-sheet scale, the uncertainty when ‘forward’ modelling  $\langle N \rangle$  is so high that it can be prohibitively challenging to accurately calibrate  $[R]$  (Matsuoka, 2011; MacGregor et al., 2015b; Jordan et al., 2016).  
190 This is due to both uncertainty in ice-sheet model temperature fields, the ionic concentrations, and the tuning of the parameters in the Arrhenius equation (MacGregor et al., 2007, 2015b). Empirical determination of  $\langle N \rangle$  using bed-echo power is also subject to sources of potential bias. In particular, the regression methods can be ill-posed when there is rapid spatial variation in attenuation (Matsuoka et al., 2012a), or when there is a thickness-correlated distribution in bed-echo reflectivity  
195 (Jordan et al., 2016).

We will later demonstrate that, unlike absolute values of  $[R]$ , local variability in bed-echo reflectivity is highly insensitive to modelled values of  $\langle N \rangle$  (Sect. 2.6). However, despite acting as a

very weak constraint, an initial estimate for ice-sheet scale variation in  $\langle N \rangle$  is still necessary to calculate reflectivity variability. The estimate for  $\langle N \rangle$  relies on previous work by Jordan et al. (2016) and uses the ‘M07’ Arrhenius equation MacGregor et al. (2007), the Greenland Ice Sheet Model (GISM) temperature field from Huybrechts (1996) as updated in Goelzer et al. (2013), depth-averaged ionic concentrations from the GRIP ice core (MacGregor et al., 2015b), and the Greenland ice thickness data set in Bamber et al. (2013a). The temperature field derives from a full 3D thermomechanical simulation over several glacial-interglacial cycles and resolves the flow on a model resolution of 5 km, which causes some smoothing of the temperature field in narrow outlet glaciers near to the coast. In the calculation of  $\langle N \rangle$  the temperature field was firstly interpolated to a 1 km resolution then vertically scaled using the 1 km representation of the Bamber et al. (2013a) ice thickness. The geothermal heat flux in GISM was initially taken from Shapiro and Ritzwoller (2004) but further adjusted with Gaussian functions around the deep ice core sites to match observed basal temperatures. Vertical temperature profiles are within 1-2°C when compared to available in-situ measurements. ~~GISM resolves the flow on a model resolution of 5 km, which causes some smoothing of the temperature field in narrow outlet glaciers near to the coast.~~

Using this model framework, it is that predicted that  $\langle N \rangle$  varies by a factor  $\sim 5$  over the GrIS, ranging from  $\sim 6 \text{ dB km}^{-1}$  in the colder northern interior to  $\sim 30 \text{ dB km}^{-1}$  toward the warmer southwestern margins (refer to Fig. 5(a) in Jordan et al. (2016) for a spatial plot).

### 2.3 Calculating bed-echo power and reflectivity variability

In this study we use simple standard deviation measures to quantify the variability of bed-echo power and reflectivity along the flight-tracks, notated by:  $\sigma_{[P_g]}$  and  $\sigma_{[R]}$ . The numerical calculation is analogous to how topographic roughness (the rms height) is calculated from bed elevation profiles (Shepard et al., 2001), and assumes an along-track ‘window’ of length 5 km evaluated at 1 km intervals (taking the standard deviation of all the points within the the window). The choice of horizontal length scale was chosen for consistency with the basal thermal state mask by MacGregor et al. (2016), which was deemed an appropriate scale for integration of radar data with thermomechanical models at the ice-sheet scale. At this 5 km length scale, high values of reflectivity variability - a diagnostic for ~~dielectric~~/wet-dry transitions (also, see Sect. 2.5) - act as a form of edge detector. The rationale for using this approach (rather than a conventional edge detector) is that it does not impose that a single along-track transition is present, which is desirable when aiming to also detect multiple smaller water bodies relative to the window size.

Since this application of along-track variability is a non-standard approach to bed-echo data analysis, we now take a closer look at the statistical properties. The formula for  $\sigma_{[P_g]}$  follows from the variance of eq. (4) and is given by

$$\sigma_{[P_g]} = \sqrt{\sigma_{[R]}^2 + \sigma_{[L]}^2 - 2\sigma_{[R],[L]}}, \quad (6)$$

where  $\sigma_{[L]}$  is the standard deviation in attenuation loss, and  $\sigma_{[R],[L]}$  is the covariance of bed-echo reflectivity and attenuation loss. Using  $[L] = 2 \langle N \rangle h$  and assuming  $\langle N \rangle$  can be approximated  
 235 as constant (justifiable at the 5 km length scale that is considered) then eq. (6) becomes

$$\sigma_{[P_g]} = \sqrt{\sigma_{[R]}^2 + 4 \langle N \rangle^2 \sigma_h^2 - 4 \langle N \rangle \sigma_{[R],h}}, \quad (7)$$

where  $\sigma_h$  is the standard deviation of ice thickness and  $\sigma_{[R],h}$  is the covariance of bed-echo reflectivity and ice thickness. Equations (6) and (7) consider the variability of log-transformed variables  
 240 (i.e. the dB/additive form of the radar power equation). This differs from the variability of the linear variables (i.e. the multiplicative form of the radar power equation) expressed in log space. The first advantage to our use of log-transformed variables is that we can better separate the variability contributions into separate components. The second advantage is that we can make a clearer connection to the dB reflection amplitudes that are typically used in radioglaciology (Bogorodsky et al., 1983; Peters et al., 2005). This includes prior applications that have considered the distribution and  
 245 standard deviation of dB reflection amplitudes in the context of water detection (e.g. Wolovick et al. (2013); MacGregor et al. (2013)).

In regions where  $\sigma_{[R],[L]} \approx 0$ , (bed-echo reflectivity has negligible covariance with attenuation loss), eq. (6) and eq. (7) are approximated by

$$\sigma_{[P_g]} \approx \sqrt{\sigma_{[R]}^2 + \sigma_{[L]}^2}, \quad (8)$$

$$250 \quad \approx \sqrt{\sigma_{[R]}^2 + 4 \langle N \rangle^2 \sigma_h^2}, \quad (9)$$

and the loss component of  $\sigma_{[P_g]}$  is solely modulated by  $\sigma_{[L]}$  which is proportional to the product  $\langle N \rangle \sigma_h$ . Whilst an approximation (in certain circumstances the second and third terms on the right hand side of eq. (6) and eq. (7) can be of comparable magnitude) this scenario provides an intuitive way to understand the interrelationship between  $\sigma_{[P_g]}$ ,  $\sigma_{[R]}$  and  $\sigma_{[L]}$ .

255 Two example profiles for  $[P_g]$ ,  $\sigma_{[P_g]}$ ,  $[R]$ ,  $\sigma_{[R]}$ ,  $[L]$ ,  $\sigma_{[L]}$  and  $h$  are shown in Fig. 2. Fig. 2(a) is a representative example from the interior of the ice sheet where  $\sigma_{[L]}$  is relatively low and  $h$  is thick ( $\sim 2.8$  km). Subsequently the profiles for  $\sigma_{[P_g]}$  and  $\sigma_{[R]}$  are very similar in appearance, with the most notable difference at distance  $\sim 360$  km where there is higher  $\sigma_{[P_g]}$  due to the subglacial trough. The peaks in  $\sigma_{[R]}$  are later related to wet dry bed material transitions in Sect. 2.5. It is also important  
 260 to note that  $\sigma_{[R]}$  can be greater than  $\sigma_{[P_g]}$  (e.g. at distance  $\sim 342$  km), which is an effect that can be explained by the covariance between attenuation loss/ice thickness and bed-echo reflectivity in eq. (6) and eq. (7)). Fig. 2(b) is a representative example from toward the ice-sheet margins where  $\sigma_{[L]}$  is higher due to more rapid variation in  $h$  (more complex bed topography) and higher values of  $\langle N \rangle$  (warmer ice). In this case,  $\sigma_{[P_g]}$  is noticeably greater than  $\sigma_{[R]}$ , with the differences largely  
 265 attributable to higher  $\sigma_{[L]}$  as anticipated by eq. (8). The examples in Fig. 2 highlight that, at this length scale, higher values of  $\sigma_{[R]}$  arise primarily due to a large single transition in  $[R]$ . However, there are also instances where multiple transitions result in local variability peaks (e.g. at distance  $\sim 380$ -385 km along 2(a)).

When calculating  $\sigma_{[P_g]}$ ,  $\sigma_{[R]}$  and  $\sigma_{[L]}$ , bed-echo coverage gaps within a 5 km bin (see Fig. 1(b)) were accounted for by neglecting bins where less than half the data corresponded to ‘good’ bed-echoes. The effects of this filtering step are demonstrated in Fig. 2(b) where, aligned with the deep subglacial trough at distance  $\sim 1324$  km, there are along-track gaps in  $\sigma_{[P_g]}$ ,  $\sigma_{[R]}$  and  $\sigma_{[L]}$ .

It is important to clarify the difference between the use of bed-echo power and reflectivity variability in this study from previous radioglaciology studies (Neal, 1982; Peters et al., 2005; Carter et al., 2007; Grima et al., 2014). These studies focused upon the ~~variability/statistics~~ statistical distribution of the peak echo power as a result of phase modulation by interfacial roughness. By contrast, in this study we suppress roughness effects by integrating power in fast time over the echo envelope (Sect. 2.1). We are therefore able to focus upon power variability that is a result of along-track changes in the bed dielectric.

## 2.4 Distributions for bed-echo power and reflectivity variability

Spatial distributions for the variability measures:  $\sigma_{[P_g]}$ ,  $\sigma_{[L]}$ ,  $\sigma_{[R]}$  are shown in Fig. 3(a,b,c) along with ice thickness (Morlighem et al., 2017) in Fig. 3(d). In general,  $\sigma_{[P_g]}$  has a strong ice thickness dependence, and increases toward the margins where ice is thinner. The attenuation correction, which primarily acts to reduce the component of  $\sigma_{[P_g]}$  that is attributable to  $\sigma_{[L]}$ , results in a more uniform ice-sheet scale distribution of  $\sigma_{[R]}$  than  $\sigma_{[P_g]}$ . Notably, there are localised patches of higher  $\sigma_{[R]}$  present in both marginal and interior regions (which are later attributed to the presence of basal water). The ice-sheet scale trends in  $\sigma_{[P_g]}$  and  $\sigma_{[L]} = 2 < N > \sigma_h$  can be related to spatial variation in  $< N >$  (MacGregor et al., 2015b; Jordan et al., 2016) and bed roughness (Rippin, 2013; Jordan et al., 2017) (which correlates with  $\sigma_h$ ).

