Automated mapping of glacial overdeepenings beneath contemporary ice sheets: approaches and potential applications

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ABSTRACT There is growing awareness of the significance of overdeepenings in ice sheet systems. However, a complete understanding of overdeepening formation is lacking, meaning observations of overdeepening location and morphometry are urgently required to motivate process understanding. Subject to the development of appropriate mapping approaches, the availability of high resolution subglacial topography datasets covering the whole of Antarctica and Greenland offer significant potential to acquire such observations and to relate overdeepening characteristics to ice-sheet parameters. We explore a possible method for mapping overdeepenings beneath the Antarctic and Greenland ice sheets and illustrate a potential application of this approach by testing a possible relationship between overdeepening elongation ratio and ice sheet flow velocity. We find that hydrological and terrain filtering approaches are unsuited to mapping overdeepenings and develop a novel rule-based GIS methodology that delineates overdeepening perimeters by analysis of closed-contour properties. We then develop GIS procedures that provide information on overdeepening morphology and topographic context. Limitations in the accuracy and resolution of bed-topography datasets mean application to glaciological problems requires consideration of quality-control criteria to (a) remove potentially spurious depressions and (b) reduce uncertainties that arise from the inclusion of depressions of non-glacial origin or those in regions where empirical data are sparse. Potential criteria are introduced to address the problem of overdeepening elongation, and discussion of this example serves to highlight the limitations that mapping approaches – and other potential applications of such approaches – must confront. We

predict that improvements in bed-data quality will reduce the need for quality control procedures and facilitate increasingly robust insights from empirical data.

Keywords: overdeepenings, automated landform mapping, glacial erosion, landscape evolution, Antarctica, Greenland.

1. Introduction & aims

The mechanisms by which glaciers and ice sheets form spectacular alpine and fjord landscapes are well known, and such landscapes have been exploited widely for purposes of palaeo-glaciology and process understanding (e.g. Glasser and Bennett, 2004; Hooke, 1991; Sugden, 1978). Recently, this understanding has been aided by implementation of simple ice-erosion laws within numerical models, which are able to simulate depths and patterns of glacial incision with compelling success (e.g. Harbor, 1992; Kessler et al., 2008; MacGregor et al., 2000). However, the mechanisms that produce overdeepenings (Fig. 1) remain unclear, and the implementation of candidate processes within ice-erosion models, including quarrying-related ice-erosion feedbacks, has met with limited success (e.g. Egholm et al., 2012).

The reverse-bed gradient that occurs in the presence of an overdeepening has been shown to exert strong influence on glacier hydrology, ice dynamics, and ice-mass stability (cf. Cook and Swift, 2012; Creyts and Clarke, 2010; Dow et al., 2011; Schoof, 2007; Stokes et al., 2014; Thomas and Bentley, 1978; Weertman, 1974) (cf. Fig. 1C), meaning a complete understanding of overdeepening formation and morphology is essential to elucidate and understand critical ice-bed processes and to inform predictions of past and present ice-mass behaviour. Cook and Swift (2012) have argued that process understanding has been disadvantaged by an absence of quantitative studies of overdeepening morphology that perhaps reflects the term's uncertain etymology and, as a possible consequence, an unconscious disregard of overdeepenings as distinct morphological features. Hence, fundamental data are urgently required to motivate process understanding, as well as to provide quantitative test-data for ice-erosion models.

In this paper, we develop a computationally efficient GIS-based methodology for mapping of overdeepenings and quantification of their morphometry. Inspired by studies that have investigated glacier depositional and erosional phenomena (e.g. Clark et al., 2009; Evans, 2006; Stokes et al., 2013), we develop methods that can be used to extract overdeepening length, width, depth and volume, as well as overdeepening long- and cross-profiles that pass through the deepest point. Measurements of these phenomena require delimitation of the basin perimeter and identification of basin in- and out-flow points, and also require identification of overdeepenings within overdeepenings (i.e. nested overdeepenings; Fig. 1D). From these phenomena, further metrics, including elongation ratio (cf. Clark et al., 2009), normal and adverse slope lengths and gradients, and planform area, can then be derived (Fig. 2).

We use comprehensive subglacial topography datasets for the Antarctic and Greenland ice sheets (Bamber et al., 2013; Fretwell et al., 2013) to illustrate a possible application of our methodology that seeks to relate overdeepening form to ice velocity. These recently published datasets provide an incentive for the development of landscape analysis tools that can systematically examine ice-sheet beds, not least because the application of traditional methods of geomorphological mapping at this scale is inappropriate. Nevertheless, automated analysis of such datasets need to be appreciative of dataset quality and resolution, the presence of features of non-glacial origin, and the inherently different timescales of ice mass and landscape response.

2. Study areas and datasets

2.1 Study areas

The landscape beneath present-day ice sheets provides an unparalleled opportunity to elucidate ice-bed processes and evolution because of the size of the ice-covered area and, where ice cover remains present, the absence of thick postglacial deposits that in palaeo-glaciated landscapes accumulate within areas of deep erosion. Comprehensive subglacial topography datasets for the Antarctic (Bedmap2) and Greenland ice sheets have recently been made available (Bamber et al., 2013; Fretwell et al., 2013) and these are used in the example application of our methods in section 5. For practical purposes, the development of these methods was undertaken on a relatively small domain surrounding the Byrd Glacier catchment and Transantarctic Mountains, adjacent to the Ross Ice Shelf, in East Antarctica (Fig. 3). This 5.22×10^5 km^2 region provides an excellent methodological test-bed, combining substantial variability in relief with the presence of a large number of bed depressions that exhibit a range of depths and areas.

2.2 Antarctic subglacial topography

The Bedmap2 dataset provides subglacial and continental shelf topography for the Antarctic continent (Fretwell et al., 2013) using the most up-to-date compilation of empirical ice-thickness measurements for the Antarctic ice sheets. Raw ice-thickness data for Bedmap2 have been derived from a variety of sources, including: direct airborne radar sounding and seismic measurements; satellite altimetry and free-air gravity surveys; and 'synthetic' data computed using a 'thin-ice' model. The rationale for including modelled topography within the source data was to prevent rock outcrops from overly skewing the ice-thickness distribution in mountainous areas where few empirical measurements exist. Although this output gives the appearance of accurate relief within ice-marginal mountain ranges, it is worth noting that this topography is not directly constrained by any empirical data. Continental shelf topography is derived from the GEBCO 2008 bathymetric compilation mosaiced with sub-ice shelf data from Timmermann et al. (2010).

The Bedmap2 topography is rendered on a 1-km grid but empirical and synthetic measurements of ice thickness were sampled at 5 km, primarily because the distribution of empirical measurements, which require interpolation (kriging) to form a continuous surface, did not warrant a higher resolution (Fretwell et al., 2013). Notably, the spatially non-uniform distribution of ice-thickness measurements obtained by airborne radar surveys, in which across-track sampling density is potentially 3 or 4 orders of magnitude lower than the density along the flight tracks, means that even large, valley-scale features may be absent or resolved poorly. Furthermore, the fragmentary nature of completed radar surveys carried out across Antarctica has left many regions sparsely constrained. For example, in Bedmap2, 80% of grid cells have data within 20 km, and the greatest distance from a grid cell to the nearest data point (the 'poles of ignorance') is ~ 230 km (Fretwell et al., 2013). For this reason, the non-genetic term *depression* is used in the methodological sections of this study, thereby avoiding the implication that all basin-like features in the digital elevation models (DEM) surface are genuine closed-depressions and/or glacial overdeepenings.

