Response to referee 1 comments.

Below we detail how we have responded to the referees' comments. In addition we provide, as a supplement a highlighted copy of the revised text to show where changes have been made. We are very grateful to the referees for their thorough, careful and constructive comments and suggestions.

General comments

A general comment relates to the subglacial water routing. Whilst it is good to have a demonstration of the application of the new data, the way it is implemented in this paper seems like a bit of an add on. The routes are presented in a very broad scale manner with little detail, when one of the key benefits of the paper is to provide an improved resolution and amount of detail.. . .

We agree with this (echoed also by referee 3) and on further consideration of the balance of the paper we have removed the section on hydraulic potential and water routing as it is not in the scope and focus of the rest of the paper.

Another general comment relates to the application of the two distance/error relation- ships to the rest of the ice sheet based on the 2000 m threshold. This requires a big leap of faith that all other outlets have a similar spatial error variability to Petermann Glacier. I am not convinced that this is the case. I would be more convinced if the exercise was carried out for at least one other outlet to demonstrate that this was the case.

This is a good point and we agree with the comment. We were aware that the scale of basal topography varied from the margin to the interior and that this affected the interpolation error depending on the wavelength and amplitude of the undulations and wished to find a way to account for this. Having now examined the basal topography in detail we note that a 2000 m elevation threshold, based on the ice surface, is not the most effective means of doing this. First, the ice surface elevation has little direct relationship to basal roughness and second, as pointed out, Petermann Glacier may well not be representative of other marginal areas.

As a consequence, we have revised the error model based on properties of the bed and this section of the m/s has been updated. We calculated basal roughness and separated the bed into low/high roughness areas and fitted exponential curves to the interpolation error for these two zones, which now reflect basal roughness in a quantitative way. Since submission of the paper OIB 2012 data have become available and we have added these data to the revised grids. We would have had to re-estimate the error map anyway having done this. A new set of figures shows the revised approach, which we believe to be more rigorous and more physically based although the resulting error map is more spatially variable as a consequence of these changes.

Specific comments

Author List: Dowdeswell is spelt incorrectly.
Fixed.

p4830, line 10: This sentence is slightly confusing, using "ice shelf thickness" and "floating tongue" for the same thing, please reword.
Reworded.

p 4830, line 19: "changes" - in what? please clarify with further background
Clarified.

p4831, line 1: Give a reference for ice velocity being proportional to the fourth power of ice thickness, otherwise this is a bit of an unqualified statement.

Reference added.

p4831, line 2: What do you mean by estimated velocities? Modelled velocities?
Yes, this should have been modelled.

p4831, line 6: what other reasons? This is a bit vague, either mention them or remove these words.
IPR campaigns have been undertaken for numerous reasons including surveying potential drilling sites, identifying subglacial lakes, basal conditions such as melting, and for particle physics experiments. We felt that adding all of these (and more) was a bit cumbersome and long winded and, more importantly, unnecessary.

p4831, line 16: give more details of what you mean by changes in "boundary conditions"
Added details.

p4832, line 21: changed how?
Rephrased

p4833, line 2: vertical accuracy of which surface?
This was ambiguous. Rephrased.

p4833, line 19, sentence starting "Several ..." is not clear, you interchange "outlets" and "glaciers" – be consistent, probably best to use "outlet glacier" each time for consistency. Use "unsurveyed outlet" rather than "new outlet". Change "flown" to "re-flown" for clarity.
Made changes as suggested.

p4834, line 14: do you mean over outlet glaciers or the whole of SW Greenland?
We mean near the margin. Have reworded to make clear.

p4834, line 4834: Change to "It has been flown..." – sounds like the radar is flying itself otherwise!
Agreed

p4834, line 24: Young et al., 2011 is probably a better reference here than Wright et al., 2012.
Agreed. Replaced.

p4835, line 25: You could add coupled ice-ocean modelling as an application that requires bathymetry too.
Done.

p4836, line 4: add (IBCAO) acronym which you use later. p4836, line 25: Define KMS.
Done.

p4837, line 3: add "reprojected and resampled onto the..."
Done.

p4837, line 19: this is the first use of the word "postings" and you sporadically use this later on the text. Is it necessary to use this word – it won't be an overly familiar word to most readers – can you just say 5 km resolution instead?

Posting has a particular meaning in spatial modelling and is quite different from resolution, which is why we make the distinction. A grid that is 5 km resolution is not the same as one that has 5 km posting. We have clarified this in the text.

p4837, line 19: Why did you use a 3 standard deviation filter – was it from trial and error? We are aware that, for a Gaussian distribution, 0.2% of observations will lie outside 3 sigma but, in the presence of a distribution that contains outliers, this is a relatively effective way of identifying them.

p4838, line 7: Why was 2000 m chosen as the threshold? Again, was this from a visual analysis or from some quantitative analysis? It would be useful to have the contour plotted on fig 1 so the reader can be more aware of the changeover.
This is now redundant as the whole ice sheet has now been interpolated at the same resolution. This is because track density is not a function of surface elevation and neither is bed roughness (as discussed above WRT errors). Although this means a larger proportion of grid points are interpolated values rather than measured, we provide a data set that indicates which are which.

p4838, line 19: rephrase the " first 100 km of each variogram" to include something about this referring to the separation distance.
Done

p4838, line 24: Why did you choose 2.5 km? It seems a little odd that this number is not a multiple of 1 km.
Now redundant, see previous reply.

p4839, lines 5-10: Does this minimum thickness introduce discontinuities anywhere around the margin? You are artificially lowering the bed near the margin, potentially making it up to 50 m lower than the neighbouring ice free topography, which could introduce an unrealistic overdeepening.
This is a good point. We have not found any artificial overdeepening and it is important to include this step because we are calculating bed elevation from the surface minus thickness. If thickness is zero, because there are no data, then the bed elev will be artificially high. It is important to note here that 50 m is not the thickness at the ice front but the mean thickness of the 1 km marginal grid cell, which we found to be a reasonable estimate based on marginal grid cells with observed thickness data in them.

