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Published in:
Boreas

Link to article, DOI:
10.1111/bor.12199

Publication date:
2017

Document Version
Peer reviewed version

Ice margin and meltwater dynamics during the mid-Holocene in the Kangerlussuaq area of west Greenland

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Carrivick, J.L., Yde, J., Russell, A.J., Quincey, D.J., Ingeman-Nielsen, T., and Mallalieu, J.: Ice margin and meltwater dynamics during the mid-Holocene in the Kangerlussuaq area of west Greenland. *Boreas doi xxxxxx*

Land-terminating parts of the west Greenland ice sheet have exhibited highly dynamic meltwater regimes over the last few decades including episodes of extremely intense runoff driven by ice surface ablation, ponding of meltwater in an increasing number and size of lakes, and sudden outburst floods, or 'jökulhlaups', from these lakes. However, whether this meltwater runoff regime is unusual in a Holocene context has not been questioned. This study assembled high-resolution topographic data, geological and landcover data, and produced a glacial geomorphological map covering ~1200 km^2. Digital analysis of the landforms reveals a mid-Holocene land-terminating ice margin that was predominantly cold-based. This ice margin underwent sustained active retreat but with multiple minor advances. During ~1000 years meltwater runoff became impounded within numerous and extensive proglacial lakes and there were temporary connections between some of these lakes via spillways. The ice-dams of some of these lakes had several quasi-stable thicknesses. Meltwater was apparently predominantly from supraglacial sources although some distributary palaeochannel networks and some larger bedrock palaeochannels most likely relate to mid-Holocene subglacial hydrology. In comparison to the geomorphological record at other northern Hemisphere ice sheet margins the depositional landforms in this study area are few in number and variety and small in scale, most likely due to a restricted sediment supply. They include perched fans and deltas and perched braidplain terraces. Overall, meltwater sourcing, routing and the proglacial runoff regime during the mid-Holocene in this land-terminating part of the ice sheet was spatio-temporally variable, but in a manner very similar to that of the present day.

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Changes in terminus position, mass and dynamics of land-terminating outlet glaciers in west Greenland have major implications for ice sheet stability and via meltwater fluxes for global sea level rise. A key to understanding the driving mechanisms of dynamic changes has been separating short-term variability from longer-term trends, in air temperatures, ice sheet surface melt, and outlet glacier velocity, for example. Specifically, analyses of remotely-sensed images of the ice surface have enabled surface meltwater generation (e.g. Harper *et al.* 2012), temporary storage in supraglacial lakes (e.g. Fitzpatrick *et al.* 2014), and the implications for subglacial meltwater dynamics (e.g. Bartholomew *et al.* 2012) and glacier velocity (e.g. van de Wal *et al.* 2015), to be interpreted over the last decade. Mernild *et al.* (2012) have compared modelled variability in meltwater runoff to that measured, and Carrivick & Quincey (2014) have analysed variability in the number and size of ice marginal lakes along the entire western margin. However, even the most long-term of these studies is limited to the satellite era and to the duration of field campaigns, i.e. over the last ~ 45 years at most and usually concentrated in the last decade. There is therefore a need to utilise longer-term datasets, such as glacial geomorphology, to place modern observations of land-terminating ice margin position fluctuations and the regime(s) of meltwater from these in the context of longer-term (Holocene) ice sheet margin character and behaviour.

Previous research on Holocene glacial geomorphology associated with the land-terminating margins of the ice sheet in west Greenland has been motivated to establish a geochronology, i.e. identifying and absolute dating of major (Last Glacial Maximum and Holocene) moraines, and improving the resolution/confidence of these dates. Major efforts have concentrated on evidence pertaining to the early Holocene in the Qeqertarsuaq (Disko Island) – Disko Bugt area (e.g. Donner & Jungner 1975; Ingólfsson *et al.* 1990; Humlum *et al.* 1995; Long & Roberts 2002; Lloyd *et al.* 2005, Long *et al.* 2006; Young *et al.* 2013), or on terrestrial evidence pertaining to the middle Holocene in the Sisimiut-Kangerlussuq area (e.g. Ten Brink & Weidick 1974; Ten Brink 1975; van Tatenhove *et al.* 1996; Levy *et al.* 2012). In recent years the use of cosmogenic surface exposure dating has added additional information to the chronology of the Late-Wisconsin and Holocene deglaciation history (e.g. Rinterknecht *et al.* 2009; Roberts *et al.* 2010). Overall, whilst major moraine systems spanning ~ 120 km of landscape have been mapped in west Greenland (Ten Brink 1975), there is a paucity of 'high-resolution' glacial geomorphological mapping and thus of complementary detailed information on the former ice areal extent, thickness, flow patterns and behaviour. By far the most notable exception is the seminal work of Ten Brink (1975) who made detailed investigations of the major moraine systems, glacial geomorphology and Holocene history of the Sukkertoppen – Kangerlussuaq – Ørkendalen (Qinnguata Kuussua) area but who was limited by glacial deposits that were < 50 m in local relief (Ten Brink 1975) and, in general 'a thin drift with little topography' (Ten Brink, 1975). Warren & Hulton (1990) made a glacial geomorphology investigation south of Ilulissat (Jakobshavn) and interpreted a topographic control (rather than climate) on still-stands of a retreating tidewater outlet glacier.