2.3 Greenland subglacial topography

Subglacial and continental shelf topography for Greenland is provided by Bamber et al. (2013). As with Bedmap2, topography in this dataset is rendered on a 1 km grid with subglacial topography mainly derived using ice-thickness measurements obtained from airborne radar surveys and satellite observations. As such, similar error sources, assumptions and levels of uncertainty exist. Continental shelf topography is sourced from the most recent IBCAO (International Bathymetric Chart of the Arctic Ocean) compilation of offshore bathymetric datasets (Jakobsson et al., 2012), supplemented with additional soundings from Jakobshavn fjord. It should be noted that where bathymetry is not well known, or observations do not exist, bed elevations are often underestimated by up to several hundred metres, particularly within fjords (Bamber et al., 2013).

2.4 Additional datasets

Higher-resolution ice thickness datasets for several areas of Antarctica were obtained from the Centre for Remote Sensing of Ice Sheets archive (CReSIS; https://data.cresis.ku.edu) to evaluate the implications of DEM resolution for the delineation of overdeepening perimeters. These products, which have restricted geographical coverage, are derived from airborne radar surveys and are published at a grid spacing of 500 m. Ice-surface velocity data derived from InSAR observations over Antarctica and Greenland were sourced from datasets compiled by Rignot et al. (2011a) and Joughin et al. (2010a), respectively.

3. Automated mapping of overdeepenings

3.1 Delimitation of overdeepenings in the landscape

A key challenge in mapping geomorphological phenomena is delineating their boundaries. For example, overdeepenings do not represent isolated pockets of deep glacial erosion in an otherwise unmodified fluvial landscape. Most frequently, overdeepenings occur as areas of deeper erosion in the floors of deep, glacially-carved valleys (cf. Cook and Swift, 2012) and, as such, the flanks of an overdeepened basin are inseparable from those of the host valley. The use of hydrological tools to delineate overdeepenings (see section 3.2.1) by means of 'filling sinks' does not therefore necessarily have a strong, physical basis. Further, subaerial and subglacial hydraulic gradients will differ, meaning that an overdeepening that in the subaerial environment contains a lake will only contain a subglacial lake if the gradient of the adverse bed-slope exceeds 11 times that of the ice-surface gradient (the 'ponding' threshold; Clarke, 2005).

To avoid consideration of such complexities, we follow the definition provided by Cook and Swift (2012), who use *overdeepening* (verb) to describe the excavation of a topographic depression that, subglacially, would require ice, water and sediment to traverse a locally reversed (or adverse) slope. This usage therefore describes the creation of a specific landform, an *overdeepening* (noun), which on deglaciation would form a sedimentary basin or lake (cf. Fountain and Walder, 1998). This definition of an overdeepening as a subaerial 'closed depression' means that the elevation at the outflow point can be used to delineate the perimeter. Mapping of closed depressions in predicted subglacial hydraulic potentials (cf. Shreve, 1972) is avoided intentionally because depression form and location would depend partly on the morphology of the ice surface, which is inherently variable. A classification based on purely morphological grounds is therefore independent of glaciological processes. Clearly, closed-depressions can also be formed by non-glacial processes, including tectonic processes (e.g. by faulting), whilst some mapped depressions are artefacts resulting from interpolation between sparse empirical data. Methods of identifying erroneous depressions and tectonic basins are also therefore considered in this study.

Despite the morphological simplicity of closed-depressions, mapping methodology must overcome several important challenges. Firstly, automated analyses of DEMs at the ice-sheet scale, even at 1 km resolution, require computationally efficient techniques. Secondly, like other bedforms (cf. Clark et al., 2009), overdeepenings tend to develop a distinctive ovoid planform (Cook and Swift, 2012), but constraints imposed by topography often produce sinuous overdeepenings that follow the axes of large troughs (e.g. Fig. 1B), while others are influenced by geological structures or changing phases of ice-flow direction, resulting in circular or irregular shapes. Thirdly, overdeepenings are frequently nested (cf. Fig. 1D), with larger examples occasionally containing many generations of nesting. Finally, many overdeepenings beneath contemporary ice sheets may be relict landforms that represent erosion during earlier stages of glaciation, which may limit meaningful analysis of mapping results. For example, ice flow direction during depression formation, and thus the location of in- and out-flow points, cannot always be established.

3.2 Delineation methods

Three GIS-based methodologies were evaluated for the purpose of delineating depressions in a DEM surface that would constitute subaerial closed-depressions in a landscape. The third approach was adopted for this study.

3.2.1 Hydrological filling

In the post-glacial landscape, overdeepenings are sinks for water and sediment (e.g. Preusser et al., 2010; van Rensbergen et al., 1999). An instinctual approach to mapping overdeepenings is therefore to use GIS hydrological tools to identify sinks (or areas of internal drainage) across the digital terrain (e.g. Arnold, 2010). ‘Fill’ tools offer the simplest approach, and work by filling sinks to capacity, thereby creating a ‘depressionless’ DEM. A major disadvantage of this method is that smaller depressions located within larger depressions cannot be delimited, meaning overdeepenings within tectonic basins or rift systems, or smaller overdeepenings nested within larger overdeepenings, go unrecognised (Fig. 3). The wide size range exhibited by overdeepenings (Cook and Swift, 2012) means those solutions to this problem that use fill criteria to limit the size of the fill area for individual depressions are unworkable at the scale of whole ice sheets.

3.2.2 Terrain filtering

A more sophisticated yet computationally simple approach to mapping overdeepenings is to apply signal-processing techniques to the DEM surface (e.g. Leonowicz et al., 2009; Stumpf et al., 2013). This approach considers the landscape as a three-dimensional waveform within which depressions and mountain peaks represent anomalous interference. By filtering the elevation data at specific wavelengths, a smoothed surface largely voided of relief (interference) can be created. Overdeepened topography can then be extracted by calculating the negative residuals beyond a given threshold compared to its original form. Figure 4 shows how a two-dimensional, circular (200 km), low-pass Gaussian filter of the form

Feil! (1)

where sigma (σ) and mu (μ) are the standard deviation and mean of the elevation distribution respectively, can be used to produce a map of probable areas of overdeepening. In contrast to basin extents mapped using hydrological filling techniques (above), this approach lacks precise thresholds required for delimiting closed depression perimeters. Furthermore, the approach does not enable mapping of nested features. A more rigorous GIS-based approach is therefore needed that as appropriate to the complex morphology of large depressions.