The two zoom regions in Fig. 3 include the flight-track profiles in Fig. 2. (north-central ice sheet, pink bounding box; northwestern margins, black bounding box). These examples serve to further illustrate the spatial interrelationship between  $\sigma_{[P_g]}$ ,  $\sigma_{[R]}$  and  $\sigma_{[L]}$  in a typical interior region with lower  $\sigma_{[L]}$  and a typical marginal region with higher  $\sigma_{[L]}$ . It is clear that the interior example has very similar spatial distributions for  $\sigma_{[P_g]}$  and  $\sigma_{[R]}$ , whereas the marginal example has higher  $\sigma_{[P_g]}$  associated with the higher  $\sigma_{[L]}$  that occurs in the subglacial troughs and more complex topography toward the edge of the ice sheet. The marginal example also demonstrates that the power variability associated with the subglacial troughs is largely removed for  $\sigma_{[R]}$ .

The corresponding frequency distributions for  $\sigma_{[P_g]}$  and  $\sigma_{[R]}$  are shown in Fig. 4. Both demonstrate a strong positive-skew, with a long-tail extending to higher values. The mean and standard deviation for  $\sigma_{[P_g]}$  is greater than  $\sigma_{[R]}$ . This is consistent with the commonly observed result that making an attenuation correction to  $[P_g]$  acts to reduce the overall decibel range for  $[R]$ , (e.g. Oswald and Gogineni (2008); Schroeder et al. (2016b)), hence more closely resembling the predicted dB range for bed materials (Peters et al., 2005).

A quantification and discussion of cross-over statistics for  $\sigma_{[R]}$  is given in Appendix A.

## 305 2.5 Interpretation of reflectivity variability as a sufficient diagnostic for basal water

Radar bed-echo reflectivity depends upon the dielectric contrast between glacier ice and bed material. For a specular, nadir reflection the Fresnel power reflection coefficient is given by

$$[R] = 10 \log_{10} \left| \frac{\sqrt{\epsilon_{bed}^{\sim}} - \sqrt{\epsilon_{ice}^{\sim}}}{\sqrt{\epsilon_{bed}^{\sim}} + \sqrt{\epsilon_{ice}^{\sim}}} \right|^2, \quad (10)$$

where  $\epsilon_{ice}^{\sim}$  and  $\epsilon_{bed}^{\sim}$  are the complex dielectric permittivity of the glacier ice and bed material respectively. The relative (real) part of the permittivity,  $\epsilon_{bed}$ , is the primary control upon  $[R]$ . A summary of dielectric and reflective properties of glacier bed materials at typical ice-penetrating radar frequencies is given in Table 1 and is collated from Bogorodsky et al. (1983); Martinez et al. (2001); Peters et al. (2005). The permittivity and reflectivity range for each material arises due to sub-wavelength dielectric mixing between either ice or water and the bed material, and takes into account typical saturation and porosity values (Martinez et al., 2001; Peters et al., 2005). In general, lower values of  $\epsilon_{bed}$  and  $[R]$  occur for dry or frozen bed materials (approximately  $\epsilon_{bed} < 7$  and  $[R] < -14$  dB), whilst higher values occur for wet bed materials (approximately  $\epsilon_{bed} > 7$  and  $[R] > -14$  dB). Dielectric mixing between bed materials can also occur at the length scale of the Fresnel zone ( $\sim 100$  m), which results in a further averaging of the observed reflectivity (Berry, 1975; Peters et al., 2005).

Due to the range of possible bed materials at the ice-sheet scale it is not possible to formulate a unique dielectric model for diagnosing water from  $\sigma_{[R]}$ . A simple ‘two-state’ dielectric model, does, however, enable us to physically motivate the water diagnostic in terms of dielectric properties (Fig. 5). The model assumes that the along-track sample window is comprised of two different bed materials: the dry ‘background’ bed material with permittivity  $\epsilon_{dry}$  and reflectivity  $[R_{dry}]$ , and the wet material with permittivity  $\epsilon_{wet}$  and reflectivity  $[R_{wet}]$ . For simplicity, it is assumed that each along-track measurement is in one of the wet or dry states, with the wet-dry mixing ratio parameterised by  $f$ . In this formulation, a single body of wet material or multiple smaller bodies of wet material, have the same formula for the reflectivity variability given by

$$\sigma_{[R]} = \Delta[R] \sqrt{f^2(1-f) + (1-f)^2 f}, \quad (11)$$

where  $\Delta[R] = [R_{wet}] - [R_{dry}]$  is the reflectivity difference between wet and dry beds. Eq. (11) is derived by considering the weighted variance for two discrete random variables and does not account for non-linear variations due to variable scattering coherence. A phase-space plot for  $\sigma_{[R]}(f, \Delta[R])$  is shown in Fig. 5(c), and shows that for fixed  $\Delta[R]$ ,  $\sigma_{[R]}$  is maximised when  $f = 0.5$  (i.e. an even mixing of wet and dry materials).

Past diagnosis of basal water typically associates the upper tail of the reflectivity distribution with water, prescribing a threshold above which the bed is interpreted as wet (e.g. Jacobel et al. (2009); Chu et al. (2016)). In this study, a similar thresholding approach is applied to the distribution of  $\sigma_{[R]}$  (Fig. 4(b)). The threshold choice for basal water ( $\sigma_{[R]} > 6$  dB) corresponds to the region greater

than the  $\sigma_{[R]} = 6$  dB contour in Fig. 5(c) and requires a minimum wet-dry reflectivity difference of  
340  $\Delta[R] > 12$  dB. In general,  $\Delta[R] > 12$  dB is only possible for a mixture of wet and dry (or frozen)  
bed materials (Table 1). For example, an even mixing of ground water and dry granite ( $f = 0.5$ ,  
 $\Delta[R] = 17$  dB) has  $\sigma_{[R]} = 8.5$  dB. The contours in Fig. 5(c) demonstrate that small perturbations  
to even mixing ( $f \neq 0.5$ ) produce similar  $\sigma_{[R]}$ , and hence that water detection is insensitive to the  
discretisation of the along-track sample window (Sect. 2.3). Overall, the threshold choice ( $\sigma_{[R]} > 6$   
345 dB) is fairly conservative and is deliberately intended to reduce false-positive detection of basal water  
(at the expense of reduced overall detection). Finally, as discussed in Sect. 2.3, the 6 dB variability  
threshold applies to the distribution of log-transformed reflectivity.

The bed-echo power aggregation in Sect. 2.1 partially mitigates for roughness-induced scattering  
loss and the along-track power variability associated with this. Supporting evidence is that the cross-  
350 over analysis for  $\sigma_{[R]}$  (see Appendix A) demonstrates there to be no significant bias for the 150  
MHz radar systems (ACORDS and MCRDS) versus MCORDS v2 (the 195 MHz radar system used  
as a benchmark). Additionally, whilst small-scale roughness transitions (transitions from specular to  
diffuse scattering) will often correlate with wet-dry transitions, this scenario will act to amplify the  
useful signal component with the  $\sigma_{[R]}$  value increasing.

355 Table 1 also indicates that, in exceptional circumstances,  $\Delta[R] > 12$  dB could be generated in  
frozen/dry regions that partially contain sandstone or till that is close to matching the permittivity  
of ice. However, if present, these regions are likely to have indistinct bed-echoes and will not be  
included in the effective coverage in Fig. 1(b).

## 2.6 Basal water distribution and robustness to attenuation model bias

360 The initial basal water predictions ( $\sigma_{[R]} > 6$  dB, [pre- sensitivity analysis](#)) are shown in Fig. 6 (red,  
blue and green data), and correspond to  $\sim 3.5$  % of bins containing predominantly ‘good’ quality  
bed-echoes (Fig. 1(b)). A full geographic analysis of the spatial distribution is performed in Sect. 3.  
To demonstrate the robustness of the predictions, we performed a sensitivity analysis with respect to  
the modelled attenuation correction  $\langle N \rangle$  (Sect. 2.2) The analysis considered a series of increas-  
365 ingly large (uniform, multiplicative) perturbations to  $\langle N \rangle$  and then tested if  $\sigma_{[R]} > 6$  dB also held  
for the perturbed model. Examples of ‘persistent’ water predictions for  $\pm 20$  % (red and green data)  
and  $\pm 50$  % (red data) perturbations to  $\langle N \rangle$  are indicated. As the perturbation size increases this  
results in a slight decrease in the overall percentage of water predictions (corresponding to  $\sim 2.6$  %  
and 2.1 % of the along-track bins for  $\pm 20$  % and  $\pm 50$  % respectively).

370 The sensitivity analysis tests the robustness of the water predictions to a number of different phys-  
ical scenarios. Firstly, inherent bias in the Arrhenius equation parameters. For example, an empirical  
correction similar to the uniform perturbation considered in Fig. 6 was proposed by MacGregor et al.  
(2015b) to model unaccounted frequency-dependence in the electrical conductivity. Secondly, bias  
in the model temperature field ( $\langle N \rangle$  is approximately equivalent to depth-averaged temperature).

375 Thirdly, bias due to assumed ionic concentration values. It is hard to formally quantify the possible  
range of these uncertainties but, based upon solution variability for  $\langle N \rangle$  using ice-sheet model  
temperature fields (Jordan et al., 2016),  $\pm 20\%$  is a reasonable estimate for temperature related un-  
certainty. Subsequently, in the comparison with other data sets in Sect. 3 the subset of red and green  
points in Fig. 6 is used. Inherent bias in the Arrhenius equation parameters could be significantly  
380 higher than temperature uncertainty (MacGregor et al., 2015b). However, since the spatial structure  
for the basal water distribution under the  $\pm 50\%$  perturbation is largely preserved, this is unlikely  
to significantly alter the conclusions that are drawn. Additionally, whilst  $\sigma_{[R]} > 6$  dB is used as the  
threshold, Fig. 3(c) demonstrates that a less conservative threshold ( $\sigma_{[R]} > 4$  dB) retains the majority  
of contiguous regions with no water predicted (e.g. the interior of southern Greenland).

385 It is important to emphasise the robustness of  $\sigma_{[R]}$  with respect to uncertainty and model bias  
in  $\langle N \rangle$  (particularly compared with bed-echo reflectivity,  $[R]$ ). An analogous sensitivity analysis  
by Jordan et al. (2016) demonstrated that systematic over and underestimates in  $\langle N \rangle$  lead to  
pronounced ice thickness-correlated biases in the distribution for  $[R]$  in northern Greenland (Fig. B1  
in Jordan et al. (2016)).