p4839, ice shelves section: This method not completely clear on the first read, it could do with some rewording. e.g. line 13: Change to something like : "... was found by combining all GIMP, ATM surface elevation measurements with surface elevation derived from airborne ice thicknesses measurements using a hydrostatic equilibrium assumption", also "most of each ice shelf area" is not very definitive. line 21: add "Ice shelf thickness was then calculated".
Have reworded as suggested to make clearer.

p4843, line 12: "...confirm its presence..." what is its referring to? the trough or the discontinuity?
The trough.

p4843, line 19: ...is now seen as..." is a bit colloquial, please reword.
Done.

p4843, line 25: remove "potential" this is confusing with hydraulic potential

This section has been removed altogether.

p4844, line 1: how did you derive the flow paths? More details needed.
See previous.

p4845, line 4: which "differences" – add more detail from the figure caption
Clarified.

p4856, Figure 3: A closer examination of Fig 3 shows that there are a few artefacts in the bathymetry in the area defined by the red box and also south of Helheim Glacier – is this inherited from IBCAO or a result of processing in this work?
It is a combination of the two. In these areas IBCAO contains no bathymetric data and does not agree with the bed elevation at the ice edge as determined by us. In fact, we find very large errors in IBCAO. For Petermann, the bathymetry appears to follow the ice shelf elevation and not the seafloor. The errors are 100s of metres. To avoid a discontinuity and to match the two where data exist we have adjusted IBCAO as explained in section 3.2.3. However, this did not include adjusting IBCAO beyond the lateral extent of the fjord/glacier mouth where the mismatch was found as there is no reasonable way to do this. In the revised data set, we have refined the correction to minimise discontinuities at the lateral boundaries of the feature. We have also stressed the point that the bathymetry is not reliable in areas where data are absent and should be treated with caution. Our view was that it was better to address the gross errors, in fjord mouths, where we found them and where we could, accepting that this introduces lateral gradients that are likely unrealistic but much smaller than the discontinuity that would have been present at the ice edge if we had not made these changes.

DataA new bed elevation dataset for Greenland

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Abstract

We present a new bed elevation dataset for Greenland derived from a combination of multiple airborne ice thickness surveys undertaken between the 1970s and 2012. Around 420,000 line kilometres of airborne data were used, with roughly 70% of this having been collected since the year 2000, when the last comprehensive compilation was undertaken. The airborne data were combined with satellite-derived elevations for non glaciated terrain to produce a consistent bed digital elevation model (DEM) over the entire island including across the glaciated/ice free boundary. The DEM was extended to the continental margin with the aid of bathymetric data, primarily from a compilation for the Arctic. Ice thickness was determined where an ice shelf exists from a combination of surface elevation and radar soundings. The across-track spacing between flight lines warranted interpolation at 1 km postings for significant sectors of the ice sheet. Grids of ice surface elevation, error estimates for the

30   DEM, ice thickness and data sampling density were also produced alongside a mask of
31   land/ocean/grounded ice/floating ice. Errors in bed elevation range from a minimum of ±10 m
32   to about ±300 m, as a function of distance from an observation and local topographic
33   variability. A comparison with the compilation published in 2001 highlights the improvement
34   in resolution afforded by the new data sets, particularly along the ice sheet margin, where ice
35   velocity is highest and changes in dynamics most marked. We estimate that the volume of ice
36   included in our land/ice mask would raise mean sea level by 7.36 metres, excluding any solid
37   earth effects that would take place during ice sheet decay.

38

1   Introduction

40   The bed elevation and ice thickness of the Greenland Ice Sheet are important boundary
41   conditions for numerical modelling. Surface ice velocity is roughly proportional to the fourth
42   power of ice thickness (Paterson, 1994) and errors in the latter can, therefore, introduce
43   substantial errors in modelled velocities for the present-day or future evolution of the ice
44   sheet. Bed and surface geometry can be used to determine hydraulic potential and, hence,
45   subglacial hydrological pathways (e.g. (Wright et al., 2008), while elucidating subglacial
46   topography can also provide insights on the origin and genesis of landforms at the bed
47   (e.g.(Young et al., 2011)). For these and other reasons, a large number of airborne field
48   campaigns have been deployed over Greenland during the last decade with a key objective of
49   obtaining ice thickness measurements. The last major compilation of these data, for deriving
50   bed elevations, was undertaken more than a decade ago (Bamber et al., 2001b) and did not,
51   therefore, include these more recent and extensive field campaigns. In particular, in recent
52   years there has been a focus on acquiring data over the fast-flowing outlet glaciers that fringe
53   the margins of the ice sheet and which are responsible, in part, for the recent acceleration in
54   mass loss observed on the Greenland Ice Sheet (van den Broeke et al., 2009;Howat et al.,
55   2011;Howat and Eddy, 2011). It is now well established that some of these marine-
56   terminating outlet glaciers can respond rapidly, and with large amplitude, to changes in the
57   force balance at the bed and/or calving front and, as a consequence, attention has been
58   focussed on modelling their past (Nick et al., 2009) and future behaviour (Nick et al., 2012).
59   For such applications, accurate basal geometry is critical for determining rates of grounding-
60   line migration and potential pinning points (Favier et al., 2012). Here, we take an approach
61   that is aimed at maximising the resolution and utility of the basal topography in these key,
62   marginal sectors of the ice sheet: areas that are often challenging for conventional ice

63 penetrating radar (IPR) systems. As a consequence, we employ a range of techniques for

64 improving the bed representation in these areas and interpolate the data at two different

65 resolutions, which are then merged into a single product. This product is intended to be

66 dynamic such that, as new data become available, they will be incorporated in new releases.

67 In addition, we intend to update the data sets with new methods that improve the interpolation

68 such as the use of a mass conservation model (Morlighem et al., 2011). This paper describes

69 the methods used, and presents the first release of the products.