3.2.3 Contour tracking

A final approach to delineating overdeepenings is to track changes in closed-contour length as an observer moves away from an elevation minimum. The novel contour-tracking process developed for this study is threefold. Firstly, a terrain analysis mask is calculated that delimits depression-like areas in the DEM surface (Fig. 5A–B). By eliminating large areas of the surface that are not depression-like, this initial step greatly enhances computational efficiency at the ice sheet scale. Secondly, locations of elevation minima are pinpointed using zonal statistical analyses to find the minimum point within each closed-contour that intersected the terrain-analysis mask (Fig. 5C). Finally, changes in contour length are tracked away from each elevation minima to identify sharp increases in length that would indicate the breach of a closed depression (Fig. 6A-B). The steps detailed below were implemented using toolboxes and commands found within ArcGIS 10.1 and GRASS GIS, with automated workflow achieved using packages within Python such as ArcPy.

Step 1 – Terrain analysis mask

Many existing methods of automated terrain analyses have their roots in differential geometry, using combinations of morphometric parameters such as slope, gradient, curvature and aspect to classify the form of the DEM surface (e.g. Brenning, 2009; Drăguţ and Blaschke, 2006; Evans, 1980; Klingseisen et al., 2008; MacMillan et al., 2000; Saha et al., 2011; Wilson and Gallant, 2000; Wood, 1996). Here, two parameters are used to delimit broad areas of depression-like relief: plan curvature (horizontal curvature, intersecting with the XY plane) and minimum curvature in the direction perpendicular to the direction of maximum curvature. The quantitative foundation to this methodology thus creates several free parameters, the sensitivity of which can be fine-tuned depending on the desired mapping criteria (cf. Table 1). For example, plan curvature and minimum curvature can be tuned to mask only small cirque-like features that are confined by high topography.

Step 2 – Finding elevation minima

Elevation minima are found by running a zonal statistical tool on the depression-like areas identified in Step 1. Where contours are completely within the terrain analysis mask, a grid of points with elevation attributes is created, from which points of minimum elevation associated with each contour are deduced. Repeat points, or erroneous elevation peaks, can be removed by systematic comparison with the elevation and the ID of the bounding contour.

Step 3 – Contour analysis

Depression perimeters are delineated by determining the highest elevation 'bounding contour' using a process of 'contour-tracking'. This method works by measuring the change in contour length between successive contours at increasing distance from the point of elevation minima (Fig. 6A). In the case of depressions, contour lengths will increase in size, and contour length will increase rapidly when a depression is 'breached'. For this study, a change in contour length of 90% from the preceding contour was chosen, which represents a factor-of-three increase in the bounded area. This method was used to identify generations of nested depressions (Fig. 6B) by running multiple passes whilst 'ignoring' previously identified depression breaches.

A range of parameters can be specified during this process (Table 1). For this study, values were chosen that provide sensible results, but these are far from definitive. In addition, classification of nested depressions can be undertaken, and for this study a simple top-down approach based on nested depression order was applied (Fig. 6C).

4. Automated extraction of overdeepening metrics

Whilst mapping methods (above) can provide qualitative information on the distribution, planform morphology and nesting of probable overdeepenings, further methods are required to extract information on overdeepening form (e.g. Fig. 2). GIS-based methods suitable for extracting such metrics from very large numbers of mapped overdeepenings are described below.

4.1 Depression in- and out-flow points

The identification of depression entry and exit points is a critical step in the measurement of depression length, elongation, and the morphology of normal and adverse slopes. In order to provide adaptability to different contexts, three approaches are presented here. The first two are topographic and hydrological methods that utilise the bed topography alone and may be appropriate where ice-thickness data are not available (e.g. in palaeo domains), although we also consider a more sophisticated hydrology based approach that makes use of ice-thickness data to infer subglacial water flow directions. The third uses only simple analysis of ice-surface elevations above the depression perimeter. All three approaches have methodological limitations, a summary of which is given in Table 2. Although methods that infer subglacial water flow may offer advantages for understanding some processes (see below), the ice-surface-based approach was considered to be the most robust (see Discussion).

4.1.1 Topography based

Where grounded ice-thickness data do not exist (e.g. continental shelves and palaeo ice-sheet domains), in- and out-flow points can be inferred by identifying broad trends in landscape elevation by means similar to the filtering method presented above (Section 3.2.2). However, thick ice sheets may become independent of topography and may subsume and dissect mountain ranges, meaning ice-flow directions can reverse as ice sheets grow. For the Byrd test domain, a 1500 km^2 moving window is required to overcome the influence of the Transantarctic Mountains and identify a general trend in relief that reflects the westwards flow of the East Antarctic Ice Sheet. By reducing the size of this moving window, more localised trends in elevation change can be extracted, which may be suitable for defining in- and out-flow points during periods when glaciation was more restricted and smaller ice masses occupied only higher elevations. However, different sectors of a single ice sheet can be characterised by contrasting styles of glaciation, meaning this method can be difficult to apply across ice-sheet scale domains.

4.1.2 Hydrology based

An intuitive approach to identifying in- and out-flow points where ice-thickness data are absent is to use ‘hydrological tools’ to calculate water flow direction and flow accumulation rasters for the DEM surface. Because ice flow in areas of deep erosion and overdeepening development is directed by topography, flow routing can be used to infer locations where maximum ice flux enters and leaves each depression. However, the assumption that subaerial water flow represents an accurate proxy for ice flow does not, as discussed above, hold true for ice sheets that subsume mountainous landscapes. If ice thickness is known, a more direct approach can be used (see 4.1.3 below), but the availability of ice-thickness data presents a further hydrology based approach, which is to infer subglacial water flow direction through calculation of hydraulic potentials (cf. Shreve, 1972) at the ice-bed interface (e.g. Livingstone et al., 2013). This method is likely to provide a better approximation of the pathways of water and sediment through an overdeepening than ice-flow-based methods, and may therefore be useful for understanding certain overdeepening-related processes and feedbacks. However, flow patterns are sensitive to changes in ice surface geometry and will not follow the deepest path through an overdeepening where the adverse slope approaches or exceeds the ponding threshold (Clarke, 2010), meaning the identification of outflow points using this method is unlikely to be robust.

4.1.3 Ice-surface based

The preferred method for identifying in- and out-flow points in this study uses the elevation of the overlying ice-sheet surface. Assuming that the surface of ice flowing immediately above an overdeepening approximates a uniform plane that slopes in the direction of flow, the points of maximum and minimum ice-surface elevation that lie above the bounding contour of the depression (Fig. 7) will provide a reasonable approximation for the principal entry and exit points in terms of the greatest flux of ice. Where multiple points of equal ice-surface elevation exist around the depression edge, a single in- or out-flow point can be determined by choosing the point most distant from the basin minima. Given the size of depression that can be mapped from a 1 km resolution DEM, the ice-surface elevation dataset is sufficiently precise to enable the identification of in- and out-flow points even for depressions in the centre of an ice sheet, and application to the test area domain demonstrates that this method is robust in most contexts (Fig. 8). The suitability of this method will, however, collapse under flat ice surfaces, such as those of ice shelves or above large subglacial lakes, and where overdeepening planform is highly complex.

4.2 Depression morphology and context

Following the identification of depression minima, bounding contours, and in- and out-flow points, a range of descriptive profiles and quantitative metrics can be extracted for each depression (e.g. Fig. 2). Many of the metrics are readily calculated using simple GIS techniques and to give a thorough description here would be unnecessary. However, some, such as the long-profile and the calculation of elongation ratio for sinuous or asymmetric depressions, require bespoke methods.