### 390 3 Results

The basal water distribution is now compared with existing analyses for the basal thermal state  
(MacGregor et al., 2016) (Sect. 3.1), geothermal heat flux (GHF) (Shapiro and Ritzwoller, 2004;  
Fox Maule et al., 2009; Martos et al., 2018) (Sect. 3.2), bed topography and subglacial flow paths  
(Morlighem et al., 2017) (Sect. 3.3), and ice surface speed (Joughin et al., 2010, 2016) (Sect. 3.4).  
395 The basal water predictions in the results are always indicated by red circles and correspond to the  
set of red and green points in the sensitivity analysis, Fig. 6. In regional zoom plots the circles are  
fixed to be 5 km in diameter (a true representation of the along-track window size and the effective  
resolution of the radar method). In ice-sheet scale plots the buffer size of the water predictions are  
increased for visualisation purposes. The radar flight-tracks represent where there are ‘good’ bed-  
400 echoes (Fig. 1(b)), and hence indicate the effective coverage.

In interpreting the maps it is important to emphasise that the basal water predictions in this study  
correspond to a subset of flight-track data where basal water is present. Specifically, they correspond  
to where there are rapid spatial transitions and gradients in the bed dielectric (i.e. small, finite, wa-  
ter bodies or the boundaries of larger water bodies). The predictions therefore act as a sufficient  
405 constraint upon the distribution of basal water rather than being a fully comprehensive flight-track  
map for water extent. Additionally, since the vast majority of the radar measurements were col-  
lected before the onset of summer surface melt, to a first approximation, the basal water predictions  
correspond to the winter storage configuration.

### 3.1 Comparison between basal water distribution and basal thermal state synthesis

410 In Fig. 7(a) the basal water predictions are underlain by the basal thermal state synthesis (frozen/thawed likelihood) map by MacGregor et al. (2016). The synthesis employed four independent methods: (i) assessment of thermomechanical model temperature fields, (ii) basal melting inferred from radiostratigraphy, (iii) basal motion inferred from surface velocity, (iv) basal motion inferred from surface texture. The four methods were then equally weighted, leading to a likelihood map for frozen beds, 415 thawed beds, and uncertain regions. Importantly, the prediction did not utilise radar bed-echo data and is therefore independent of our basal water predictions.

The reflectivity variability water diagnostic enables a positive discrimination of basal thaw, since  $\sigma_{[R]} > 6$  dB is deemed as a sufficient (but not necessary) criteria for basal water. Positive discrimination of frozen regions is not, however, possible. This is because low reflectivity variability ( $\sigma_{[R]} <$  420 6 dB) could correspond to many different scenarios: a frozen region, a drier region at or above the PMP, or a wet region that is smoothly varying in bed-echo reflectivity. Since basal water enables a positive discrimination of thaw, red circles in likely thawed (pink) regions indicate agreement and red circles in likely frozen (blue) regions indicate disagreement with the basal thermal state synthesis. Absence of basal water in likely frozen regions is an indicator of general consistency between 425 the two methods.

There is general agreement (water in predicted thawed regions) for the beds of major outlet glaciers and their upstream regions. This includes Helheim, Kangerlussuaq, Jakobshavn, and the other fast flowing regions identified in Fig. 1(c). There is also general agreement between basal water and the extent of predicted thaw in the NEGIS drainage basin. Major regions of disagreement 430 (water in predicted frozen regions) are highlighted in the zoom plots, Fig. 7(b)-(f). The most obvious disagreement is the quasi-linear ‘corridor’ of basal water in the north-central ice sheet (Fig. 7(d)). This feature tracks close to the central ice divides and extends from the NorthGRIP region in the south toward Petermann glacier in the north. There are also noticeable areas of disagreement to the north and east of the Camp Century borehole (Fig. 7(b)), in the far north (Fig. 7(c)), to the east of 435 GRIP (Fig. 7(e)), and around Kangerlussuaq (Fig. 7(f)). There is also an absence of water in many predicted frozen regions indicating consistency. This includes parts of the southern interior, north of the NEGIS drainage basin, and the majority of the interior region between the Camp Century and NEEM boreholes.

### 3.2 Comparison between basal water distribution and geothermal heat flux models

440 The basal temperature of glacier ice is governed by surface temperature, GHF, strain heating from internal deformation, frictional heating, and diffusive and advective heat transport (e.g. van der Veen (2013)). In the interior of the ice sheet, close to the ice divides, GHF, vertical advection, and diffusion are the dominant processes which influence basal temperature. In this scenario, the thermodynamic



(temperature) equation can be approximated by the classical Robin model which predicts that basal  
445 melting occurs when GHF is above a certain threshold. For typical values of ice thickness and surface  
accumulation rate, which controls the rate of vertical heat advection, the minimum GHF forcing for  
melt is anticipated to be  $\sim 55\text{-}70 \text{ mW m}^{-2}$  in the interior of the ice sheet (Dahl-Jensen et al., 2003;  
Greve, 2005; Buchardt and Dahl-Jensen, 2007). In the water-GHF comparison we therefore define  
'elevated' GHF (i.e. likely to produce basal melt) as  $> 60 \text{ mW m}^{-2}$ . This definition is also informed  
450 by the lower range of values, ( $37\text{-}50 \text{ mW m}^{-2}$ ) that are typically associated with non-altered ancient  
continental crust (Artemieva, 2006; Rogozhina et al., 2016).

In Fig. 8 the basal water predictions are underlain by three different GHF models: the seismic  
model by Shapiro and Ritzwoller (2004), and two models derived from magnetic anomalies by  
Fox Maule et al. (2009) and Martos et al. (2018). The GHF model by Shapiro and Ritzwoller (2004)  
455 is based upon the correlation between a 3D tomographic model of the crust and mantle tempera-  
ture. The GHF models by Fox Maule et al. (2009) and Martos et al. (2018) are based on a thermal  
model of the lithosphere with the lower boundary defined by the Curie depth which is determined  
from magnetic anomalies. Martos et al. (2017) further describes this approach and the additional  
spectral processing method used to produce Fig. 8(c). An older tectonic GHF model by (Pollack  
460 et al., 1993) is not considered and a spatial plot for the GrIS can be found in Rogozhina et al. (2012)  
along with a discussion of the caveats of the different types of model. A summary of local GHF  
estimates using borehole temperature profiles and thermomechanical model inversions (Weertman,  
1968; Dahl-Jensen et al., 1998, 2003; Greve, 2005; Buchardt and Dahl-Jensen, 2007; Petrunin et al.,  
2013) are provided by Rezvanbehbahani et al. (2017); Martos et al. (2018), and demonstrate general  
465 consistency between Fig. 8(c) and local estimates at GRIP, NEEM, NorthGRIP and Camp Century.  
Local estimates of GHF at Dye 3 ( $\sim 20\text{-}25 \text{ mW m}^{-2}$ ) are significantly lower than all three GHF  
models.

In interpreting Fig. 8, we limit the comparison to the ice sheet interior where the spatial corre-  
lation between GHF and basal water should be strongest. The model by Shapiro and Ritzwoller  
470 (2004), Fig. 8(a), predicts low GHF ( $< 60 \text{ mW m}^{-2}$ ) over the vast majority of the central and north-  
ern interior. There is therefore no correlation between elevated GHF and basal water. The model by  
Fox Maule et al. (2009), Fig. 8(b), predicts elevated GHF around GRIP and the southern and eastern  
boundaries of the NEGIS basin and basal water is also present in this region. There is, however, no  
correlation between elevated GHF and basal water along the ice divides north of NorthGRIP. The  
475 model by Martos et al. (2018), Fig. 8(c), exhibits strong overall spatial correlation between basal wa-  
ter and elevated GHF in the interior of the northern ice sheet. Notably, there is a striking correlation  
between elevated GHF and the quasi-linear 'corridor' of basal water that extends from NorthGRIP  
toward Petermann glacier. All three models predict regions of elevated GHF in the southern interior  
including the Dye 3 region. However, there is only isolated radar evidence for basal water.

480 In the comparison between the flight-track water predictions and GHF distributions in Fig. 8 it is important to bear in mind that the GHF distributions are evaluated at a lower spatial resolution. For example, the resolution of the GHF distribution by Martos et al. (2018) is a consequence of the spectral method (window size and overlap) which has an effective resolution of  $\sim 75$  km.

### 3.3 Comparison between basal water distribution, bed topography and subglacial flow paths

485 In Fig. 9 the basal water predictions are underlain by the most recent Greenland bed topography digital elevation model (DEM) (Morlighem et al., 2017). To motivate further discussion about water storage locations and hydrological connectivity, a predicted subglacial flow path network is also included. The network structure is governed by gradients in the hydraulic pressure potential (Shreve, 1972) which was calculated using the bed elevation and ice thickness surfaces at a grid cell resolution of 600 m (derived from Morlighem et al. (2017)). The flow-routing algorithm was implemented in ArcGIS using the inbuilt flow accumulation tool on a hydraulic potential surface that had been processed for sink removal using the method of Wang and Li (2006). Likely hydrological flow paths were identified by excluding flow paths where 50 or fewer neighbouring cells cumulatively contribute to a given location.

495 Fig. 9 demonstrates that the vast majority of the basal water predictions are well aligned with predicted subglacial flow paths. This alignment is most visually pronounced toward the margins and zoom plots are shown for the Petermann catchment in Fig. 9(b) and northwestern margins in Fig. 9(c). Fig. 9(b) also demonstrates that basal water is present along sections of the ‘mega-canyon’ feature identified by Bamber et al. (2013b) - for example, north west of the intersection ( $80^{\circ}\text{N}$ ,  $50^{\circ}\text{W}$ ). In the interior of the ice sheet, where the horizontal gradients in ice thickness are small, local minima **sinks** in the hydraulic potential surface should correlate with topographic depressions. The water storage locations in the interior generally conform to this behaviour (Fig. 9(d)).

### 3.4 Comparison between basal water distribution and ice surface speed

In Fig. 10 the basal water predictions are underlain by ice surface speed (Joughin et al., 2016) which is based upon a temporal average from 1st December 1995 to 31 October 2015. The ice surface speed is determined using interferometric synthetic aperture radar (InSAR) as described in Joughin et al. (2010). Whilst there is a complex overall relationship between basal water and ice velocity, there are some clear spatial patterns. Notably, in the topographically less constrained northern and western outlet glaciers, basal water is often concentrated in the fast-flow onset regions and tributaries whilst it is absent from the main trunks. This behaviour is particularly evident for the Petermann glacier catchment (Fig. 10(b)). In the topographically more constrained southeastern outlet glaciers, there is widespread evidence for basal water storage in both the fast flowing glacial troughs and upstream regions. This includes both the Kangerlussuaq catchment and the tight network of subglacial troughs to the south (Fig. 10(c)), and the Helheim catchment (Fig. 10(d)).