The aim of this study is therefore to critically analyse the landform record of ice margin and meltwater activity during the mid-Holocene in west Greenland. This is achieved via construction of a high-resolution glacial geomorphological map, taking advantage of digital terrain analysis where possible, in combination with geological data, landcover data and field observations. We focus on the Kangerlussuaq area because of its accessibility and its prominence in ongoing research into the nearby glaciers, rivers, geology, soils and vegetation.

Study site

The Kangerlussuaq area (Fig. 1) has an arid, continental low-arctic climate (Mernild *et al.* 2015). The bedrock in the area is predominantly of Archaean ortho-gneisses that were reworked under high grade metamorphism in the palaeo-proterozoic (Van Gool *et al.* 2002). The bedrock mainly has a dip of 70˚ east and there are several major faults and joints extending through the area (Aaltonen *et al.* 2010). The large-scale topography (Fig. 1), which is primarily controlled by these faults and joints, has been further exaggerated by glacial erosion that formed U-shaped valleys and produced areal scouring (Sugden 1974), and subsequently post-glacial faulting that provides a lineation predominantly in the east-west direction (Aaltonen *et al.* 2010). The meso-scale topography is typical of area scouring by subglacial erosion of crystalline bedrock and includes dome-like summits, rounded ridges, smoothed floors of incised cols, and streamlined 'stoss and lee' bedrock hummocks (Sugden 1974; Ten Brink 1975). Valley fill deposits at the fjord heads include major terraces that are composed of both glacifluvial and marine sediments (e.g. Storms *et al.* 2012).

Between 10 000 and 11 000 cal. a BP the ice sheet margin in west Greenland began to retreat inland from a position near the present coastline (Funder 1989). The local marine limit is likely to be slightly higher than 40±5 m a.s.l. (Storms *et al.* 2012) and this is important note in this study because any shorelines much above this altitude are thus related to palaeolakes. The general retreat was punctuated by numerous still-stands, or perhaps minor readvances, which left a series of large-scale and near-continuous sinuous and lobate major moraine systems across west Greenland, as classified, dated and named by Ten Brink & Weidick (1974), Ten Brink (1975), van Tatenhove *et al.* (1996), and Levy *et al.* (2012) (Fig. 1). Further insights into the geochronology and landscape evolution of the Kangerlussuaq-Russell Glacier area have been gained by analyses of lake sediments (Eisner *et al.* 1995; Aebly & Fritz 2009; Anderson & Leng 2004; Young & Briner 2013, 2015) and valley fill sediments (Storms *et al.* 2012). The most recent consensus is that the Fjord moraine system (not shown on figure 1) dates to 8340 – 9080 cal. a BP, the Umîvît moraine system to 7360 to 7960 cal. a BP, the Keglen moraine system to 6490 to 7190 cal. a BP, and the Ørkendalen moraine system to 6400 to 7030 cal. a BP (Storms *et al.* 2012, Fig. 1). The landscape in the area of this study (Fig. 1) therefore represents the mid-Holocene and <1000 years of evolution during deglaciation.

Between 4200 and 1800 cal. a BP the ice sheet margin was (an unknown distance) farther inland from the present position (Young & Briner 2013, 2015), but at Russell Glacier a neoglacial advance at 2000 cal. a BP closely corresponds in position to Little Ice Age moraines and to the position of the present ice margin (Forman *et al.* 2007).