4.2.1 Long-profiles

For depressions with sinuous planforms, the path of a long profile that follows the deepest route through the depression is far removed from a straight line that joins the in- and out-flow points. A convenient solution is to calculate a 'least-cost' path between the in- and out-flow points that passes through the depression minimum (Fig. 8B). The in- and out-flow locations and position of the depression minimum can then be used to divide the profile into normal and adverse slopes. For adverse slopes with gradients that are below 11 times the ice surface gradient (Clarke, 2005), this path will approximate the route taken by subglacial water through the depression, and as noted previously this may offer some advantages. However, steeper adverse slopes will cause flow to flow around the overdeepening or distribute across the adverse slope. Consequently, hydrological methods for determining the depression long-profile were deemed unsuitable.

4.2.2 Depression shape

The shape of each mapped feature can provide important information on its probable origin and history of erosion. For example, many glacigenic geomorphological phenomena demonstrate ovoid or elongate forms (see section 5.2), meaning ovoid depressions may be considered more likely to have a glacial origin than those with more complex planform morphologies, and a glacial origin may be considered even more secure for those with ovoid planforms that are elongate in the direction of ice flow. Small, isolated and circular depressions appear to be characteristic of artefact depressions created by kriging at flightline intersections in regions of sparse empirical data, giving a 'pockmarked' appearance to the DEM surface.

Two shapes are thus of interest: circular and elongated (i.e. ovoid). Absence of elongation can be assessed using a minimum bounding geometry methodology, whereby a depression is enclosed within a polygon that is defined by its minimum possible area (Fig. 9A). Where a depression fills more than 60% of a square polygon, it can be classified as 'circular'. However, elongation cannot be assessed using this method if a depression exhibits strong sinuosity (e.g. Fig. 1B) or a complex planform. As a result, have developed a novel method for assessment of elongation in which elongation is determined with respect to the presumed direction of ice-flow at all points along the depression long-profile. This was be achieved by calculating the mean width of the depression perpendicular to the least-cost transect at regular intervals along the long-profile (Fig. 9B). Further, the threshold for elongation was defined as an elongation ratio (transect length divided by mean width) that exceeds 2. Depressions that exhibit neither circularity nor elongation in the direction of ice flow are deemed 'unclassified'.

4.3 Contextual classification

Overdeepening form and location is likely to be influenced by a range of local factors that affect erosion potential, including lithological changes or weaknesses and the location of moulins that direct surface runoff to the glacier bed (Cook and Swift, 2012; Herman et al., 2011; Hooke, 1991). Often, the simplest method for isolating such external drivers will be by cross-referencing depression location and/or relevant metrics with other numerically modelled or empirical datasets. Other factors can be isolated by automated classification of mapped depressions using such datasets. For example, empirical observations have indicated that topographic-focussing of ice flux in regions of high relief is a strong control on overdeepening location and depth (e.g. Kessler et al., 2008; Roberts et al., 2010). In this instance, a depression can be classified as 'topographically confined' using a simple proximity-based GIS-method that calculates the mean elevation of the topography within a small (20 km) buffer of the depression perimeter.

5. An assessment of the influence of overriding ice velocity on overdeepening morphology

Application of the methods outlined above to bed-topography datasets for Antarctica and Greenland (Bamber et al., 2013; Fretwell et al., 2013; Patton et al. in prep) produces a database of >13,000 bed depressions (including nested depressions). To demonstrate the potential insight that can be gained from such datasets, we explore the relationship between overriding ice velocity and overdeepening morphology. We nevertheless acknowledge important uncertainties associated with current subglacial topography datasets (see Discussion) and apply strict quality-control criteria to our mapping of bed depressions and analyses of their metrics.

5.1 Motivation

It has been proposed that many glacially moulded bedforms, including flutes, drumlins and mega-scale lineations lie on a continuum of scale that reflects the velocity of overriding ice (e.g. Clark et al., 2009; Heidenreich, 1964; Stokes and Clark, 2002). The elongation ratio (ER) in particular is assumed to correlate strongly with ice velocity for these phenomena, with analysis of high landform ER values often used to infer ice-streaming conditions (e.g. Clark, 1993; King et al., 2009; Ó Cofaigh et al., 2013; Stokes and Clark, 1999). A similar correlation may exist for overdeepenings because fast-flowing ice should enhance rates of headward erosion by quarrying and abrasion (Herman et al., 2011; cf. Hooke, 1991) and rates of sediment evacuation and abrasion at the overdeepening lip (cf. Alley et al., 2003). Proof of this relationship would have significant value for palaeoglaciological research because it would provide information on former ice velocities in regions where erosional processes have dominated or the preservation of depositional bedforms has been poor.

5.2 Quality-control procedures

Comparison of mapping results for Bedmap2 and for the higher-resolution CReSIS dataset (Fig. 10) demonstrates that the spatially variable distribution of empirical measurements in both datasets produces artefacts that are mapped as bed depressions. In the Bedmap2 topography, these artefacts mainly constitute smaller, isolated, spherical depressions that are aligned with the flightlines used to collect airborne radar measurements of ice thickness (Fig. 10C). Mapping of the same domain from the higher-resolution dataset identifies a significantly greater number of bed depressions, many of which have a similar isolated, spherical appearance, albeit at a smaller scale commensurate with the increased resolution of the dataset and density of flightlines (Fig. 10D). In contrast, the first-order characteristics of larger depressions do not differ substantially when mapped using the higher-resolution dataset, as demonstrated by planform (Fig. 10C–D) and long-profile (Fig. 10E) characteristics of the Byrd Glacier depression. Mapping from higher-resolution datasets therefore improves the detail in respect of the outlines (and thus metrics) of larger and some smaller depressions, but artefact depressions are still present.

To avoid the inclusion of spurious metrics from artefact depressions, mapping of overdeepenings from gridded datasets requires application of quality control criteria regardless of data resolution. For this study, a suite of criteria have been considered and applied that are based on known dataset uncertainties (cf. section 2.2):

1. Bed-elevation uncertainty. Absolute bed-uncertainty data beneath grounded ice is provided with the DEMs for Greenland and Antarctica (Fretwell, et al., 2013; Bamber et al, 2013). Although this is a good measure for estimating uncertainties in overdeepening absolute depth (i.e. the elevation of the deepest point in relation to sea level), it is not a robust criterion for assessing adequate delineation of mapped features, which is dependent on relative uncertainties in the immediate area of the bed. For this reason, criteria based on flightline density and depression size (below) were also considered.

2. Flightline density. Criteria were used to specify a minimum depression width in regions of sparse empirical data, resulting in the removal of small, isolated depressions characteristic of artefact depressions produced by kriging. Areas of sparse data were identified using a flightline density mask that showed the density of flightlines within a 10 km radius of each grid cell. Depressions with widths less than 20 km were excluded if they did not intersect areas with densities > 0.11, which is roughly equivalent to two flightlines within the given radius. The choice of criteria reflects the observation by Fretwell et al. (2013) that absolute errors in elevation generally increase over distances of up to 20 km, beyond which errors appear largely uncorrelated with distance.