70

71 ## 2   Datasets

72 ### 2.1   Airborne Datasets

73 In contrast to the previous compilation (Bamber et al., 2001b), bed elevation is the

74 interpolated parameter rather than ice thickness. The latter is derived from the difference

75 between the bed and surface elevation. This means that the bed elevation varies smoothly and

76 realistically across the land/glaciated boundary. To determine the bed elevation, data from a

77 number of airborne IPR missions have been collated, alongside new data sets for the

78 unglaciated bedrock. At present, we have collated ice thickness data from seven sources, each

79 of which is described below and detailed in Table 1. Figure 1 shows the spatial distribution of

80 the different data sources.

81 ### 2.1.1 Center for Remote Sensing of Ice Sheets (CReSIS)

82 The majority of the data included in this compilation are from a series of instruments

83 developed and flown by CReSIS at the University of Kansas (Gogineni et al., 2001). These

84 data were collected between 1993 and 2009 with the 1993-1999 data being identical to that

85 used previously (Bamber et al., 2001b). Between 1993 and 2002, data were acquired by the

86 Improved Coherent Radar Depth Sounder (ICoRDS). Between 2003 and 2005 the Advanced

87 Coherent Radar Depth Sounder (ACoRDS), between 2006 and 2009, the Multi-Channel

88 Radar Depth Sounder (MCRDS), and since 2010, the Multi-channel Coherent Radar Depth

89 Sounder (MCoRDS) were flown. Since 1999, around 65% more data were available

90 compared with the previous compilation and from 2006, more effort has been spent on

91 focused campaigns with dense grids over individual outlet glaciers that have changed the

92 most.

93

94   The MCoRDS instrument is operated over the 180 - 210 MHz frequency range with a 10 - 30
95   MHz adjustable bandwidth and multiple receivers developed for airborne sounding and
96   imaging of ice sheets. Multiple receivers allow cross-track surface clutter to be suppressed so
97   that relatively weak ice-bed echoes can be retrieved. MCoRDS has been flown on the NASA
98   P-3 and DC-8. Aircraft navigation was from kinematic GPS and a precision laser altimeter
99   was regularly mounted coincidentally. Using ice core sites, the vertical accuracy of the
100 thickness measurement was estimated to be ±10 m (Gogineni et al., 2001), but it is worse in
101 areas where the ice/bed interface is ambiguous or complex (e.g. in hilly terrain or when off-
102 nadir scattering obscures the desired basal return) and larger cross-over errors are commonly
103 seen in these regions. Data processed to contain location and ice thickness are available from
104 CReSIS and were used in the main. Geolocated radar echo strength profile images (often
105 known as radargrams) were used only to verify the removal of some data which appeared to
106 be unphysical.

107

108 In a region close to the outlet of Jakobshavn Isbrae, CReSIS produced a 125 m resolution bed
109 DEM. This includes all data collected in the region by them between 1997 and 2007. This
110 grid was used in place of the original CReSIS data as they reprocessed all the data in the
111 channel and collocated the data with coincidental lidar surface elevations. Additionally, they
112 included ASTER data in bare rock areas and fjord soundings to complement the airborne data.

113

114 Since 2010, the MCoRDS instrument was flown as part of the NASA Operation IceBridge
115 (OIB) programme (Studinger et al., 2010). OIB is designed to provide airborne data to fill the
116 gap between the end of the ICESat satellite record in 2009 and the launch of ICESat-2 which
117 is scheduled for 2016. MCoRDS operates on all flights where ice thickness measurements can
118 be retrieved. Several of the existing dense grids over rapidly changing outlet glaciers were re-
119 flown as well as tracks over previously unsurveyed glaciers. These data significantly improve
120 coverage particularly where narrow, fast flowing outlet glaciers were previously unsurveyed
121 around the north-west and south-west coasts of Greenland. The gridded datasets include
122 flights from the 2012 OIB season, which increases the total coverage by some 30%
123 (supplementary Fig S2).

124

2.1.2 Alfred-Wegener-Institut (AWI)

The AWI airborne instrument is capable of penetrating 4 km of ice with better than 50 m vertical accuracy and 3.25 m along-track resolution (Nixdorf et al., 1999). It operated at 150 MHz transmitting bursts of 60 ns and 600 ns duration. In earlier years a combination of GPS and inertial navigation was used and, since 1997, differential GPS was employed. Horizontal navigation errors are up to ±100 m. Data were collected in 1996-9, 2004 and 2010, operating out of the NGRIP camp site in central Greenland (Nixdorf and Goktas, 2001) and from coastal airstrips at Station Nord (Meyer et al., 1999) and Qaanaaq. This gives dense coverage in central Northern Greenland (Fig. 1).

2.1.3 Warm Ice Sounding Explorer (WISE)

WISE is an airborne sounder designed to measure ice thickness in areas of warm and fractured ice. It is based on the MARSIS planetary sounder used on Mars. It operates at a 2.5 MHz centre frequency with a monopole antenna with navigation using conventional GPS. It was operated by the NASA Jet Propulsion Laboratory (JPL) on an Air Greenland Twin Otter as an IPY deployment. Data were collected in 2008, 2009 and 2010 in marginal areas. WISE is the only radar that has returned bed echoes for some marginal areas of southwest Greenland.

2.1.4 Pathfinder Advanced Radar Ice Sounder (PARIS)

PARIS flew in 2009 and was operated by Johns Hopkins University on an Operation Ice Bridge flight (Raney, 2010). PARIS successfully demonstrated high altitude soundings with a delay Doppler design. The along track resolution of the data was 250 m with a vertical accuracy of 12.5 m.

2.1.5 High-Capability RadarSounder (HICARS)

HICARS is operated by the University of Texas Institute for Geophysics (Peters et al., 2007;Peters et al., 2005) and was operated in 2011 and 2012 as part of a joint UK/US, NERC/NSF funded project called Greenland Outlet Glacier Geophysics (GrOGG). It is a 60 MHz phase coherent pulsed radar with a 15 MHz bandwidth. It has been flown extensively in Antarctica including surveys over the Thwaites Glacier catchment and surveys of large sectors of East Antarctica (Young et al., 2011). New algorithms employed with HICARS allow a horizontal resolution of less than 5 m and vertical resolution of ± 10 m.