Concomitant with this general ice margin retreat, the Kangerlussuq – Russell Glacier area landscape has developed as a geomorphological record of that activity. Ten Brink (1975) suggested that excellent landform preservation in the area is due to the fact that in this region the ice advanced up-slope. Additionally, the semi-arid climate and generally stable bedrock (Aaltonen *et al.* 2010) means that mass movements are limited in type and number and frequency and mostly to river banks. Furthermore, surface water drainage is limited so major channels pertain to different environmental conditions than at present. Consequently, soil development has been slow and thus soil cover is very thin (Ozol & Brull 2005) permitting bedrock form to remain visible especially via aeolian deflation. Vegetation cover is limited and so where sediments and depositional landforms do exist their morphology is clear and their bulk composition can be relatively easily inferred from natural exposures such as aeolian deflation cusps and hollows.

Datasets and methods

In this study we obtained and corrected fine-resolution (2 m grid) topography data, compiled existing geological information and created our own landcover (30 m grid) classification. These datasets and our methodology are described and explained in the Supporting Information. These datasets in combination and with our own field studies enabled creation of a digital geomorphological map of the Kangerlussuaq – Russell Glacier area. Our scrutiny of the landforms included development of novel digital analyses.

Glacial geomorphology

Our glacial geomorphology map was compiled by firstly creating a geodatabase of existing geomorphological data, most notably the 'major moraine systems' originally mapped by Ten Brink (1975) but developed in terms of chronology by van Tatenhove *et al.* (1996) and Storms *et al.* (2012). None of these previous efforts completely cover the area considered in this study (Fig. 1) and none at a fine-resolution. Additionally, a geomorphological assessment of part of the Leverett Glacier ice margin by Scholz & Baumann (1997) was used, as were Little Ice Age (LIA) and neoglacial landforms by Forman *et al.* (2007), some parts of the Ørkendalen moraines by Levy *et al.* (2012), and aeolian landforms by Willemse *et al.* (2003). Lidberg (2011) reported a wealth of field photographs, description and reasoned interpretation of the glacial geomorphology of a part of the area considered in this study. Secondly, we primarily utilised our topography data set (Fig. 1), but with reference of that to our geological (Fig. 2) and land cover datasets (Fig. 3), to extend and infill this previous mapping with our identification of glacial geomorphology. Mapping was supported by field observations and the authors' own oblique aerial photographs both spanning multiple years from 1985 to 2015.

Mapping of glacial geomorphology focussed on remotely identifying:

(i) Moraine ridges to interpret the position and probable style of ice margin advances (or still-stands but they are far less likely in this area: Ten Brink 1975) of the ice margin.

(ii) Till/drift mantle/veneer distribution, and kame and kettle topography (Ten Brink 1975) so as to refine suggestions of active ice margin retreat versus ice stagnation – disintegration, respectively.

(iii) Palaeochannels to interpret past meltwater routing and style.

(iv) Spillways to identify major meltwater routes between adjacent valleys.

(v) Shorelines to determine the presence and extent of former lakes and the likely dam type or associated ice margin configuration.

(vi) Perched fans and deltas, braidplains and marine terraces, as marked by sloping and horizontal terrace edges, respectively, to identify changes in meltwater-sediment discharge regime and/or changes in base level.

For each of these categories criteria including position and association, shape and size and texture were used to discern landforms (Table 1). Since this study encompasses such a large area that is relatively remote and inaccessible, virtually all of our geomorphological information is derived from the fine-resolution topography and from superficial landform character as represented in the topography, geology and landcover maps. This meant that assessing whether a particular ridge or hummock was bedrock or sedimentary depended on multiple enquiries. Firstly, its context (what else was in its vicinity) was considered. For example multiple gully heads at a similar elevation on a hillslope could indicate a change in substrate hardness. Secondly, its form was considered. For example bedrock hummocks in the area tend to have at least one steep side, owing to the geological inclination and dip, whereas especially older moraine ridges tend to be much more subdued forms. Thirdly, its texture was informative. Comparison between the DEM and the landcover map showed that grasses and sedges appeared 'smooth' on the DEM and indicated soil and thus sediments, whereas bare rock was visually rough. Bedrock areas could also be detected with geological structure (faults, cracks, lineations), 'stoss and lee' hummocks and superimposed isolated boulders.