3. Depression size. Several size criteria were employed. Firstly, large features (e.g. tectonic basins) were removed by excluding depressions with bounding contours exceeding 2,000 km in perimeter (an area equivalent to 1.5 times the catchment area of Pine Island Glacier, Antarctica; cf. (Vaughan et al., 2006). Depressions beyond this size are unlikely to have a glacial origin. Secondly, in regions with flightline densities > 0.11, depressions with adverse slopes shorter than 5 km were excluded because depressions of this size were unlikely to be adequately resolved by empirical measurements. Finally, a minimum overdeepening depth of 40 m was applied regardless of other criteria because shallow depressions are likely to have many sources, including kriging, bed elevation uncertainty, and geology. This value is intermediate between the minimum published absolute uncertainty values for the Bedmap2 (±66 m) and Greenland (±10 m) datasets (Fretwell, et al., 2013; Bamber et al, 2013).

4. Elongation with respect to the current ice-flow direction (cf. section 4.2.2). In accordance with the majority of landforms sculpted by flowing ice (e.g. flutes, drumlins, roche moutonnées, troughs), overdeepenings are generally elongate in the direction of ice flow. This criterion can therefore be applied to exclude potentially non-glacial depressions. Though some genuine overdeepenings will be excluded, including those with complex planforms formed under previous ice-flow configurations, strict filtering of landforms on the basis of ice flow direction will be beneficial for many applications because it should remove 'relict' landforms or those

with complex morphologies that are unlikely to be in equilibrium with present ice sheet processes.

5. Topographic confinement. Empirical observations indicate that overdeepenings are most common where ice flow is topographically confined (e.g. within valleys and outlet glacier troughs), meaning this criterion can be applied in conjunction with (4) to exclude probable non-glacial depressions. Notably, by favouring depressions that are located within troughs where ice flow direction is likely to have been stable over many glacial cycles, this criterion will exclude depressions with morphologies that may have evolved under varying ice-flow configurations. For this study, depressions were classified as topographically confined if the mean elevation of topography surrounding the depression exceeded the elevation of the lip by a value greater than 500 m (cf. section 4.3; Fig. 9C).

5.3 Results and interpretation

To test the relationship between overdeepening ER and ice velocity for contemporary ice sheets, our mapping methods allow the ER of Greenland and Antarctic depressions to be plotted against ice-surface velocities (Joughin et al., 2010b; Rignot et al., 2011b) measured above the deepest point in each depression (Figure 11). Because of dataset limitations and uncertainties regarding the origin and morphological inheritance of nested depressions, we restrict our analysis to 'parent' depressions that are elongated in the direction of ice flow and that pass the other quality criteria detailed above. In addition, we separately analyse the subset of depressions that can be categorised as 'topographically confined'. Not unexpectedly, these plots show substantial scatter, demonstrated by very low R^2 values, that is consistent with the limitations of the source datasets and the simplicity of our approach. Nevertheless, significance values for three of the plots are < 0.05 and therefore support a probable relationship, indicating that improvements in bed-data quality and quality control criteria will provide a stronger foundation for such a link. Furthermore, significance values are strongest for topographically confined depressions, despite the number of depressions within these subsets being significantly smaller. This is particularly evident for Greenland, where the relationship for elongated depressions is not statistically significant. The application of strict quality criteria means that the inclusion of spurious depressions arising from DEM artefacts is not thought to be a significant influence on the observed relationships.

6. Discussion

6.1 Overdeepening identification and mapping approaches

A necessity for automated mapping approaches at the scale of whole ice sheets is the development of simple but robust methods of landform delimitation (e.g. Saha et al., 2011). Overdeepenings present a considerable challenge in this respect because closed-depressions can arise from glacigenic and geological processes, though large tectonic basins (e.g. Wilkes subglacial basin) are easily excluded and smaller tectonic basins (e.g. the Vostok basin in East Antarctica) are relatively rare. An overriding concern, however, is how to define overdeepenings as mappable landforms.

For many applications, the identification and mapping of overdeepenings as 'subaerial' closed-depressions, which is a well-established definition (e.g. Fountain and Walder, 1998), would seem to be appropriate. However, because the movement of ice, water and sediment at the base of an ice sheet is driven largely by the ice surface gradient, overdeepenings mapped using our methods are not necessarily 'closed-depressions' in a subglacial context. For example, an overdeepening will only form a 'closed-depression' in the subglacial hydraulic gradient, thereby forming a lake, where the ratio of adverse slope to ice surface slope gradients exceeds the 'ponding' threshold (Clarke, 2010). Further, the 'surface' of the subglacial lake will be inclined in the opposite direction to ice flow, and the gradient of the subglacial lake 'surface' will vary in response to changes in ice-surface and hydraulic gradients during glacial advance and retreat. For purposes of understanding the formation of features that evolve over time periods that span many glacial cycles, the dependence of such glaciological thresholds on the ice surface gradient means they are unsuitable for the development of mapping criteria. Our approach may not therefore be suited to applications that aim to understand specific glaciological purposes, but, in the absence of a more appropriate definition, our approach provides an effective and robust method at the ice sheet scale.

The timescale of overdeepening formation nevertheless presents challenges for applications that seek to relate overdeepening characteristics to former or present ice sheet parameters, as we have attempted to achieve in our example test of a relationship between overdeepening ER and overriding ice velocity. First, changes in ice-sheet flow configuration associated with ice sheet and landscape evolution process that span many glacial cycles means some overdeepenings, perhaps in particular ice sheet sectors, may have formed under conditions unlike those that have prevailed in more recent ice sheet history. Second, changes in ice sheet geometry in response to shorter-term climatic variations will affect ice flow patterns and subglacial hydrological gradients, meaning the precise location of in- and out-flow points relevant over the time scale of overdeepening formation can never be precisely known. Third, it has been proposed that spatial patterns of erosion and sedimentation at the glacier bed, which are dictated by the ice surface gradient, mean overdeepening morphology should maintain equilibrium with ice sheet geometry (e.g. Hooke, 1991; Alley et al., 2003). However, the timescales required for subglacial processes to produce adjustments in overdeepening morphology in response to even slow changes in ice sheet geometry are unknown and may mean that this assumption is invalid.

In light of the complex issues described above and poor knowledge of the geology beneath the present ice sheets, our study has developed methods and criteria that utilise information on present ice sheet flow and geometry and, in particular, that focus our analysis on depressions for which a glaciological origin is most secure. This approach is particularly important for analysis of overdeepening morphology because this requires identification of overdeepening adverse and normal slopes and the exclusion of overdeepenings that do not appear to conform to present ice flow configurations. Further, the latter involves a particular focus on features beneath topographically constrained outlet glacier systems where ice-bed processes and characteristics are most likely to demonstrate equilibria with present ice geometry. The first important consideration therefore is the identification of depression in- and out-flow points, for which we utilised ice surface elevation to infer the points where maximum ice flux enters and exits each depression (section 4.1.3). Prediction of in- and out-flow points using calculated subglacial water fluxes are unsuitable because flow is sensitive to subtle changes in ice-surface gradient that may cause flow to 'pond' and deviate around the adverse slope. As we have indicated above (sections 4.1.1 and 4.1.2), without the presence of an overlying ice sheet, there will be significant uncertainty concerning ice flow direction, especially for depressions in subglacial mountain ranges and across large areas in the ice sheet interiors.