2.1.6 Technical University of Denmark (TUD)

A small amount of data are included which were collected in the 1970's by the TUD using a 60 MHz echo sounder, analogue recording and inertial navigation. The quality of these data is, in general, poorer than the more recent data so they are only used when there are no other data sources within 50 km (Fig. 1) (Bamber et al, 2001).

2.2 Surface elevations

All the airborne data were provided as ice thickness measurements but we require bed elevation. Therefore, a surface elevation estimate is needed. For most of the CReSIS and OIB campaigns, the flights also operated the NASA Airborne Topographic Mapper (ATM): an airborne laser altimeter. ATM scans the surface and is resampled to a horizontal spacing of 50 m with data averaged into 80 m diameter platelets. ATM was not always available and so our secondary source of surface elevations was from a DEM of the whole of Greenland. The surface DEM used was produced as part of the Greenland Ice Mapping Project (GIMP). It has a horizontal resolution of 90 m and is a multi-sensor DEM derived from data collected between 2000 and 2009. This DEM was created from MODIS, AVHRR, ASTER, SPOT and RADARSAT datasets merged with the ASTER GDEM and the (Bamber et al., 2001a) DEM vertically co-registered using ICESat data. Validation against ICESat data, indicated vertical errors of ± 5 m on the ice sheet and ± 7m for the unglaciated margins (Howat, pers comm).

2.3 Bathymetry

Numerical modelling over long timescales, such as glacial-interglacial cycles, or spinning up the thermodynamics in an ice sheet model requires basal geometry that extends out to the continental shelf: i.e. as far as the maximum glacial extent which can reach to the shelf edge several hundred kilometres from the present-day ice limit under full-glacial conditions (Evans et al., 2009); (Dowdeswell et al., 2010); (Cofaigh et al., in press). To achieve this requires inclusion of bathymetric data. Here we used the International Bathymetric Chart of the Arctic Ocean (IBCAO) v3 (Jakobsson et al., 2012). This is an interpolation of various bathymetric data from the entire Arctic Ocean and a DEM for Greenland, Ellesmere Island and Iceland. It was supplemented with additional data from soundings in the Jakobshavn fjord, which were included in the CReSIS Jakobshavn grid described above. Other changes to this dataset, and the reasons for these changes, are as described in Sect. 3.2.3.

3   Methods

All the airborne data were transformed onto a polar stereographic projection with standard parallel at 71°N and a central meridian of 39°W. All invalid data, as defined by the instrument teams, all data outside Greenland, and all data with ice thickness less than 0 m were removed. Rather than interpolating ice thickness, which has a discontinuity at the ice sheet margin, we interpolate bed elevation, as this varies smoothly across the ice edge. By combining the high resolution surface topography with the bed data we can create a more realistic ice margin. Where CReSIS thickness data were acquired within 2 days and within 1 km of an ATM surface elevation, the ATM estimate is used to convert to bed elevation. For 18% of ice thickness estimates from CReSIS when ATM was also flown, no surface elevation estimate was recorded. In those cases, and for all other datasets, the GIMP DEM was used. This introduces a potential error in the derived bed elevation if there has been a change in surface elevation (dh/dt) between the acquisition of the ice surface and thickness data.

For most of the data (74%) this is not relevant because simultaneous LIDAR data (such as ATM) are available. For the remainder, the GIMP DEM was used. The exact time stamp of GIMP is unclear as it was derived from a mosaic of images. For Jakobshavn, Helheim, and Kangerdlugssuaq glaciers, however, it was based on imagery from 2007, and for Petermann glacier from 2003 (personal comm; I. Howat). As the largest dh/dt values are found over these outlet glaciers, a correction was applied in these areas. Dh/dt values were taken from an ICESat based estimate for Greenland, covering the period 2003-2009, interpolated to 1km resolution (Hurkmans et al., submitted). Annual dh/dt values were added for years between the IPR data point and the time stamp for the area of interest. We assumed GIMP is representative for approximately mid 2003 and 2007, respectively, and using the the month of the IPR measurement, the appropriate fractions of the dh/dt values for the GIMP year (2003 or 2007) and the IPR year were taken into account. Of the 26% of data points for which GIMP was needed, 22% were located over one of the four glaciers mentioned above and have been corrected, therefore, for dh/dt. Most of the remainder (e.g. nearly all the AWI data) are in the interior where dh/dt values are at the few cm a^{-1} level.

To interpolate the bed, it is necessary to delineate glaciated and ice-free terrain. A land surface mask was created by merging a number of data sources. The coastline of Greenland was smoothed to a 1 km resolution from the Danish Ministry of Environment (formerly KMS)

218   1:2,500,000 scale vector maps of the coast. The Canadian Arctic and Iceland were separated

219   from the Greenland coast by identifying distinct polygons. The ice sheet and periphery ice

220   caps were identified using a binary mask (Howat I.M and A. Negrete, in prep) produced from

221   a combination of Landsat 7 panchromatic band imagery from July-Sept 1999-2001 and

222   RADARSAT-1 SAR amplitude images from autumn 2000. Data were provided at 180 m

223   horizontal spacing and reprojected onto the polar stereographic grid at 1 km resolution.

224   Gridboxes with over 50 % ice cover were considered to be ice sheet / ice cap. Ice shelves

225   were categorized separately from the rest of the ice sheet. The existence of an ice shelf was

226   determined by the presence of a grounding line from Interferometric Synthetic Aperture

227   Radar. Grounding lines were provided by Eric Rignot (JPL) and are a sub-set of those

228   previously published (Rignot et al., 1997). The ice shelf front was determined by a number of

229   means. In areas with ATM coverage, the ice front was located by a step in the elevations. This

230   was confirmed by ice front locations from the KMS maps. In the absence of ATM data, the

231   KMS ice fronts were used alone. On Storstrommen and Ostenfeld glaciers, no front was

232   present in maps or ATM data but InSAR confirmed the presence of an ice shelf. In these

233   cases, a small shelf was added based on Google Earth imagery. No attempt was made to

234   determine the bed elevation of peripheral glaciers where no thickness data exist (see Fig. 1)

235   and, in this case, it is the glacier surface elevation that is obtained, if resolved at all at 1 km:

236   i.e. ice thickness is zero.