515 In the interior of the ice sheet basal water is predicted near to the head of the NEGIS ice stream. However, basal water is also predicted in some of the slowest flowing regions of the ice sheet interior. Notably, close to the central ice divides between NorthGRIP and Petermann and north east of GRIP.

## 4 Discussion

### 4.1 Basal water, basal thermal state and temperature

520 The basal water distribution in this study and the basal thermal state synthesis by MacGregor et al. (2016) represent two independent approaches to predict where the bed beneath the GrIS is thawed. There is greatest agreement (basal water in likely thawed regions identified by MacGregor et al. (2016)) toward the ice margins where ice surface speed is generally higher. The most noticeable regions of disagreement (basal water in likely frozen regions identified by MacGregor et al. (2016))  
525 all occur where the ice surface speed is low. This includes the north-central ice-divide (Fig. 7(d)), the region east of GRIP (Fig. 7(e)), and the region west of Kangerlussuaq (Fig. 7(f)). The regions of agreement/disagreement are, perhaps, unsurprising, since three of the four methods employed by MacGregor et al. (2016) - ice-velocity, surface texture and radiostratigraphy - associate basal thaw with present (or past) ice sheet motion. In general, a thawed bed is a necessary (but not sufficient)  
530 condition for appreciable basal motion, and there is likely to be a subset of thawed regions where basal motion is negligible. This subset naturally incorporates water/thaw near the ice divides (since driving stress is low), and in the eastern ice sheet (since ice flow is topographically constrained).

Another key difference between the water predictions in this study and the thaw predictions by MacGregor et al. (2016) is that their study employed techniques better tuned to identify continu-  
535 ous regions of basal thaw, whereas the basal water predictions are localised. This provides another means to reconcile regions of disagreement, since in some instances the basal water predictions may correspond to localised patches above the PMP in an otherwise frozen region. A final explanation for discrepancies, is that the model temperature fields included in the basal thermal state synthesis were often tuned around knowledge of GHF at the time (i.e. Shapiro and Ritzwoller (2004); Fox Maule  
540 et al. (2009)).

There is no evidence for basal water at the location of the temperature boreholes, which, based upon the resolution of our method, corresponds to within 5 km. Since high reflectivity variability is not necessary for thaw this is consistent with both frozen and thawed borehole temperatures. Water is, however, observed fairly close to two frozen boreholes:  $\sim 10$  km south of GRIP and  $\sim 7$  km  
545 northeast of Camp Century (Fig. 7). At GRIP this is less surprising, since the basal temperature is 6 degrees below the PMP (Dahl-Jensen et al., 1998; MacGregor et al., 2016) and GHF is likely to be elevated in this region (Fig. 8). The basal water predictions near Camp Century are more surprising, since in the late 1960s basal temperatures were measured to be 11.8 degrees below the PMP (Weertman, 1968; MacGregor et al., 2016). One possible explanation, which was recently

550 invoked to explain the presence of a lake less than 10 km from South Pole (where the bed is frozen),  
is that the basal water is yet to reach thermal equilibrium (Beem et al., 2017). Another possible  
explanation is that the presence of hypersaline water could result in a depression of the PMP. This  
situation arises at Lake Vida in East Antarctica (where liquid water exists at  $-13\text{ }^{\circ}\text{C}$ ) (Murray et al.,  
2012) and at Devon Ice Cap in the Canadian Arctic (Rutishauser et al., 2018).

## 555 **4.2 Basal water and geothermal heat flux**

The comparison between basal water and the different GHF models in Fig. 8 (Shapiro and Ritz-  
woller, 2004; Fox Maule et al., 2009; Martos et al., 2018) demonstrates greatest consistency with  
the distribution by Martos et al. (2018). Notably there is a pronounced spatial correlation between  
elevated GHF and the new predictions of basal thaw in the northern ice sheet. A recent machine  
560 learning-derived map for GHF beneath Greenland by (Rezvanbehbahani et al., 2017) is also consis-  
tent with there being extensive basal thaw in this region. However, establishing definitive attribution  
of regions of the basal melt to GHF forcing (rather than frictional and strain heating, low advection  
from colder ice above, and/or surface melt water storage) will require integration with thermome-  
chanical ice-sheet models. The basal water predictions could also be used as a constraint in a wide  
565 variety of other numerical modelling contexts. For example, experiments with 3D models to recon-  
struct the full ice temperature history over the last glacial cycle(s) can constrain the minimum GHF  
required to produce basal melting at the predicted basal water locations (Huybrechts, 1996). Other  
studies include investigating the sensitivity of ice-sheet dynamics to the thermal boundary condi-  
tion (Seroussi et al., 2013) or basal lubrication (Shannon et al., 2013), and thermal models of the  
570 underlying lithosphere (Rogozhina et al., 2016).

Recent analyses imply that much of the spatial variation in GHF beneath the northern GrIS can  
be explained by Greenland's passage over the Iceland mantle plume between roughly 35 and 80  
million years ago (Rogozhina et al., 2016; Martos et al., 2018). The magnetic GHF map in Fig.  
8(c), alongside gravity data (Bouger anomalies), was recently used to by Martos et al. (2018), to  
575 infer the most likely passage of the hotspot track. The most likely predicted path (corresponding to  
going forwards in geological time) follows the quasi-linear region of elevated GHF in Fig. 8(c) from  
Petermann glacier to NorthGRIP, and follows a path previously anticipated by Forsyth et al. (1986).  
The spatial correlation between elevated GHF and the quasi-linear basal water 'corridor' provides  
an additional source of evidence for the predicted path.

## 580 **4.3 Basal water, bed topography and subglacial flow paths**

There is growing evidence that much of the present day subglacial flow path network beneath the  
GrIS is paleofluvial in origin. This includes the dendritic flow path networks in the Jakobshavn  
(Cooper et al., 2016) and Humboldt catchments (Livingstone et al., 2017), along with the prominent  
'mega-canyon' feature which extends from the NorthGRIP region in the south to Petermann glacier

585 in the north (Bamber et al., 2013b). The comparison between the predicted flow paths and basal  
water in Fig. 9 enables a revised assessment of the hydrological flow paths that are likely to be  
utilised in the contemporary ice sheet. For example the flow routing analysis demonstrates that basal  
water originating in the Petermann catchment is likely to route through sections of the canyon toward  
the ice-sheet margins. Fig. 9(b) supports this hypothesis, since ~~the~~ there is evidence for basal water  
590 along the majority of the canyon. However, it is important to stress that more rigorously assessing  
hydrological connectivity will require incorporation of DEM uncertainty when performing the flow  
routing (e.g. Schroeder et al. (2014)) and use of a coupled hydrological ice flow model (e.g. Le Brocq  
et al. (2009)).

#### 4.4 Basal water and ice-sheet motion

595 Both observational (e.g. Moon et al. (2014); Tedstone et al. (2013)) and theoretical studies (e.g.  
Creys and Schoof (2009); Schoof (2010)) point toward a complex spatio-temporal relationship be-  
tween basal water and ice surface speed in fast flowing regions of the ice sheet. This ultimately  
depends upon the details of how the subglacial drainage system responds to surface melt water. **It is  
therefore essential to re-emphasise that the basal water predictions generally correspond to the win-  
600 ter storage (pre- surface melt) configuration and may, therefore, be of limited utility in understanding  
spatio-temporal patterns related to ice dynamics.**

In addition to basal water and temperature, spatial variation in the underlying geology and lithol-  
ogy of the GrIS (notably, presence or absence of deformable sediment) will also influence ice-sheet  
motion. It is widely anticipated that much of the interior of the ice sheet is underlain by hard pre-  
605 Cambrian rocks, with more limited sedimentary deposits toward the margins (Dawes, 2009; Hen-  
riksen, 2008), and in the NEGIS drainage basin (Christianson et al., 2014). It is therefore entirely  
plausible that much of the basal water predicted in the interior lies upon a hard undeformable bed  
(particularly in the context of the igneous rock that would be associated with the geological remnants  
of the Iceland hotspot track) and therefore experiences little motion due to bed deformation.

#### 610 4.5 Comparison with past RES analyses of basal water and disrupted radiostratigraphy in Greenland

Despite acknowledged calibration issues, due to both variable radar system performance and spatial  
variation in attenuation, the bed-echo reflectivity analysis of 1990s PARCA RES data by Layberry  
and Bamber (2001) anticipated some of the water predictions in this study. This includes prior basal  
615 water predictions in the NEGIS onset region, and the upstream areas of Kangerlussuaq, Petermann  
and Humboldt glaciers.

There is a mixed agreement between the basal water predictions in this study and those from Os-  
wald and Gogineni (2008, 2012) who performed joint bed-echo reflectivity **and** scattering analysis of  
the 1990s PARCA data. In general, better agreement with our results occurs in smoother topographic

620 regions in the ice-sheet interior such as close to the NorthGRIP borehole. Since the effects of spatial bias due to attenuation uncertainty are lower in the interior of the ice sheet, this is where bed-echo reflectivity as a water diagnostic should be more robust. Additionally, the water detection method proposed by Oswald and Gogineni (2008, 2012) will generally not be able to discriminate water in many outlet glaciers and tributaries including Petermann and the northwestern margins (Jordan et al., 625 2017). This is because these regions tend to exhibit a diffuse scattering signature (associated with fine-scale roughness) whereas the method proposed by Oswald and Gogineni (2008, 2012) is specifically tuned to detect water bodies that exhibit a spatially continuous, (near-) specular scattering signature. By contrast, comparison between the water predictions in this study and the radar-derived bed roughness maps in Rippin (2013) and Jordan et al. (2017) demonstrate a lack of modulation 630 by bed roughness, with basal water present in rougher marginal regions and the generally smoother ice-sheet interior.

Basal units of disrupted radiostratigraphy are widely present in Greenland RES data sets (Bell et al., 2014; Wolovick et al., 2014; Bons et al., 2016; Dow et al., 2018). The features have been attributed to a supercooling ‘freeze-on’ process (Bell et al., 2014), ‘stick-slip’ mechanisms, (Wolovick 635 et al., 2014) and the rheological anisotropy of ice (Bons et al., 2016). In the fast flow initiation regions of some outlet glaciers (e.g. Petermann and the northern tributaries of NEGIS) these basal units are closely aligned with the basal water predictions.