Use of these criteria (Table 1) was aided by digital topographic analysis, such as interactively taking vertical transects off the DEM, analysis of departures in elevation from a local trend to quantify local landform relief, visual assessment of hillshaded images especially for surface texture, for example (Fig. 4A, B, C, D, respectively), 3D visualisation (Fig. 5) and field photographs (Fig. 6). Regarding detrending, we generally used the ArcGIS 10.2 tool 'focal statistics' (with calculation of 'mean' elevation in a circular moving window) because the typical width of moraine ridges in the area is 40 to 60 m and the typical width of palaeochannels is 10 to 20 m. Palaeolake shorelines, which in this study area do not contain any record of glacio-isostatic uplift, were digitally drawn using the 'create contour' function in the 3D Analyst extension of ArcGIS 10.1. Terraces were mapped and queried for elevation at the boundary/edge between the sloping riser and the relatively flat tread morphological units. Horizontal terrace edges identified incised fluvial floodplains/braidplains and incised marine terraces, and sloping terrace edges identified incised glacifluvial fans and deltas. Variations in spatial density, orientation and geometry were computed via export of landform centroid coordinates and attributes to a text file.

Results and interpretation

A total area of ~1200 km^2 was mapped for its glacial geomorphology and included 2076 'moraine' polygons, 428 palaeochannels, 261 terrace edges, 24 deltaic- or fan-shaped landforms and several streamlined bedrock hummocks (Fig. 7). A .pdf version of this map enabling zoom and pan functions and with layers that can be switched on and off is available as Supporting Information (Fig. S7). Moraine ridges on plateau areas and those that cross valley floors (Fig. 5A) are generally aligned north-south, i.e. transverse to regional slope, and comprise sinuous ridges typically with 5 m local relief (Fig. 7). They are also sparsely-spaced; typically the distance between the ridges is approximately ten times the width of the ridge (Fig. 5A, D). Moraine ridges on valley sides often occur in sub-parallel stacks and with direct association of minor palaeochannels (Fig. 5A, 7), so are most likely moraine-kame terrace complexes (Weidick 1968; Ten Brink 1975) reflecting re-working by meltwater of moraine during progressive ice margin retreat and thinning. Overall, the moraines mapped in this study exhibit considerable complexity in planform and have asymmetry in position across local valleys and across adjacent hillsides and local plateaux (Fig. 7). The complexity and asymmetry of these moraine ridges and their occurrence both within and between the previously identified 'major moraine complexes' means that only local (e.g. for a single valley), not regional ice margin retreat patterns are distinguishable in our mapping (Fig. 7).

Nonetheless, on an individual landform basis the moraines mapped in this study closely correspond to those mapped by Ten Brink (where there is overlap with his Plate 2 and our study area) but in general we identify at least an order of magnitude more ridges. Discrepancy in position between our moraines and Ten Brink's (1975) could be due to Ten Brink's use of aerial photographs that had not been orthorectified, as hinted at by the 'approximate scale' label on his maps. Discrepancy between the number of moraine ridges that we identify and those by Ten Brink (1975) is due partly to the higher resolution of the data we have to hand, and partly due to the rigid criteria imposed by Ten Brink (1975) that only moraines that extended continuously for several kilometres and only those in similar topographic positions on both sides of valleys were included.

Furthermore, whilst Ten Brink was motivated to identify large-scale ice margin advances, we are interested in revealing the detail and complexity of ice margin dynamics, and in particular the nature of meltwater at the ice margin. With this interest in mind, and targeting where local cross-cutting relationships or else direct contact between moraines and meltwater landforms exist, we describe in detail four sub-areas of our map, informally named here as the 'western spillway', the area around Aajuitsup Tasia, a gorge emanating from the Leverett Glacier proglacial area, and the Ørkendalen valley (Fig. 7). The central part of our study area immediately to the north of Aajuitsup Tasia contains evidence for multiple palaeolakes, multiple shorelines of these palaeolakes, and exchange of water between them via spillways. The southern part including the Sandflugtdalen (Akuliarusiarsuup Kuua) and Ørkendalen (Qinnguata Kuusua) valleys (Fig. 1) includes recessional moraines and extensive suites of ice marginal channels. For brevity we only describe herein the geomorphology of the four sub-areas that we then go on to discuss.