237   **3.1   Data editing**

238   The data were gridded with 5 km postings and a 3 standard deviation filter was applied twice

239   to remove elevation outliers. This removed 0.3% of datapoints. Visual inspection indicated

240   that the filter had removed noisy data but a small number of anomalous measurements

241   remained. These were tracks which were not picking the ice/bed interface but appeared to be

242   tracking an internal layer. A coarse filter was applied whereby a data point was removed if the

243   bed elevation was more than 500 m different from the previous estimate (Bamber et al.,

244   2001b). This was only applied in areas where there was previous data coverage and not in

245   areas with large relief or areas of high surface velocity (>100 m/yr). Visual inspection also led

246   to the removal of several other tracks from the CReSIS data after examination of the

247   echograms. In all, 98.6% of the data from the various campaigns were determined to be over

248   ice and of sufficient quality to be included.

3.2 Interpolation

All data which passed the quality checks were locally averaged into a quasi-regular 1 km and 2.5 km resolution grid, which reduced the disparity in along and across track spacing of the data. The resolution of the two grids was chosen based on the data density as indicated in Figure 1 and supplementary Fig. S3. A 1 km posting DEM results in about 20% of grid points containing data in areas where the across-track spacing is greater than 20 km. For sectors of the interior, across-track spacing can be more than 50 km (Fig 1). In general, however, bed gradients are smaller in these areas and a lower resolution is adequate for capturing the large scale relief. In areas identified by the mask as being unglaciated, surface elevation data from the GIMP DEM were included in the quasi regular grid.

The 1 and 2.5 km quasi-regular grids were interpolated to regular grids using ordinary kriging. The GSLIB library (Deutsch and Journel, 1997) was used to calculate variograms. Separate variograms for the two resolutions were calculated and an exponential function was fitted to the first 100 km of each variogram using a non-linear least squares scheme. These variograms were used to interpolate the quasi-regular grids using a nugget of 50 m to take account of uncertainty in the airborne data. A maximum of 50 quasi-regular grid points were considered and the maximum search radius was set to 250 km so that a result was obtained everywhere. The 2.5 km grid was bilinearly interpolated to 1 km resolution. This avoids artefacts produced from interpolating sparse data at 1 km, and results in a grid at a single horizontal posting, which affords a simpler data structure compared with a multi-resolution or nested grid approach. The two grids were combined, with the higher resolution one being used for all areas which had sufficient data density (supplementary Fig. S3) and the lower resolution grid used elsewhere. The merging was done using a Hermite basis function of width 20 km across the boundary.

Ice thickness was derived by subtracting the bed from the GIMP DEM for every grid point defined as ice covered in the mask. The minimum ice thickness at the margin was set to 50 m. Wherever the thickness was less than this, the bed was lowered to be 50 m below the GIMP surface elevation. This is necessary because thickness data around much of the margin of the ice sheet does not exist (Fig. 1) and therefore, subglacial bed elevations cannot be determined

280   directly in these areas but must be interpolated close to a discontinuity in thickness and
281   surface elevation.

3.2.1 Ice shelves

283   Nine ice shelves are present around the Greenland Ice Sheet according to our land mask. The
284   surface elevation for these was, in general, found by combining all GIMP, ATM with airborne
285   ice thicknesses data, which had been converted to surface elevation using the assumption of
286   hydrostatic equilibrium. Some ice shelf grid-points contain no elevation data from any source,
287   in which case, nearest neighbour interpolation from other ice shelf elevations was used. For
288   Nioghalvfjerdsbrae and Zachariae Isstrom, only ATM and airborne elevations were used as
289   GIMP elevations were not in agreement with the other data sources.

290

291   Ice shelf thickness was calculated from surface elevation using the assumption of hydrostatic
292   equilibrium (see (Griggs and Bamber, 2011) for a full description of the method). A constant
293   firn density correction of 10 m was used throughout. An ocean water density of 1027 kg m^{-3}
294   and an ice density of 917 kg m^{-3} were chosen. The same parameters were used to convert the
295   airborne ice thicknesses to surface elevations and back again, which was done to increase data
296   coverage. In areas where the surface elevation was less than 10 m above sea level, and ice
297   thickness would be calculated to be negative, surface elevations were interpolated from
298   thicker neighbouring datapoints. In areas where hydrostatic equilibrium is invalid such as the
299   shear margin of Petermann Glacier, ice thickness was interpolated from surrounding
300   datapoints.

3.2.2 Mass conservation

302   In many areas, there were no airborne data within a few kilometres of glacier termini. If there
303   are unglaciated regions closer to the terminus than the closest airborne data, they dominate the
304   interpolated elevation at the terminus, artificially raising the bed elevations. This was
305   corrected for in two ways. In four areas, we had an airborne survey line which crossed the
306   fast-flowing region of the outlet. In these cases, we took the bed elevations at the airborne
307   data points along with the known velocity (Joughin et al., 2010) and the 30-year average
308   modelled surface mass balance (Ettema et al., 2009) and calculated the expected bed elevation
309   based on the principle of mass conservation. The direction of flow of the ice from the location
310   of the airborne data to the terminus was determined by the direction of the velocity vector. A

more sophisticated approach has been developed and demonstrated on 79North Glacier (Morlighem et al., 2011) and we intend to incorporate results using this method in future releases of the product.

Figure 2 shows the result of correcting the bed elevations using this method for one outlet. In (a) the original elevations are shown and (b) shows the bed after the application of the technique. A region of 7 km length is altered from the position of an airborne track inland, following the region of faster flowing ice as shown in Fig. 2(c). It is clear that after this change, thick ice is now able to flow along the outlet where previously a ridge was mapped due to the inclusion of measurements of unglaciated terrain closer to the terminus than legitimate ice thickness data.