#### 4.6 Limitations of bed-echo reflectivity variability as a RES technique to detect basal water

Bed-echo reflectivity variability provides a practical way to automate the detection of a subset of 640 basal water with high confidence at the ice-sheet scale. In particular, as a form of edge detector, the technique is well-tuned to detect finite water bodies with sharp horizontal gradients in water content. These attributes are thought likely to be common to basal water in the (likely hard-bedded) interior of the ice-sheet. It is, however, important to note that the approach will fail to identify basal water/wet regions with a homogeneous dielectric and reflective character. This includes the 645 centre of large subglacial lakes (based upon the resolution of our method lakes greater than 5 km in horizontal extent) and regions of more uniformly saturated subglacial till. Since all identified subglacial lakes in Greenland are < 5 km in horizontal extent (Palmer et al., 2013; Howat et al., 2015; Palmer et al., 2015; Willis et al., 2015) we believe that the former scenario is likely to be rare. However, extensive regions of saturated till that evade detection are likely to be present, particularly 650 beneath larger outlet glaciers. This interpretation is supported by comparison with the bed-echo reflectivity and basal water maps of the Petermann catchment in Chu et al. (2018). Specifically, there is a good agreement between the two water maps in the interior and fast flow initiation region, but this study fails to predict the basal water (likely to be wet sediment) in the main trunk of the outlet glacier. Therefore, if focusing on the catchment-scale subglacial hydrology of Greenland outlet 655 glaciers, or other glaciologically similar regions of Antarctica, a suite of existing RES techniques to

detect and characterise basal water (e.g. Peters et al. (2005); Jacobel et al. (2009); Schroeder et al. (2013); Young et al. (2016)) are better suited. Finally, it is important to add that part of the subglacial water budget is likely to comprise groundwater (Siegert et al., 2017), which would be practically undetectable by RES.

660 As is generally the case in RES analysis, certain simplifications were made in this study when interpreting bed-returned power. Notably, we did not account for: (i) power modulation due to birefringent propagation/ice fabric anisotropy (e.g. Matsuoka et al. (2012b)); (ii) rough surface transmission (e.g. Grima et al. (2014); Schroeder et al. (2016a)); (iii) scattering by near-surface water. The first of these mechanisms could potentially influence power variability near the ice-divides, since  
665 abrupt fabric transitions can be present in these regions (Martin et al., 2009; Drews et al., 2012). However, the zoom plots in these regions (e.g. Fig. 7(b,d), Fig. 9(d)) indicate that the water predictions occur over a multiple range of flight-track orientations relative to the ice divide (which should correlate with orientation of the principle dielectric axes). We therefore can discount ice fabric having a dominant influence upon the results. The second and third of these mechanisms will influence  
670 power variability primarily in faster flowing regions toward the margins, as these regions have higher surface roughness (e.g. MacGregor et al. (2016)) and surface melt (e.g. van de Wal et al. (2008)). The degree of surface-induced power variability depends upon both the surface permittivity and the roughness regime (relative to the radar wavelength) (Schroeder et al., 2016a). It is, however, important to note that we are likely to get the majority of ‘false positives’ (elevated power variability due to  
675 surface modulation) in regions where the bed is predicted to be thawed with high certainty by MacGregor et al. (2016). In turn, the central results in this study - the ‘new’ regions of basal water/thaw identified in Fig. 7 - are likely to be largely unaffected by surface modulation.

## 5 Summary and conclusions

This study placed a spatially comprehensive observational constraint upon the basal water distribution beneath the GrIS and, hence, regions of the bed at or above the PMP of ice. The distribution of  
680 basal water is influenced by, and has influence upon, multiple ice-sheet and subglacial properties and processes. Subsequently, with a focus upon ice-sheet scale behaviour, we performed an exploratory comparison with related data sets for the GrIS. This included an up-to-date synthesis for the basal thermal state (MacGregor et al., 2016), three different GHF model distributions (Shapiro and Ritzwoller, 2004; Fox Maule et al., 2009; Martos et al., 2018), bed topography (Morlighem et al., 2017)  
685 and predicted subglacial flow paths, and ice surface speed (Joughin et al., 2010, 2016).

Central to the methods in the study was the use of bed-echo reflectivity variability (rather than bed-echo reflectivity) as a RES diagnostic for basal water. Our use of this diagnostic (a form of edge detector) was motivated by its insensitivity to radar attenuation at the ice-sheet scale, and the  
690 pragmatic advantages when performing data combination for multiple RES field campaigns. The

reflectivity variability diagnostic is, however, only able to detect wet to dry (or wet to frozen) transitions in bed material and is a sufficient (but not necessary) condition for basal water. It will therefore need to be combined with other information to fully map basal water extent and classify basal water bodies.

695 There was much agreement between the basal water distribution and the thawed marginal regions predicted by MacGregor et al. (2016). However, we identified regions of basal water/thaw in the interior of the ice sheet that were previously classified as likely to be frozen. The most extensive ‘new’ region of predicted thaw is a quasi-linear ‘corridor’ feature which extends from NorthGRIP in the south to Petermann in the north. This feature, and the majority of basal water in the northern interior, spatially correlate with elevated GHF inferred from magnetic data by Martos et al. (2018).  
700

The comparison with bed topography (Morlighem et al., 2017) and predicted flow paths, demonstrated good overall agreement between the basal water storage locations and the geometric constraints imposed by the hydrological pressure potential. However, many of the basal water predictions in the ice-sheet interior occur where ice surface speed (and hence basal motion) is negligible.

705 One plausible explanation is that much of the interior lies upon a hard and undeformable bed. Future investigation of basal control upon GrIS dynamics, should integrate information about basal water and the basal thermal state with better constraints upon bed lithology and geology.



## Appendix A: Cross-over analysis for bed-echo reflectivity variability

To assess the internal consistency of the bed-echo reflectivity variability data, Fig. 3(c), we carried  
710 out a cross-over analysis at flight-track intersections (defined, as bin centres separated by  $< 2.5$  km in  
correspondence with 5 km window size). In this analysis we decomposed the data by the radar system  
categories in Fig. 1(a), using MCoRDS v2 (the most spatially extensive and recent measurements)  
as a benchmark. We also discounted adjacent postings from the same flight-track.

The cross-over statistics are shown in Table 2, and demonstrate standard deviations that range  
715 from 1.13 dB (MCoRDS v2 - MCoRDS v2) to 1.69 dB (MCoRDS v2 - MCRDS). The relatively  
high standard deviation for MCoRDS v2 - MCRDS is likely to arise because the MCRDS cover-  
age is almost exclusively in outlet glacier regions where power variability due to surface roughness  
and crevassing will be higher. There is no significant bias due to radar centre frequency (MCoRDS  
v2 and MCoRDS are for 195 MHz, whereas MCRDS and ACORDS are for 150 MHz). However,  
720 MCoRDS does have a small negative bias. Since the underlying mechanism for this is unclear we  
do not empirically correct the data. We note that this is a conservative approach (since  $\sigma_{[R]}$  is under-  
estimated), and is logically consistent with  $\sigma_{[R]}$  being used as a sufficient condition for basal water  
(i.e. basal water may be present that is not above the prescribed variability threshold).

The cross-over standard deviation values for  $\sigma_{[R]}$  in Table 2 should not be interpreted as standard  
725 errors (and are likely significant overestimates) as the flight-track windows do not necessarily sample  
the same region of the glacier bed. This contrasts with performing cross-over analysis of bed-echo  
power/reflectivity where the Fresnel zone (length scale  $\sim 100$  m) defines a spatial overlap. As a  
form of edge detector, the purpose of the  $\sigma_{[R]}$  metric is to identify a signal attributable to basal water  
within a 5 km region (rather than to coarse-grain an average value). Additionally, the along-track  
730 data in Fig. 2 shows that  $\sigma_{[R]}$  can rapidly fluctuate at a 5 km length scale. We therefore point toward  
the high degree of spatial structure for  $\sigma_{[R]}$  in Fig. 3(c), and the water predictions in Figs. 7 - 10 as  
evidence for the robustness of the approach.

## Data availability

The CSARP L1B RES data are available from CReSIS at <https://data.cresis.ku.edu/data/rds/> and are documented in Paden (2015). The profile in Fig. 2(a) is from data segment 2012050804 from the 2012 P3 season and the profile in Fig. 2(b) is from data segment 2014051601 from the 2014 P3 season. The Greenland basal thermal state synthesis (MacGregor et al., 2016), ice thickness and topography data sets (BedMachine V3) (Morlighem et al., 2017), and ice surface speed (Joughin et al., 2016), are archived by NSIDC at <http://dx.doi.org/10.5067/R4MWDWWUWQF9>, <https://nsidc.org/data/idbmg4> and <https://nsidc.org/data/NSIDC-0670/versions/1> respectively. The GHF maps by Fox Maule et al. (2009) and Shapiro and Ritzwoller (2004) and available at <http://www.dmi.dk/dmi/dkc09-09.pdf> and [http://ciei.colorado.edu/~nshapiro/MODEL/ASC\\_VERSION/index.html](http://ciei.colorado.edu/~nshapiro/MODEL/ASC_VERSION/index.html) respectively. *We have contacted NSIDC about archiving of the water data set. We hope to provide the link for this, along with the PANGEA link for the heat flux map by Martos et al. 2018, in the final TC paper. Pending review;*

~~the basal water distribution will be provided as csv files on both a season-by-season basis and for the full (13-season) data set. The data columns will correspond to: (A) latitude, (B) longitude, (C) water binary value, at a 1 km along-track posting. The water binary value will correspond to: 1  $\equiv$  a 5 km bin with water ( $\sigma_{[R]}$  > 6 dB, robust to a  $\pm 20\%$  perturbation in  $\langle N \rangle$ ), 0  $\equiv$  a 5 km bin with a majority of good quality bed echoes but no water, NaN  $\equiv$  a 5 km bin with a majority of poor quality bed echoes (interpreted as no coverage). A GIS shape file for the water predictions and radar flight tracks will also be provided.~~

## Author contributions

The study was initiated by JLB and co-advised by MJS as part of the Basal Properties of Greenland Project. TMJ wrote the paper, analysed the radar data, and, with advice from DMS, developed the water detection method. CNW and MAC developed the GIS mapping environment and modelled the subglacial flow paths. JDP and DMS both advised on radar data processing. YMM contributed the new heat flux distribution and advised on the interpretation of the heat flux models. PH contributed the GISM temperature field and advised on ice-sheet thermodynamics. All authors commented on the paper draft and contributed to the interpretation of the science.