The second important consideration is overdeepening planform morphology and context because overdeepenings that do not have simple ovoid forms and that are elongate in the direction of ice flow are likely to be non-glacial or 'relict' features. These are therefore unlikely to possess morphologies that are in equilibria with ice sheet processes. Criteria applied here utilised analyses of planform shape, elongation and ice flow direction to exclude features with circular or complex planform morphologies and with ovoid morphologies that are not elongate in the direction of ice flow. Given that ice flow configurations in Greenland and Antarctica are likely to be stable over many glacial cycles, and given that assumptions concerning overdeepening planform shape are indeed valid, the application of these criteria should provide a very robust means of limiting the mapped dataset to overdeepenings that are appropriate for the investigation of overdeepening evolution and ice-erosion feedbacks. Nevertheless, because the majority of ice flow is organised into outlet glacier or ice stream systems that occupy deep troughs that extend far into the ice sheet interiors (e.g. Morlighem et al., 2014), further confidence can be achieved by means of excluding overdeepenings that are not confined by step topography. For this study, the high relief threshold used to define topographically confined overdeepenings (section 4.3) further limits the dataset to overdeepenings in deep troughs where ice flow configuration is likely to be highly stable.

6.2 Relevance and scientific potential

The compilation of glacial-landform datasets alongside improvements in the resolution of remotely sensed data has led to numerous and detailed analyses of subglacial phenomena across palaeo-glaciated domains, including drumlins (Clark et al., 2009), glacial lineations (Greenwood and Clark, 2009; Spagnolo et al., 2014; Stokes et al., 2013), meltwater channels (Margold et al., 2011) and ribbed moraine (Dunlop and Clark, 2006). Much of this work has been driven by the need to decipher the glaciological significance of such landforms, as well as the need to obtain robust morphological data to enable testing of numerical models that simulate ice sheet and landscape evolution processes (e.g. Evans, 2009; Jamieson et al., 2010; Melanson et al., 2013). For overdeepenings, process understanding has been disadvantaged by an absence of quantitative studies, meaning characterisation of overdeepening distribution and morphology falls far behind that achieved for comparable glacial landforms, including cirques, valleys and drumlins. Given the strong influence exerted by reverse bed slopes on subglacial hydrology, ice-flow dynamics and ice-sheet stability (e.g. Cook and Swift, 2012), the acquisition of similar data for understanding of overdeepening form and evolution is important. Moreover, comparison of overdeepening morphology with ice sheet parameters has the potential to yield major advances.

In terms of ice sheet behaviour, overdeepenings are known to have potentially far-reaching implications for ice, water and sediment movement. First, a large body of empirical (e.g. Iverson et al., 1995) and theoretical work (e.g. Röthlisberger and Lang, 1987; Creyts and Clarke, 2010) demonstrates that overdeepenings raise subglacial water pressures by reducing the transmissivity of the subglacial drainage system, thus modify important boundary conditions that affect ice-bed coupling and rates of basal sliding. Second, the amplitude and wavelength of topography at the ice sheet bed is of fundamental significance to ice-sheet sliding laws (Schoof, 2005). Third, the significance of overdeepenings for rapid ‘collapse’ of marine terminating outlet systems during forced retreat has been demonstrated theoretically (Schoof, 2007) and empirically (e.g. Nick et al. 2009), although the backstress provided by reverse slopes indicates that overdeepenings beneath grounded ice may have a stabilising influence on ice sheet flow. Given the availability of ice velocity data and knowledge of subglacial topography, the possibility that overdeepenings promote or resist ice flow according to the specific morphology of the adverse slope (cf. Cook and Swift, 2012) is therefore eminently testable. Further, a large dataset of overdeepening characteristics enhances the potential to gain process-based insights by stimulating development and testing of numerical ice-erosion models (e.g. Egholm et al., 2012).

Fundamental questions also remain in terms of overdeepening origin and evolution. Cook and Swift (2012) propose that patterns of erosion and deposition that are dictated by the ice surface gradient (cf. Hooke, 1991; Alley et al, 2003) mean all glacier and ice sheet beds should tend toward a uniformly overdeepened long-profile, with deepest erosion occurring near the long-term average equilibrium line altitude (ELA) (e.g. Hallet et al., 1996; Boulton, 1996; Anderson et al., 2006). However, observations indicate that beds with multiple overdeepenings are common, indicating that ice-water-sediment-erosion (IWSE) feedbacks produce very localised deep glacial erosion by reinforcing patterns of erosion and sedimentation (e.g. Hooke, 1991; Alley et al., 2003) and thereby act as an important constraint on overdeepening size and morphology (Cook and Swift, 2012). The ability to map overdeepenings and compare quantitative information on their morphology with information on ice sheet data offer significant potential to test assumptions relating to the controls on overdeepening location and form. In addition, quantitative measurements are again necessary to motivate and test numerical ice-erosion models. Notably, parameterisation of IWSE processes in numerical ice-erosion models that couple the flow of ice, water and sediment is now being realised (e.g. Egholm et al., 2012), and increasingly sophisticated treatments of the ice flow (Egholm et al., 2011), erosion (e.g. Iverson, 2012) and water flow (e.g. Werder et al., 2013), promising unprecedented ability to explore the implications of coupled IWSE processes on the evolution of subglacial landscapes.

6.3 Potential application to overdeepening form

Exploration of the formation and evolution of overdeepenings under contemporary ice sheets offers significant advantages. First, the presence of ice-cover means landform morphology in areas of deep erosion is not obscured by postglacial sedimentation, which complicates analyses of overdeepenings and other erosional features, including tunnel valleys, in palaeo settings (Huuse, 2000; Hansen et al., 2009; Preusser et al., 2010; Moreau and Huuse, 2014). Second, the data that are available that describe ice sheet characteristics, including velocity and thermal regime, are vastly superior to that which can be obtained for former ice sheets, for example from numerical modelling. Third, the present ice sheets have remained largely stable features for much of the recent geological past (e.g. Huybrechts, 1993), meaning landscapes and landforms shaped by characteristically slow subglacial processes, and the ice sheets themselves, are more likely to have achieved equilibrium forms that are in balance with climatic, glaciological and tectonic forces. Further, the proposed ability of overdeepening form to maintain equilibrium with ice geometry (see section 6.1) means the existence of subglacial sediments within overdeepenings does not necessarily preclude the ability to obtain process insights.

Exploration of the full range of potential applications of our methods is beyond the scope of this paper. Rather, as a single illustration of the potential power of our approach, we have tested for the potential relationship between overdeepening ER and overriding ice velocity. It is nevertheless recognised that evidence for such a relationship is not strong, and this motivates consideration of data and methodological limitations that similar studies must address.