Western spillway

The 'western spillway' (Fig. 8) exits into a local valley known as Ringsødal and comprises a 40 to 60 m deep and ~250 m wide gorge with sub-linear planform and a v-shaped cross-section. The floor of the gorge at 295 m a.s.l. contains bedrock hummocks with a streamlined planform and steep (cliff) sides. A major palaeolake fed into this spillway as evidenced by a very distinct shoreline at 340 m a.s.l (Fig. 8). This shoreline is altitudinally far above the local marine limit and is open-ended meaning that the dam for the lake water no longer exists. For the lake to form a shoreline 45 m above the spillway floor, the spillway must not have existed during lake formation. A moraine dam could have existed at the northern (inlet) end of this spillway (see black arcs and question marks in Fig. 8) but whether failure of this possible dam was the mechanism of formation of the spillway remains ambiguous; but it can be said that the streamlined bedrock hummocks support an outburst flood hypothesis. For note, all of the moraines in the 'western spillway' area (Fig. 8) are likely to belong to the Umîvît moraine system, dating to 7360 to 7963 cal. a BP since they broadly match the extent of that moraine system as presented by Storms et al. (2012; Fig. 1).

North of Aajuitsup Tasia

The area immediately to the north of Aajuitsup Tasia, including that known locally as 'Maniitsoq', contains: (i) several laterally-extensive and horizontal benches on multiple valley sides and these are shorelines at 292 m a.s.l. and 312 m a.s.l. (Fig. 6A) (ii) cols between these valleys with a box-shaped cross-section and with a floor with mean elevation 300 m.asl, i.e. between these two shorelines and thus indicative of a spillway (Fig. 9), (iii) fan-shaped deposits with an apex situated at the spillway exits, usually with a steep down-slope edge, often with incised gullies set into this edge (Figs 5B, 9), and (iv) sub-parallel > 5 m local relief ridges with arcuate crests trending transverse to (palaeo)ice flow and in direct contact with shorelines but situated (only) in elevation below the shorelines (Figs 5C, 6A). These sub-parallel low-relief ridges are topographically and geomorphologically analogous to the arcuate moraines described on Baffin Island by Andrews & Smithson (1966) and are most likely De Geer moraines (c. f. Lindén & Möller 2005).

The Aajuitsup Tasia – Sanninasoq valley and palaeolake(s) apparently received meltwater from numerous sources, including directly from the Russell Glacier ice margin and from meltwater draining over at least two cols (Figs 6B, 9), i.e. spillways. A fan-shaped landform on the slopes immediately to the south and altitudinally beneath each spillway indicates sedimentation into a lacustrine environment from debris-charged meltwater routing over the spillways (Fig. 5C). Therefore the Aajuitsup Tasia –

Sanninasoq valley palaeolake(s) were contemporaneous with meltwater routing through the two spillways.

This entire valley has a reverse bed gradient (i.e. towards the east) so the deepest part of any palaeolake (at ~50 m) was towards the present ice margin. This depth of water and this reverse bed gradient means that palaeolake(s) in this valley were likely ice-dammed at the easternmost end and by an advanced Russell Glacier terminus. The reverse bed gradient also explains why this valley holds such well-preserved moraines on the valley floor, because erosion by (late Holocene) meltwater and submergence by (late Holocene) sedimentation has not occurred, in contrast to other neighbouring valleys where moraines are predominantly preserved on valley sides.

South of Leverett Glacier proglacial area

The southernmost extent of the Leverett (palaeo-)proglacial area (Figs 6B, 6C) connects to the Ørkendalen valley via an impressive 4.2 km long gorge (Fig. 5D). This gorge drops >200 m in elevation and has slopes of up to 105 and 55 m high on the northern and southern sides, respectively (Fig. 10 inset). Several inflexions in the northern side slopes may be associated with a palaeosurface on the southern side (Fig. 10 inset). Indeed in the lower part of the gorge, where it broadens to >400 m wide, several (discordant) terrace flights occur. These flights indicate sedimentation with a higher base level (altitudinally far above palaeo-sea levels), which was most likely due to the presence of the Ørkendalen valley glacier effectively blocking the southernmost part of the gorge. Indeed the wider southern part of the gorge is within the outer limit of Ørkendalen lateral moraines (Fig. 10). It is ambiguous as to whether the gorge head has been overridden or infilled by moraine and glacifluvial sediment. Therefore this gorge could either be contiguous with, or could pre-date, the most extensive ice margin of the Leverett Glacier between 6406 and 7028 cal. a BP (Storms *et al.* 2012). The 2.7 km broad and 600 m long fan of sediment at the mouth of this gorge (Figs 5B, 10) has a relief of just 33 m and such a low-relief fan slope indicates relatively fine-grained deposition from fluidal flows.