## 760 Competing interests

JLB is an advisory editor of The Cryosphere. The authors declare that they have no conflict of interests.

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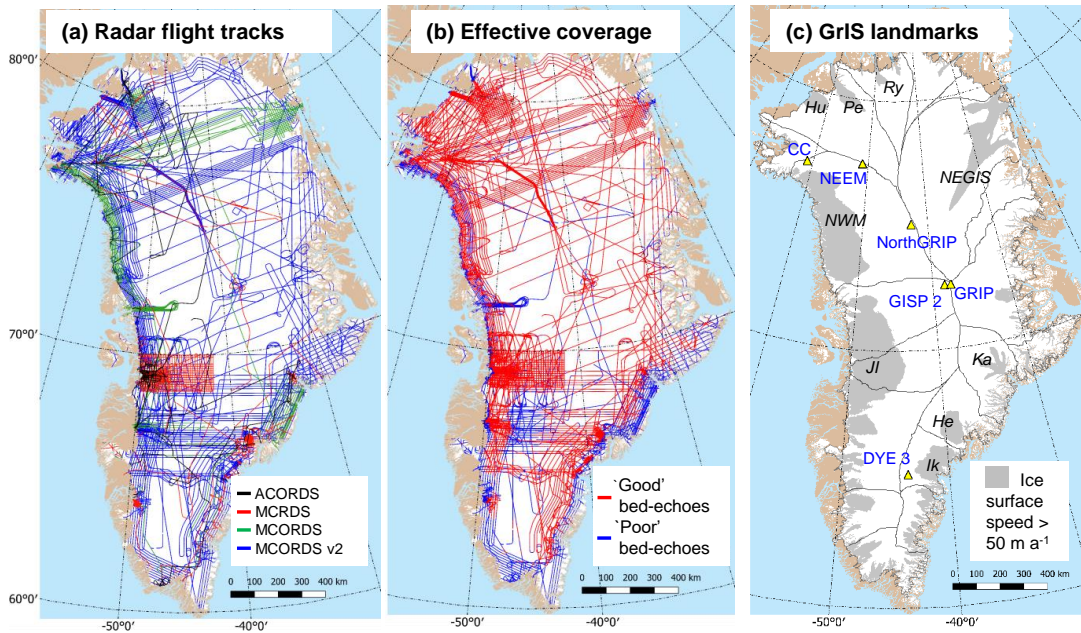
*We have shortened the label of the 2nd column to make this suitable for a single column width.*

**Table 1.** Dielectric and reflective properties of subglacial materials based upon a compilation of past values by Bogorodsky et al. (1983); Martinez et al. (2001); Peters et al. (2005). The bulk values take into account typical ranges of saturation and porosity for the dielectric mixing of water and ice with the background material. The relative dielectric permittivity of ice is 3.15, which means that ‘dry’ (just the background dielectric) or ‘frozen’ (a mixture of the background dielectric with ice) produce a similar range for  $[R]$ .

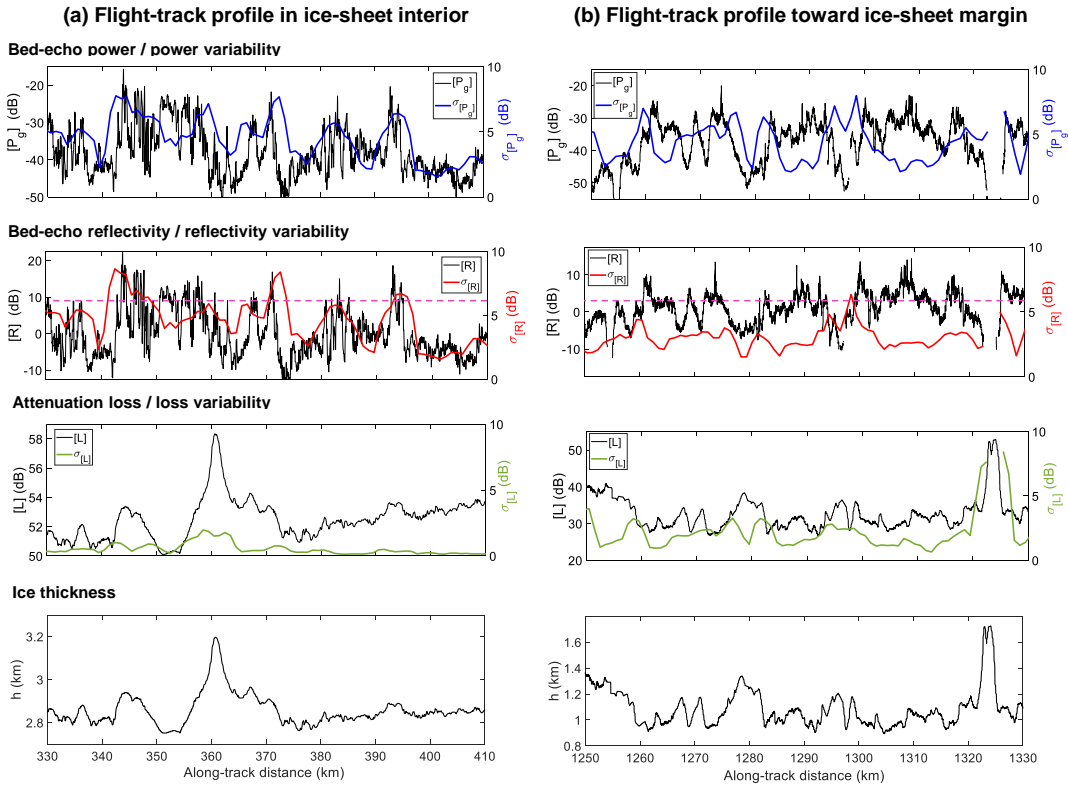
Bed material	Permittivity $\epsilon$	Reflectivity $[R]$ (dB)
Ground water	80	-2
Wet till	10 to 30	-11 to -6
Wet sandstone	5 to 10	-19 to -11
Dry/frozen granite	5	-19
Dry/frozen limestone	4 to 7	-26 to -14
Dry/frozen till	2 to 6	negligible to -19
Dry/frozen sandstone	2 to 3	-37 to -16

**Table 2.** Cross-over statistics for  $\sigma_{[R]}$  for the different radar instrument classes in Fig. 1(a).

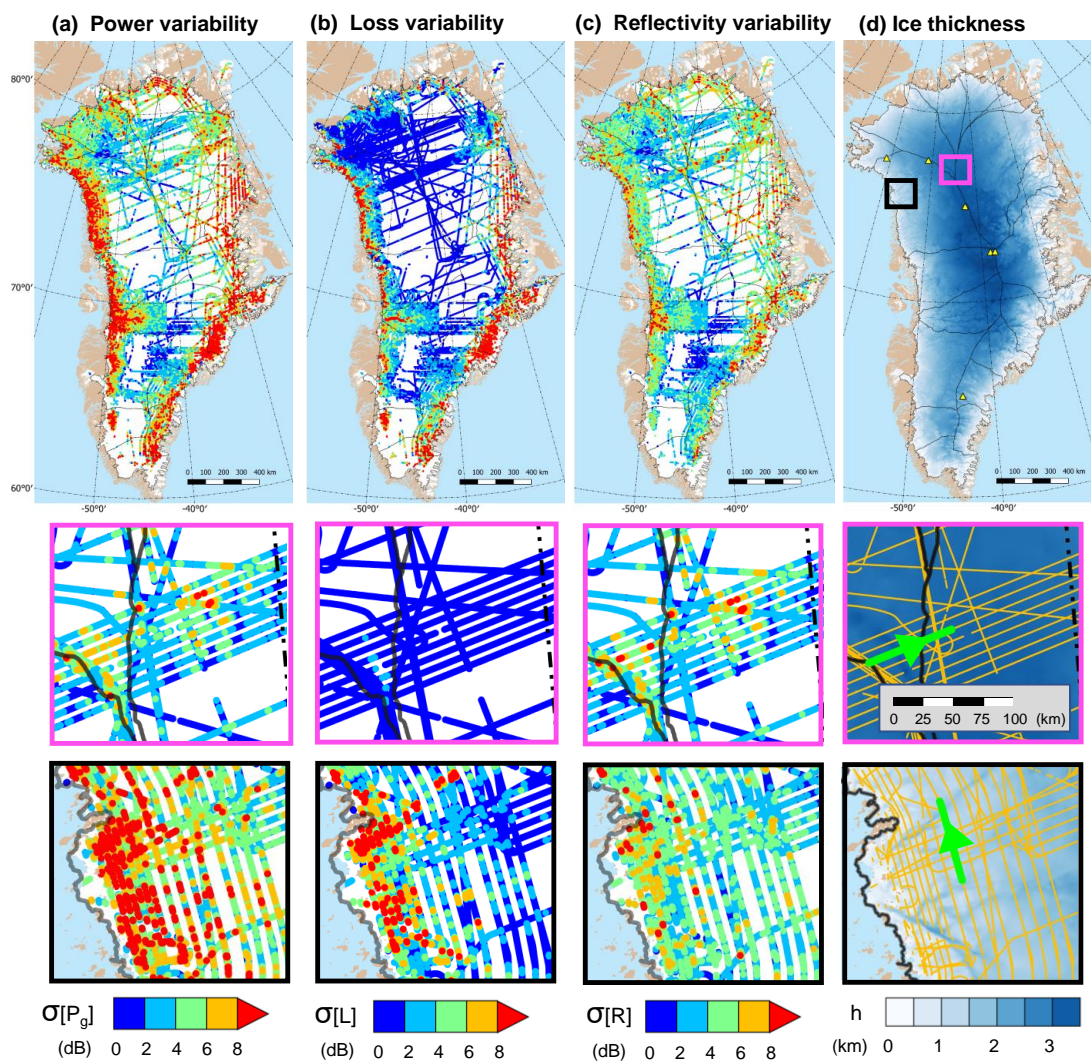
Data comparison	Mean (dB)	SD (dB)
MCoRDs v2 - MCoRDs v2	N/A	1.13
MCoRDs - MCoRDs v2	-0.43	1.18
MCRDS - MCoRDs v2	-0.07	1.69
ACORDS - MCoRDs v2	0.00	1.24