In a number of other locations, the same situation arises but either there are no velocity measurements, the velocity vectors make the ice flow out of the region of fast flow, unsurveyed tributaries join the main flow oceanward of the airborne data, or the airborne measurements do not cross the entire fast flow channel. In these cases, our mass conservation scheme cannot be readily implemented and instead, linear interpolation was used to remove the artificial ridges at the terminus. In all cases where data were changed from the original interpolated values, a mask is provided which notes the change made, the reason and the original interpolated value.

3.2.3 Bathymetry interpolation

The interpolated bed elevations were merged with the known bathymetry of the fjords and oceans around Greenland (Jakobsson et al., 2012). A smoothing distance of 3 km was used on the ocean-ward side of the coastline to merge the two datasets. No smoothing was used in the south-west where the coastline consists of many narrow fjords. Any smoothing zone in this area would result in the fjords being completely removed. In a number of areas the bathymetry is not well known and IBCAO does not contain any observations (see figure here: http://www.ngdc.noaa.gov/mgg/bathymetry/arctic/images/IBCAO-sources.jpg). In some of those areas, our coastal bed elevations suggest that the IBCAO bathymetry is significantly too high. In a number of fjords, including Jakobshavn and Petermann, the bathymetry data are particularly high and appear to follow the ice surface rather than the fjord bottom (which is at

about -500 m depth) . In Jakobshavn fjord, we replaced the IBCAO bathymetry with that from the CReSIS grid described in Sect. 2.1.1. In other areas, no bathymetric data were available and the IBCAO interpolation contains a fjord which is much shallower than the bed elevation at the grounding line or glacier terminus. For example at Hagen Brae, the fjord just seaward of the grounding line was 200 m shallower than the bedrock elevation at the grounding line. To ensure realistic ice flow pathways, we lowered the bathymetry in regions affected in this way to create a smooth surface without a discontinuity at the land/ocean margin. Where ice shelves are present, a similar approach was undertaken involving interpolation of the bed elevation at the grounding line seaward to the first measured IBCAO value. A minimum depth for the cavity beneath the ice shelf of 10 m was imposed, distal from the grounding line.

We have not included any new bathymetric data as this is beyond the scope of this study. In the affected areas, we interpolated between the grounding line/glacier terminus and beyond the fjord mouth using triangulation. The results were smoothed over 2 km around the triangulated values. End points for the interpolation were chosen to ensure that ice flows out of the fjord. The approach is somewhat subjective and results depend upon the choice of endpoints and can produce steep gradients in the across fjord direction beyond the lateral region of adjustment. As a consequence, we provide a mask with the data which indicates where the bathymetry has been changed in this manner.

Section on IBCAO v3 deleted.

4   Results and Discussion

The final bed DEM was referenced to the EIGEN-GL04C geoid (Forste et al., 2008) and is shown in Fig. 3. All the figures plotting bed elevation are referenced to the geoid, which has a range of around 10 m in the north-west to 65 m in the south-east. Alongside the bed DEM, we also produced grids of ice thickness, surface elevation, error maps of surface and bed elevation, the land surface mask, the geoid-ellipsoid separation, a mask showing changes made to the bed elevation post-interpolation, the bed elevation and ice thickness without any intervention, a mask showing the data sources used for the ice shelves and a grid of the number of airborne data per grid-point. The dataset includes features which were not visible in previous compilations and improves the representation of many features previously observed. This improvement is most noticeable in areas of relatively high relief close to the ice margin and, in particular, where dense grids were flown (Fig. 1, and supplementary Fig.

S1). Under Jakobshavn (Fig. 4 and 5), there is a dendritic, channel system extending for about 325 km from the current grounding line into the interior almost as far as the ice divide (Hoch et al., 2011). It seems likely that this is a paleo-fluvial feature that predates ice cover in Greenland and is clearly an important control on ice dynamics in the basin. The presence of a deep trough extending into the interior has been previously reported, based on SAR processing of MCoRDS data (Hoch et al., 2011).

We make a direct comparison between the new bed DEM and the one from the previous compilation (Bamber et al., 2001b) in two regions to illustrate the improvements made. Figure 4(a) shows the Jakobshavn region in the new dataset (see black box in Fig. 3 for location) and (b) shows the same region in the older dataset. First, it is apparent that the true resolution (as opposed to the grid spacing) of the new DEM is significantly improved throughout the region, better characterising the undulating sub-glacial mountains to the north of the trough between 0-100 km along the x axis and 120-200 km along the y-axis. Secondly, the deep trough under the main fjord is present. The previous compilation showed almost no evidence of this trough due to i) a lack of bed returns in this area and ii) the resolution of the older grid at 5 km. The width of the trough in the new DEM is 3-4 km. The region of fastest flow coincides fairly well with the location of the deepest ice (Fig. 4c). The trough in the new DEM is 1366 m below sea level at its deepest point compared to a maximum depth over the entire region of 556 m below sea level in the older dataset. The main trough of Jakobshavn Isbrae is not continuous in the new dataset, disappearing around 100 km on the x-axis and reappearing at about 140 km. This does not demonstrate that the trough is discontinuous, but only that there are insufficient data to confirm the trough's presence or otherwise in this region.

Figure 4(d) and (e) show a similar comparison for a 200 km by 500 km area along the north-west coast of Greenland (red box in Fig. 3). In the older DEM, there is some evidence of bed troughs but they are wide and not well aligned with the areas of fast flow seen in Fig. 4 (f). Figure 4(d) shows numerous troughs, all aligned with areas of fast flow. The wide unconstrained minimum seen in Fig. 4(e) at a depth of 438 metres below sea level is has now become a deep, narrow trough under a region of fast flow with a maximum depth of 1219 m below sea level. This improvement is due mainly to the increased data coverage and to a lesser extent the higher resolution of the DEM. Similar comparisons can be made elsewhere

and the full fidelity of dataset becomes apparent when examining smaller regions of a few hundred kilometres in extent. To further illustrate the spatial detail of the new bedrock topography and the bathymetry we display shaded relief plots of the Jakabshavn and Kangerdlugssuaq regions in Figure 5 a and b, respectively. Fig 5a indicates the small-scale structure of the dendritic network of troughs extending inland and the complex bathymetry near the mouth of the glacier. The impact of coarser track spacing on the topographic detail can be clearly seen at -2200 to -2260 Northing, -150 to -50 on the y axis. This region possesses a feature that, likely, would be deeper and narrower if adequately resolved, like the trough adjacent and just north of it. The Kangerdlugssuaq fjord is a region where the IBCAO bathymetry had to be lowered by several hundred metres to avoid an artifical "cliff" at the ice margin. The fjord in this region is up to 700 m below sea level in our data set, which is only partially captured by the original IBCAO dataset. Inland, the trough that the glacier follows is around 1500 m in depth and in places less than 10 km in width. This type of extreme topographic relief requires both dense data sampling (Fig. 1) and appropriate resolution interpolation (1 km in this region) to adequately capture its structure.