Whilst the limitations of available data are considered in the section below, methodological limitations concern the validity of the proposal relationship and the ability of the data to provide a robust test. The validity of the relationship is supported by analogy with a wide range of glacigenic landforms and the statistical significance shown by the plots (Figure 11). Nevertheless, substantial scatter could indicate a variety of confounding factors. First, numerous factors are likely to play a role in overdeepening morphology. For example, it is likely that elongation ratio is influenced by valley width, meaning width is constrained in a way that length is not, whilst erosion processes do not necessarily scale linearly with ice velocity (cf. the likely influence on quarrying rates of subglacial water pressure variation; e.g. Hooke, 1991; Egholm et al., 2012). Further consideration also needs to be given to the significance of 'nested' depressions, which may indicate several different processes and controls. Second, the likely timescales of overdeepening formation mean measured ice velocity is not necessarily indicative of mean ice velocity over the timescale of overdeepening formation. Third, the complexity of subglacial landscapes indicates that filtering methods that attempt to exclude depressions that are unrelated to present ice flow configurations may require refinement. For example, limiting the mapped dataset to elongated and topographically confined depressions, which intends to focus analysis on overdeepenings in fast-moving outlet glacier systems that extend inwards from the ice sheet margins, will not necessarily exclude overdeepenings in buried mountain ranges in the interior of the East Antarctic Ice Sheet that were formed under warm, early ice sheet conditions (e.g. Bo et al., 2009?) where ice is now dominantly cold-based. Finally, the role of adverse slopes in providing backstress and elevating basal water pressures means overdeepenings may accelerate or modulate velocity depending on the specific morphology of the adverse slope.

These limitations and others are likely to be relevant to study of other aspects of overdeepening form, including adverse slope morphology and its significance. Notably, sparse data on subglacial process rates and sediment fluxes mean it is impossible to have confidence that overdeepening morphologies are able to maintain equilibrium with ice sheet processes. For Antarctica in particular, slower rates of subglacial erosion and sediment transport may result in overdeepening morphologies remaining in permanent disequilibrium. In addition, modelling evidence indicates rapid landscape evolution during initial glacial cycles, followed by relative landscape stability (e.g. Jamieson et al., 2010; Kessler et al., 2008), meaning long-lived erosional features that include overdeepenings reflect conditions that prevail early in ice sheet history and are not typical of the present day. Finally, we recognise that different mapping approaches may be required where it is necessary to understand the importance of specific subglacial processes. For example, the subglacial lakes prevalence and morphology reflects 'closed depressions' in the subglacial hydraulic gradient (e.g. Livingstone et al., 2013?) that are qualitatively and quantitatively different from their topographic counterparts.

6.4 Data limitations, quality control, and methodological recommendations

The quality of mapping results and the process-based insights that can be obtained using associated empirical data are dependent on the resolution and quality of available bed topography data. In addition, artefacts may be introduced during interpolation of raw bed-elevation measurements. For mapping and analysis of overdeepenings, automated methods can be used to exclude artefact depressions, but these are not without problems.

A major limitation, particularly for Bedmap2, is the absolute uncertainty of large swathes of the subglacial topography. For some areas of the East Antarctic interior, absolute bed-elevation uncertainties range up to 1,008 m (Fretwell et al., 2013). Conversely, for the most recent DEM covering Greenland (Bamber et al., (2013), some of the largest errors occur in the mountainous coastal fjord regions where extrapolation, rather than interpolation, has been required to resolve bed elevations. These fjord regions are prime locations for overdeepening development, where topographic confinement dominates the configuration of outflowing ice. Nevertheless, comparison of features mapped from large-scale datasets with those from higher resolution subset domains (e.g. Fig. 10) reveals an encouraging level of consistency. Notably, for the Byrd Glacier depression (Fig. 10 C–D), long-profile form (Fig. 10E) is resolved relatively well by Bedmap2 in comparison to the higher-resolution product, with only minor differences in maximum absolute depth. On the other hand, the size, form and number of smaller depressions in the Byrd Catchment (Fig. 10C-D) is very much influenced by flightline density (Fig. 10A–B), and artefact depressions created at flightline intersections by the interpolation method (i.e. kriging) are evident at both dataset resolutions. Quality control criteria that remove such depressions, as applied in the example study above, are therefore necessary regardless of resolution. Acquisition of even higher-resolution datasets using novel extrapolation approaches (e.g. Morlighem et al., 2014) offers the potential to further understand and improve bed-elevation uncertainties, but kriging is still required where ice flow velocities are low and empirical ice thickness measurements are scarce. The scarcity of empirical measurements across large areas of the present ice sheets means therefore that significant areas of uncertainty will remain.

Morphological studies of subglacial phenomena thus require strict appreciation of the uncertainties inherent within the source datasets. The application of multiple quality criteria to mapped results, such as minimum flightline density, is viewed as essential to minimise the introduction of unreliable or spurious data. As such, the limitations of existing subglacial topography datasets mean that the subglacial area that is suitable for landform analyses of the kind presented here is only a fraction of the total area (e.g. only 36% of the grounded Antarctic bed is constrained by measured data at a 5 km resolution in Bedmap2). Formerly glaciated areas on the adjacent continental shelves offer potential to greatly increase the mappable area: for example, studies in

Antarctica show that glacial and post-glacial sedimentation in offshore areas may be only 4-5 m thick (e.g. Dowdeswell et al., 2004) and is therefore well below the minimum bed elevation uncertainty in subglacial areas. However, detailed sediment thickness data for offshore areas is available for only restricted areas, and there is limited potential to relate landform location and morphology to ice sheet parameters. In addition, fjord depths are poorly constrained along much of the Greenland coastline, meaning subglacial and offshore topographies are often mismatched by as much as several hundred metres (Bamber et al., 2013). Further strategic data collection is therefore required to address areas of uncertainty, both in the interiors of ice sheets and at present ice-sheet margins.

7. Conclusions

The publication of relatively high-resolution bed-topography datasets covering the entire Antarctic and Greenland ice sheets (Bamber et al., 2013; Fretwell et al., 2013) provides a major opportunity to advance our understanding of the evolution of subglacial landscapes and relationships or feedbacks between ice sheet and landforming processes. Overdeepenings are a key landform in this respect because of the lack of consensus on their origin and their far-reaching influence on critical ice-bed processes and ice-sheet behaviour (cf. Cook and Swift, 2012). To address the need for quantitative data on overdeepening characteristics, we have explored automated methods for mapping glacial overdeepenings and the extraction of metrics the describe their form. Our main methodological findings can be summarised as follows:

- Hydrological tools and terrain filtering methods fail to adequately capture the complex morphologies of overdeepenings, primarily because terrain filtering lacks precise thresholds required for delimiting closed-depression boundaries and because both methods lack the ability to resolve depression nesting.
- A novel rule-based GIS method has been proposed that quantitatively tracks changes in the length of closed-contours from initial points of elevation minima. This method provides consistent, morphologically based mapping results and is computationally efficient at ice sheet scales. Its application is therefore not dependent on a particular bed-dataset resolution, requires no abstract threshold parameters to be defined, and is unlikely to be restricted or compromised by anticipated improvements in dataset quality and detail.
- The limitations of available datasets mean that mapped features require robust scrutiny. A suite of simple quality control criteria that are applicable to the 1 km gridded datasets for Greenland and Antarctica have been proposed that may be adapted according to the nature of the bed-topography dataset or the specific focus of a given study.