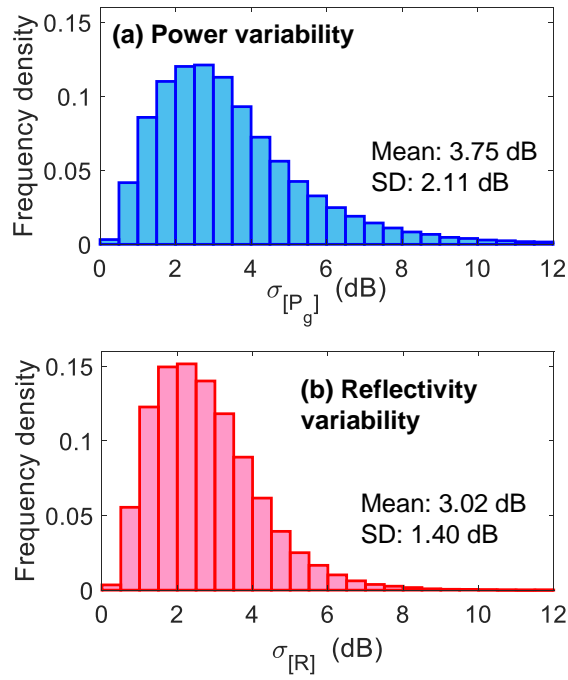
**Figure 1.** (a) Ice-penetrating radar flight-tracks for different CReSIS radar systems. (b) Effective coverage for ‘good quality’ radar bed-echoes (corresponding to peak power 10 dB above noise floor). (c) Summary of key GrIS landmarks: temperature boreholes, major drainage basin boundaries (Zwally et al., 2012), and major regions of fast flow identified from ice surface speed (Joughin et al., 2010, 2016). Abbreviations in (c) correspond to: Camp Century (CC), Humboldt (*Hu*), Petermann (*Pe*), Ryder (*Ry*), North East Greenland Ice Stream (*NEGIS*), northwestern margins (*NWM*), Jakobshavn Isbrae (*JI*), Kangerlussuaq (*Ka*), Helheim (*He*) and Ikertivaq (*Ik*). The projection is a Polar Stereographic North (70°N, 45°W) and is used in all future plots.



**Figure 2.** Example flight-track profiles for: bed-echo power and variability,  $[P_g]$  and  $\sigma_{[P_g]}$ , bed-echo reflectivity and variability,  $[R]$  and  $\sigma_{[R]}$ , attenuation loss and variability,  $[L]$ ,  $\sigma_{[L]}$ , ice thickness,  $h$ . Example (a), left of figure, is from the north-central interior of the ice sheet and (b), right of figure, is from the northwestern margins (locations both shown in Fig. 3). The threshold that is later used for water detection,  $\sigma_{[R]} > 6$  dB, is indicated by the dashed pink line and applies to the right axis. The values for  $[R]$  are relative with zero mean. The variability measures are all calculated at a 5 km length scale with 1 km posting.

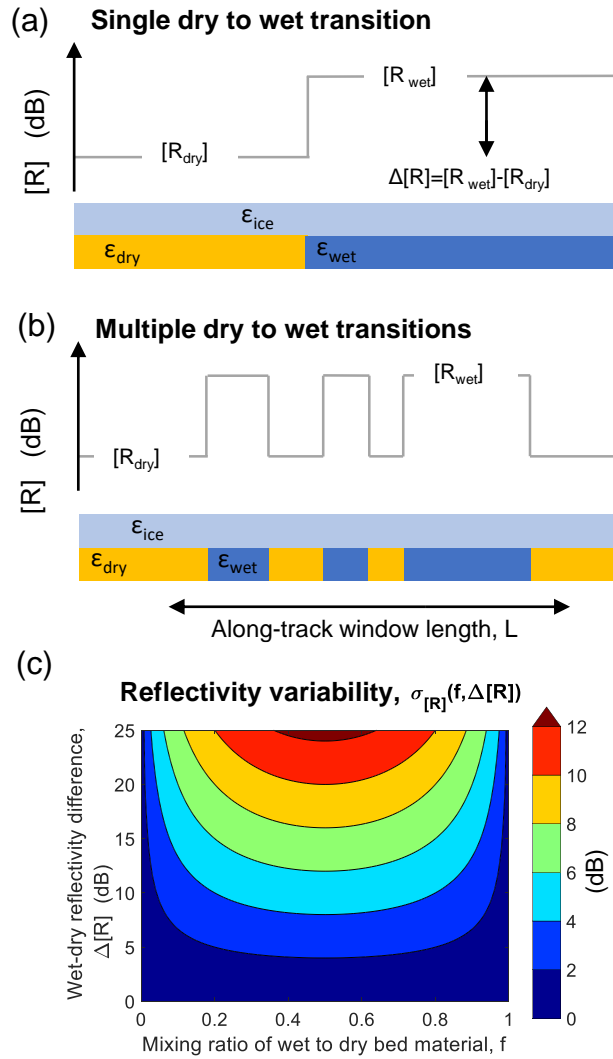


**Figure 3.** Spatial distributions for: (a) bed-echo power variability  $\sigma_{[P_g]}$ , (b) attenuation loss variability  $\sigma_{[L]}$ , (c) bed-echo reflectivity variability  $\sigma_{[R]}$ , (d) ice thickness,  $h$ . Zoom plots with flight-track data at true buffer size (5 km) are shown for the north-central ice sheet (pink bounding box, containing profile in Fig. 2(a)) and northwestern margins (black bounding box, containing profile in Fig. 2(b)). The profiles are indicated in bold green in the ice thickness zoom plots. In plots (a)-(c) higher variability data is stacked on top of lower variability data, which acts to emphasise higher variability. The zoom plots are all have the same scale ( $\times 8$  the resolution of the ice-sheet scale plots).

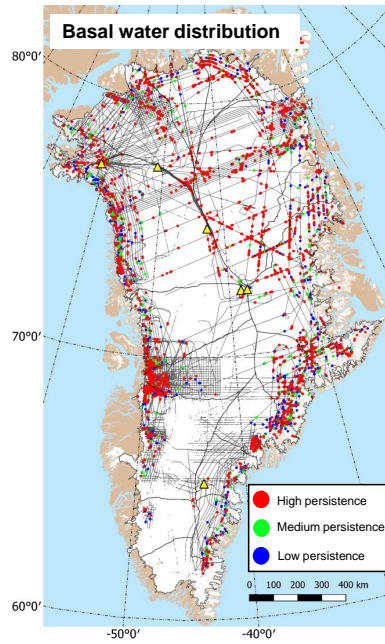


**Figure 4.** Frequency distributions for: (a) bed-echo power variability,  $\sigma_{[P_g]}$  (corresponding to Fig. 3(a)); (b) bed-echo reflectivity variability,  $\sigma_{[R]}$  (corresponding to Fig. 3(c)). Later in the study  $\sigma_{[R]} > 6$  dB used as threshold criteria for diagnosing the presence of basal water.