Hydraulic potential

Important information about potential water-flow routing beneath the ice sheet can be deduced from high quality surface and bed topography (e.g. (Wright et al., 2008). Here we use the theory developed by (Shreve, 1972) to calculate the subglacial hydraulic pressure potential and the flow paths that follow the gradients in this potential. Figure 5 shows the major subglacial flow paths colour coded, darker colours indicating a larger catchment contributing to the flow. In the main the flow paths follow the direction of ice flow. For a given thickness of water per unit area, the colour coding provides a measure of the volume of water associated with a particular flow path which is largely controlled by the size of the ice surface catchment area for a given outlet glacier. Not surprisingly and similarly to a previous analysis (Lewis and Smith, 2009), glaciers such as Helheim, Kangerdlussuaq and Jakobshavn Isbrae, which have large catchments, have the potential for substantial subglacial drainage beneath their main trunks. The Northeast Greenland ice stream is also a major conduit for subglacial water flow and it is interesting to note that high geothermal activity has been proposed to produce a large volume of water near the onset of this feature (Fahnestock et al., 2001). Combined with modelled estimates for basal melt and surface percolation to the bed,

437 ~~the flow routing shown in Fig. 5 can be used to estimate subglacial water discharge at the ice~~
438 ~~margin and compared, for example, with field based estimates (DeFoor et al., 2011).~~

4.1 Error assessment

440 For most applications, a reliable estimate of the uncertainty in the DEM is essential. To
441 determine the total error we considered i) the random error in the thickness observation and ii)
442 the impact of interpolation. The former was assessed from a track-to-track difference analysis
443 of the data (a combination of repeat track and quasi-crossover differences). We consider the
444 differences between any two measurements obtained at different times within a 50 m area.
445 This has the advantage of including along track differences in cases where aircraft fly repeat
446 tracks in subsequent seasons as well as crossovers. In total, 24.1 million differences were
447 calculated. The data were split by campaign to calculate both inter and intra-campaign
448 differences. These data, strictly, only provide information on repeatability and not systematic
449 biases but, because we are considering inter-campaign differences, we believe that biases due,
450 for example, to timing, navigation or radar calibration errors, will be captured in this analysis.
451 Only biases common across all campaigns (such as a common error in the radio wave velocity
452 in ice used) will not be seen.

453

454 Table 2 summarises the differences for each set of campaigns grouped by instrument and/or
455 institute. Figure 6a shows the histogram of cross-over and along-track differences, which has
456 a Gaussian distribution with a bias close to zero. We estimate the random error as $1/\sqrt{2}$
457 (because a difference contains two observations) of the standard deviation of the inter-
458 campaign difference of the nearest data to a grid point. In the case of TUD data, where there
459 are no differences, we use the standard deviation of all inter- and intra- campaign values.

460

461 Next we consider the error due to interpolation. This increases with distance from an
462 observation but is also a function of the properties of the underlying surface. We estimate this
463 uncertainty using a bootstrap approach for two classes of bed topography: coastal and interior.
464 To differentiate these two classes, we calculated the standard deviation of the bed elevation in
465 overlapping 50 km boxes and used this as a measure of basal roughness (supplementary Fig
466 S3). From visual inspection, a standard deviation threshold of 170 m was used to distinguish
467 the two classes of bed. For these two zones, all CreSIS data from 2000 onward were used in a

bootstrapping approach to determine the effect of interpolation as a function of distance from an observation. The difference between the interpolated and observed elevation as a function of distance was used as a measure of the interpolation error. The standard deviation of these differences for the two zones is shown in Fig. 6b. Exponential curves were fit to the differences and are also plotted in Fig 6b. These curves were used to determine the error due to interpolation as a function of distance.

~~We took the CReSIS data from the 2000s in two locations that are broadly representative of these two categories: first, the basin of the Petermann Glacier being representative of coastal, high relief and, secondly, a 550 by 1200 km region centred on Summit, being representative of the interior. In both areas, we repeated the interpolation while removing a random portion (of up to 90%) of data. The difference between the interpolated and observed elevation as a function of distance was used as a measure of the interpolation error. The standard deviation of these differences for the two areas is shown in Fig. 6b. A line with correlations of r^2=0.92 for the Petermann basin and r^2=0.80 for the central plateau was fitted to the differences. This fit was used to determine the error due to interpolation as a function of distance. Areas with surface elevation below 2000 m were assigned as "coastal" and areas above this elevation as "interior".~~

The two sources of error were combined in quadrature to produce a map of the uncertainty in the bed elevation (Fig. 7). As expected the error increases significantly with distance from an observation and the largest errors occur in coastal regions where extrapolation, rather than interpolation, was needed. For the peripheral ice caps where, in most cases, there are no airborne data, the bed elevation is poorly constrained and these areas are included only for completeness. Better estimates of the ice thickness in these areas could be estimated using an ice-surface area/volume scaling approach (Bahr et al., 1997).