In addition, we have illustrated our method by testing a proposed relationship between overdeepening elongation ratio and overriding ice velocity for a large sample of depressions mapped beneath the Antarctic and Greenland ice sheets, and discuss the potential insights and limitations that are demonstrated by such an approach. We conclude that:

- The ability to relate overdeepening characteristics to present ice sheet characteristics indicates significant potential to gain insight into critical ice-bed processes, including those that influence the location, timescale and nature of overdeepening formation, and the co-evolution of ice sheets and their subglacial topographies. However, the limitations of present datasets and the simplicity of our approach mean that strong statistical relationships cannot necessarily be expected.
- Quality criteria are necessary when analysing subglacial topography datasets. Criteria can avoid the inclusion of spurious depressions arising from DEM artefacts and address limitations are associated with the density and accuracy of empirical bed-elevation measurements. However, the level of insight that can be acquired is dependent upon the quality of the metrics that such data permit.
- Qualified relationships between overdeepening ER and ice velocity were found and, given anticipated improvements in dataset accuracy and resolution, further work is needed to better resolve them. Possible confounding factors include the existence of additional controls on overdeepening length or width and the long timescale of overdeepening formation relative to that of climatic changes that influence present ice velocities.
- The significance of reverse bed-slopes in glacial systems (Cook and Swift, 2012) indicates that the introduction of overdeepenings into ice sheet beds is itself expected to modulate ice velocities, meaning overdeepening metrics may demonstrate non-linear relationships with ice velocity that reflect a more fundamental process of ice sheet co-evolution with the underlying topography.

This work demonstrates that mapping and analysis of even the largest subglacial landforms, including troughs, cirques and overdeepenings, will be an exercise marked with varying aspects of uncertainty that requires strict quality control procedures and close scrutiny of mapping and metric outputs. Nevertheless, anticipated improvements in the accuracy and resolution of bed-topography datasets, including novel extrapolation methods that utilise surface ice velocities, will reduce the need for quality control procedures and achieve convergence of measured landform attributes on ‘true’ values that will facilitate increasingly robust insights from empirical data. We encourage collaboration between the geomorphological, glaciological and numerical modelling communities to identify real and simulated landscape features and IWSE feedbacks that maximise the potential to test simulated landscapes and to expedite process understanding.

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Table captions

Table 1: Merits of alternative methods for identification of in- and out-flow points of individual enclosed-depressions. Methods are listed in order of preference.

Figure captions

Figure 1: Examples of overdeepenings beneath contemporary and palaeo ice sheets. (A) Subglacial Lake Ellsworth in East Antarctica (Ross et al., 2011) occupies an overdeepening in a major subglacial trough that cross-cuts the Ellsworth subglacial mountains (figure courtesy Neil Ross). (B) The post-glacial lake Veitastrondvatnet, shown in an oblique aerial view looking due SW, occupies an overdeepening in a sinuous trough confined by the steep topography of the Sognefjord region, Norway. The lake is approximately 17 km long (image: Google Earth). (C) Trough-floor profiles for Jakobshavn Isbrae, Greenland, and Recovery Glacier, Antarctica, derived from bed topography datasets (Bamber et al., 2013; Fretwell et al., 2013) exhibit numerous overdeepenings. Colours highlight reverse-bed slopes: red indicates slope gradients that exceed the supercooling threshold (cf. Alley et al., 2003); blue indicates slope gradients that exceed the ponding threshold (cf. Clarke, 2005). (D) Bathymetry of the bedrock basin occupied by post-glacial Lago Fagnano, Tierra del Fuego, derived from high-resolution single-channel seismic data, showing numerous ‘child’ basins in black nested within the ‘parent’ overdeepening (here defined by the present lake margin). Glacial erosion has been conditioned by tectonic processes and the location of numerous faults. MF: Martínez fault; RT: Río Turbio fault; RC: Río Claro fault. Figure modified from Esteban et al. (2014) .

Figure 2: Cartoon showing the long-profile of a subglacial ‘parent’ depression containing nested ‘child’ depressions. Various metrics that can be used to describe the form the depression and associated child depressions are defined that can be extracted using GIS-based techniques.

Figure 3: Potential methods for identifying overdeepenings from the subglacial Bedmap2 DEM: A) the application of hydrological tools by means of ‘filling’ sinks (i.e. closed-depressions). The

inappropriateness of this method is highlighted by the application to differently sized domains: the visible boxed domain (red) and the entire Bedmap domain (blue). B) The application of 'terrain filtering', using a low-pass, circular (200 km) Gaussian filter and a standard deviation value of 25. Negative residuals (original elevation values minus filtered elevation values) ≤ -450 m are draped over the original Bedmap2 topography.

Figure 5: Identification of elevation minima. (A) Initial Bedmap2 DEM; (B) quantitative terrain analysis mask identifying areas of depression-like topography (see text); and (C) points of elevation minima within enclosed depressions (closed contours) contained by the terrain analysis mask.

Figure 6: Identification of enclosed-depression perimeters and their nestings using the contour tracking method. (A) Linear transects drawn from each elevation minima intersect with contours. The length change between adjacent contours is calculated to identify abrupt increases in contour length that indicate that a contour is beyond the contour that defines the closed-depression perimeter (see text). (B) Parent and child depressions are identified using multiple-passes of the contour-tracking algorithm and using a threshold contour-length increase of > 190% (break of slope) to identify the closed-depression boundary. (C) Parent and child depressions classified using a top-down approach.

Figure 7: Cartoon illustrating the use of the ice-surface elevation data to identify of overdeepening in- and out-flow points for an elongate overdeepening oriented in the direction of ice flow. Topographic focussing of ice flow into the overdeepening means these points are suitable proxies for the principal overdeepening entry and exit points in terms of the greatest ice flux.

Figure 8: Identification of overdeepening in- and outflow points using ice surface elevation data (cf. Fig. 7) and overdeepening long-profiles. (A) Ice-surface elevation (coloured scale) draped over the Bedmap2 subglacial topography (hillshaded DEM beneath the colour). (B) In- and out-flow points of parent depressions identified from (A) and long profiles, comprising adverse and normal slopes, calculated using a 'least-cost' routing analysis (see text).

Figure 9: Depression-shape classification methods and results. (A) Parent depressions enclosed by a polygon representing the smallest rectangle possible by area. (B) A magnified view of the depression boxed in red in (A) showing the bounding rectangle and transects (red lines oriented perpendicular to the least-cost path) used to calculate depression mean width. (C) Example classification output.

Figure 10: Comparison of subglacial topography, flightline density, and mapping outputs for Bedmap2 versus a higher resolution dataset (CReSIS; see text) covering a sector of the Byrd Glacier catchment. (A–B) Radar flightline tracks (black) and interpolated subglacial topography (see legend in B). (C–D) Mapping results, showing differences in the shape, size and number of parent and child depressions as a consequence of interpolation of data from contrasting flightline densities to produce datasets with contrasting resolutions. (E) Long-profiles for transect A–A' in (C) and (D).

Figure 11: Example application of data on overdeepening form, showing overdeepening elongation ratio versus ice-surface velocity above the overdeepening minima for (A) Antarctica and (B) Greenland. Red points are overdeepenings that are elongated in the direction of ice flow; green points are the subset of these overdeepenings that have been classified as topographically confined. Values for best-fit regression lines show the significance of the regression relationship; R^2 values are ≤ 0.03, reflecting considerable scatter that is discussed in the text.