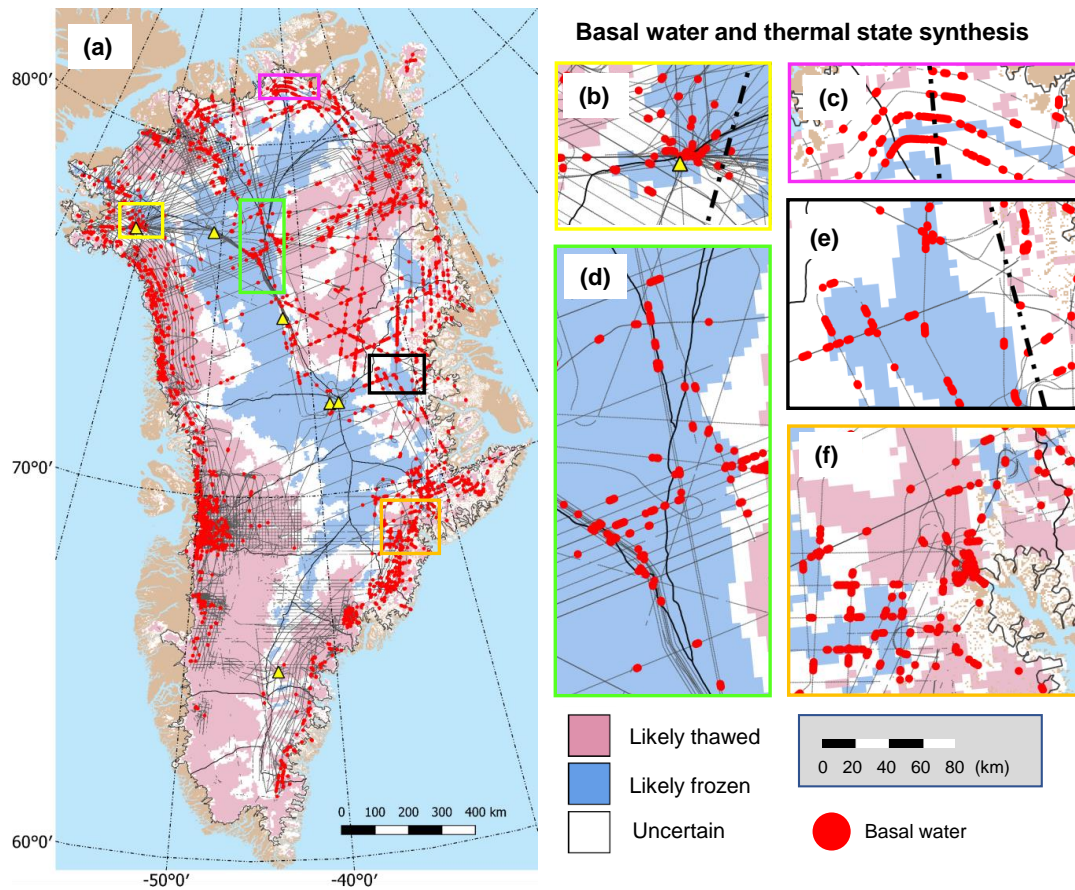




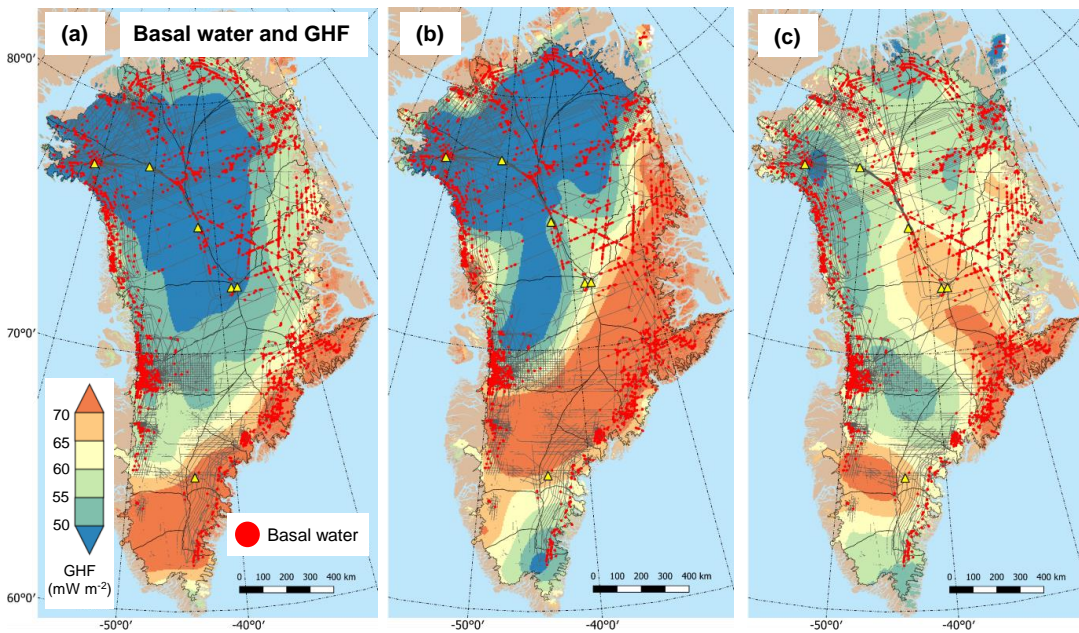
**Figure 5.** Interpretation of bed-echo reflectivity variability,  $\sigma_{[R]}$ , as a diagnostic for basal water. (a,b) Schematics of the two-state dielectric model for single and multiple along-track transitions in dry to wet bed material. Both scenarios are identically parameterised by the wet-dry mixing ratio  $f$  (visually, the fraction of blue to yellow) and wet-dry reflectivity difference,  $\Delta[R] = [R_{wet}] - [R_{dry}]$ . (c) Phase-space plot for  $\sigma_{[R]}$  as function of  $f$  and  $\Delta[R]$ .  $\sigma_{[R]} > 6$  dB is used as a threshold for positive discrimination of basal water (corresponding to green, red and yellow regions).



**Figure 6.** Basal water distribution and robustness to perturbations in the attenuation rate estimate,  $\langle N \rangle$ . The original predictions ( $\sigma_{[R]} > 6$  dB) are represented by all three colours. ‘Persistent’ water predictions ( $\sigma_{[R]} > 6$  dB for  $\pm 20\%$  and  $\pm 50\%$  perturbations to  $\langle N \rangle$ ) are indicated by the subset of green and red points, and the subset of red points respectively. The subset of red and green points is used in the rest of the paper.

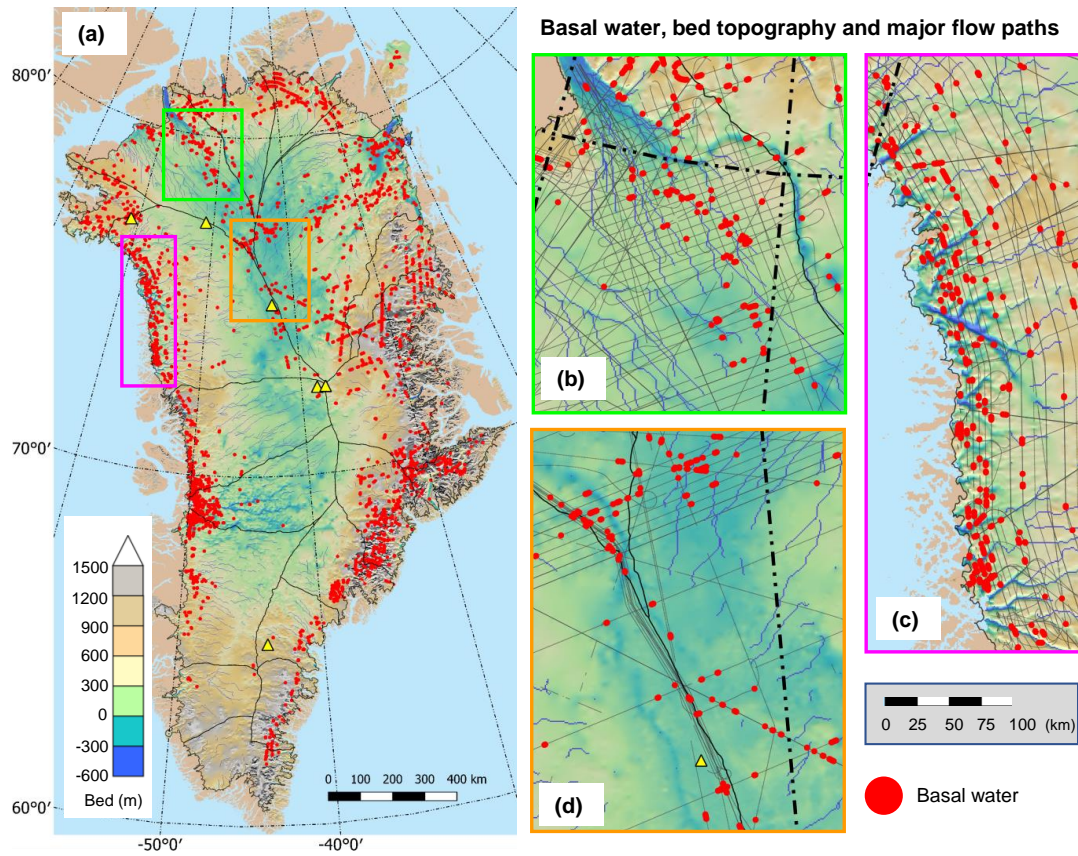


**Figure 7.** Comparison between basal water distribution and basal thermal state synthesis by MacGregor et al. (2016). (a) Ice-sheet scale. Major regions of disagreement (water in likely frozen regions) are highlighted in the zoom plots. (b) Camp Century. (c) Far north. (d) North-central ice sheet. (e) East of GRIP. (f) Around Kangerlussuaq. The zoom plots all have the same scale ( $\times 5$  the resolution of (a)).

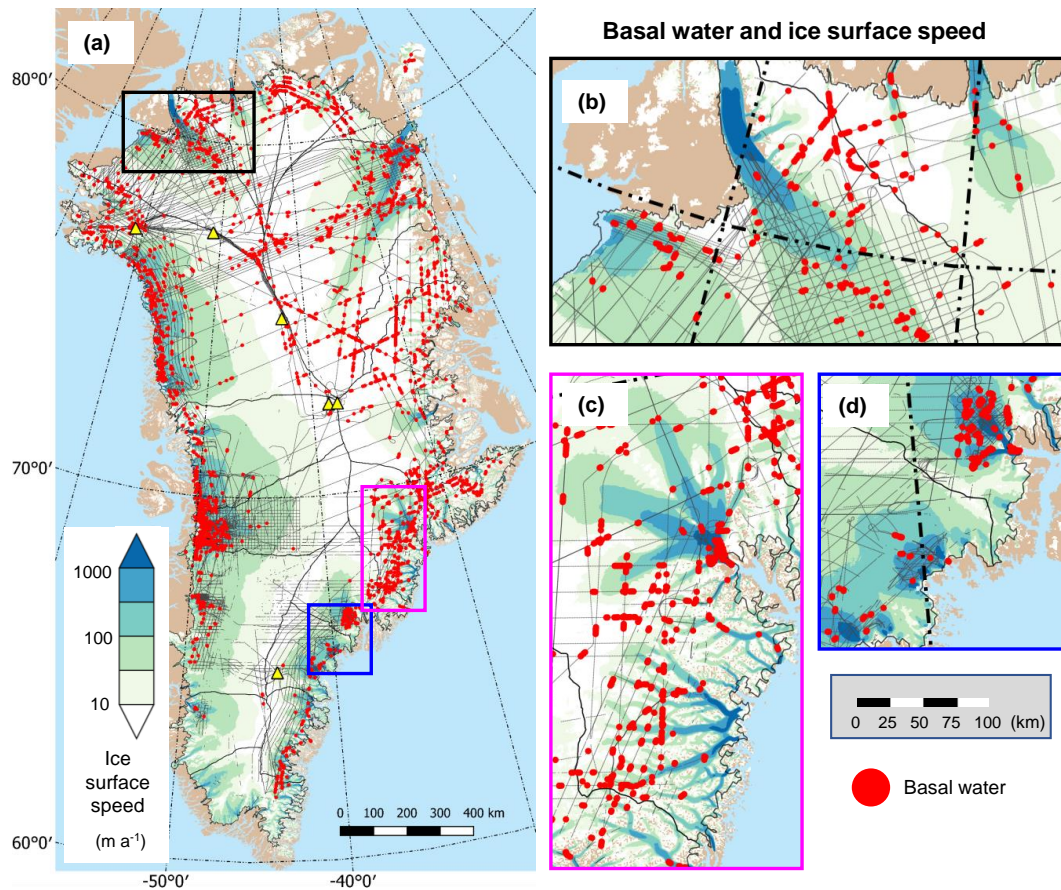


**Figure 8.** Comparison between basal water distribution and geothermal heat flux (GHF) models. (a) Seismic GHF model by Shapiro and Ritzwoller (2004). (b) Magnetic GHF model by Fox Maule et al. (2009) using satellite data. (c) Magnetic GHF model by Martos et al. (2018) derived from spectral methods using airborne data. The colour bar scale is the same in all plots and is truncated to emphasise the spatial variation in plot (c).





**Figure 9.** Comparison between basal water distribution, bed topography (Morlighem et al., 2017), and major subglacial flow paths (blue lines). (a) Ice-sheet scale. (b) Petermann catchment. (c) Northwestern margins. (d) North-central ice sheet. To improve clarity the radar flight-tracks are removed from (a) and a hillshade is applied to the bed topography. The zoom plots are all have the same scale ( $\times 4$  the resolution of (a)).



**Figure 10.** Comparison between basal water distribution and ice surface speed (Joughin et al., 2010, 2016) (logarithmic-scale). (a) Ice-sheet scale. (b) Humboldt, Petermann and Ryder. (c) Kangerlussuaq and region to south. (d) Helheim (north of plot) and Ikertivaq (southwest of plot). The zoom plots are all have the same scale ( $\times 4$  the resolution of (a).)