5   Conclusions

A large volume of high quality new data have become available since the last comprehensive dataset of ice thickness in Greenland was compiled (Bamber et al., 2001b). We improve on the earlier compilation in several respects. Most importantly, we have included extensive new data sets acquired by several different groups over the last decade. A significant effort has been made, during this period, to sound the bed of fast-flowing outlet glaciers which were

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either missed or proved challenging targets for the previous generation of IPR systems due to high attenuation and clutter. These new datasets also provide improved coverage of the, previously, sparsely surveyed interior. The greater coverage and dense network of flightlines in many coastal areas, and some inland regions, (Fig. 1 and supplementary Fig. S1) supports a grid spacing of 1 km, five times higher than the previous compilation. As a consequence, many basal features, in particular basal troughs containing outlet glaciers, are now adequately resolved in the bed topography. In the ice sheet interior, flight lines are less dense but the topography is generally smoother and, therefore, a coarser resolution captures the salient features. For convenience, the bed DEM is provided at a single posting of 1 km alongside a grid indicating whether the value is interpolated or based on observation. An error map for the DEM was also calculated and indicates areas where additional data would be particularly useful. The estimated volume of the ice sheet is $2.96 \times 10^6 \, \text{km}^3$ compared with $2.93 \times 10^6 \, \text{km}^3$ obtained previously (Bamber et al., 2001b). 22% of the ice sheet bed is below sea level and, accounting for this and the thickness of the firn layer but excluding any glacio-isostatic adjustment, we estimate that the ice sheet has the potential to raise global mean sea level by 7.36 m were it all to melt.

The data presented here represent a major advance in our knowledge of the topography of the bedrock of Greenland. However, there are several areas where data are currently lacking such as along much of the central and north eastern margins of the ice sheet (Fig. S1). While the interior of Greenland is relatively smooth, there are still many areas where the distance to the closest observation is more than 50 km, resulting in uncertainties in bed elevation exceeding 100 m and missing potential short wavelength relief that is evident, for example, inland from Jakobshavn Isbrae (Fig 5a). Although this may be less important for numerical modelling, such detail provides valuable insights into the genesis of the subglacial landforms and the geomorphology of the bed. Recent results suggest reduced uncertainty and significantly increased spatial detail can be obtained from radar tomography (Jezek et al., 2011;Paden et al., 2010), while mass conservation approaches also show promise for poorly sampled outlet glaciers (Morlighem et al, 2011). Wide application of these techniques could significantly improve current mapping by decreasing the amount of interpolation needed over deeply incised outlet glaciers. Bathymetry seaward of the glaciers and beneath ice shelves is currently poorly characterised in some areas and non existent in most. This is a major gap in our current knowledge and requires further effort by the community. New data are being

534 acquired over Greenland each year and we intend to issue new releases of the products when
535 sufficient new data warrant this. The complete set of grids, metadata and documentation are
536 available in netcdf and geotiff format from the lead author (JLB). Users will be notified of
537 new releases as they become available.

538

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547

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670

671   Table 1: Airborne data sources used in this study.

Source	Time Period	Reference	Line kilometres used
CReSIS	1993 - 2012	(Gogineni et al., 2001)	337,650
Alfred-Wegener-Institut	1996-9, 2004, 2010	(Nixdorf et al., 1999)	58,350
Warm Ice Sounding Explorer	2008 - 2010		13,180
Pathfinder Advanced Radar Ice Sounder	2009	(Raney, 2010)	5880
High-Capability Radar Sounder	2011	(Peters et al., 2007;Peters et al., 2005)	5270
Technical University of Denmark	1970's	(Bamber et al., 2001b)	70*

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673   * Approximately 30,000 km of line km were flown by TUD but we only used these data if no

674   other, more recent, observations were available. Hence the lower number shown in Table 1.

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676

677   Table 2: Inter-track differences for intra-instrument campaigns: i.e. the differences are

678   estimated for CreSIS vs. CreSIS, AWI vs. AWI etc. only.

Data Source	Standard deviation	Number of differences
CReSIS 90's	47.95	52621
CReSIS 00's	58.27	15,240,399
AWI	11.18	154
WISE	90.87	22
HICARS	54.56	8,033,817
PARIS	20.27	3327

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Figures

Figure 1: Data sources. CReSIS90 includes all data from 1993 to 1999 as used in (Bamber et al., 2001b). CReSIS00 includes all data derived from CReSIS instruments between 2000 and 2012.

Figure 2: (a) Outlet glacier on the north-west coast before mass conservation approach was applied. Red box outlines the area where the mass conservation approach altered the bed elevations. The location of the region is shown in the blue rectangle on Fig. 3. (b) Bed elevation after mass conservation has been applied in the topographically constrained fast flow region. (c) Ice surface velocity resampled to 1 km postings (Joughin et al., 2010). Bed elevations in all plots are with respect to the EIGEN-GL04C geoid.

Figure 3: (a) Bed elevation and bathymetry. Bathymetry is plotted on an opaque scale. Black and red boxes show the location of the Jakobshavn and north-west coastal regions shown in Figs. 4a-c and d-f, respectively. The blue box shows the location of the region shown in Fig. 2. The black contour is at 0 m elevation. (b) Ice thickness as determined from the difference between surface and bed DEMs.

Figure 4: (a) Bed elevation for Jakobshavn sector from the new DEM, (b) as for a) but from old DEM (Bamber et al., 2001b), (c) ice surface velocity from InSAR (Joughin et al., 2010), (d) bed elevation along north-west coast from the new DEM, (e) as for d) but from old DEM (Bamber et al., 2001b) and (f) ice surface velocity from InSAR (Joughin et al., 2010). For locations of the regions shown, see Fig. 3.

Figure 5: (a) Shaded relief plot looking inland from the coast with the mouth of Jakobshavn Isbrae at about -2150 km northing. (b) as for (ab) but for the Kangerdlugssuaq glacier basin.

Figure 6: (a) Histogram of inter-track, intra-instrument differences for all airborne datapoints within 50 m of each other. (b) Standard deviation of difference between bed elevations created using all data and using a sub-set of the data in "marginal" areas with a roughness standard deviation greater than 170 m (asterisks) and the more "inteiror" region where the

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712   roughness was below this threshold (diamonds) (see supplementary Fig. S3). An exponential

713   fit to the data is shown as solid ("margins") and dashed ("interior") lines.

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715   Figure 7: RMS error in bed elevation due to the combined uncertainties in the IPR data and

716   those due to interpolation.