The Ørkendalen valley

The Ørkendalen valley is notable for numerous stacked sequences of sub-parallel ridges that each have crests with alignment that is sub-parallel to local topographic contours

(Fig. 11). The ridge crests are typically 5 to 10 m but occasionally >15 m above the local (detrended) surface. Topographically, these moraine ridges can be partially distinguished from each other and from the surrounding hillslopes by minor palaeochannels (Fig. 11). These palaeochannels tend to be relatively small (<20 m in cross-section) and occur in especially well-developed series on valley sides. Larger palaeochannels (tens of metres in cross-section) are isolated features and have greater sinuosity than the smaller channels. Some of the larger palaeochannels have an anastomosing planform and some have an undulating long-profile, all of which is indicative of subglacial channels (Greenwood *et al.* 2016), at least partially in bedrock. Specifically, we interpret that these channels represent a sub-marginal setting where ice surface drainage reached the bed through thin (a few tens of metres) ice (c. f. Margold *et al.* 2013). These local landform assemblages are therefore most likely moraine-kame terrace complexes (Weidick 1968; Ten Brink 1975) and they record progressive thinning and retreat of the Ørkendalen valley glacier over ~1000 years.

Discussion

Moraine types

An absence of steep rock walls surrounding the ice sheet margin, the generally massive and hard crystalline geology of the region, and the aridity of the climate means that during the mid-Holocene, as at present, frost weathering and thus accumulation of supraglacial debris was very restricted. Consequently the mapped moraines are probably composed of debris that has melted-out from basal ice, as has been described in this study area for contemporary and LIA moraines by Knight et al. (2002), Adam and Knight (2003) and Forman et al. (2007). The mapped position, spatial arrangement and geometry of individual ridges can be used to suggest three distinct moraine types. Characterisation of different types of moraines based on fine-resolution topography is absent from the Greenland literature and yet is important given a lack of opportunity for sedimentological analyses due the problems with accessing this terrain.

Sparsely-spaced and discontinuous moraine ridges with irregular and sinuous planform, with undulating and relatively sharp-crested ridges, with symmetrical cross-sectional shape and situated on plateaux and less commonly on valley sides are most likely to be end moraines. Specifically, this geometry and geomorphology suggests that they are probably push and squeeze moraines and thus they are analogous to the moraines along the present-day northern flank of Russell Glacier (Knight *et al.* 2002, Adam & Knight 2003). The present-day moraines are possibly of larger dimensions than those of the mid-Holocene because they are accretions from several advances (Forman *et al.* 2007). These types of moraines develop incrementally over multiple seasons and may relate to episodes of glacier thickening (Evans & Heimstra 2005) but we acknowledge that sedimentological information is required to unravel the exact sequence of formational processes.

Mapped multiple ridges that are closely-spaced, sub-parallel and often in concentric arcs, are restricted to positions on wide and low-angle valley floors, specifically the Leverett Glacier proglacial area and at the westernmost extent of several of the lakes north of Aajuipsup Tasia (Fig. 9). They tend to have asymmetric cross-sections. They are thus most likely to be composite ridges, or thrust-block moraines and suggest compressive (ice advance) interaction of the glacier terminus with frozen ground and results in subglacial sediment becoming elevated (Hambrey & Huddart 2006). This processes will be greatly facilitated where a glacier is flowing uphill out of an overdeepening, as at Leverett Glacier (see maps in Morlighem *et al.* 2013), or onto a locally-inverse bed slope such as at the westernmost extent of several of the lakes north of Aajuipsup Tasia (Fig. 9).

At the easternmost end of Aajuitsup Tasia the mapped moraines are situated below the altitude of palaeo-lake levels and if they are contemporaneous with the shorelines, which they seem to be due to an apparent direct physical contact, then they are most likely De Geer moraines (De Geer 1889; Lindén & Möller 2005). De Geer moraines are indicative of grounded ice margin retreat within a water body; in this case an ice marginal lake. The most eastward De Geer moraine corresponds in position to the most eastward extent of shoreline(s), hence it is apparent that a grounded Russell Glacier ice margin retreated eastwards progressively into deeper water and that this water was impounded at its easternmost end by an ice-dam.

Meltwater landform formation

The palaeoglaciological significance of the moraines and meltwater landforms mapped in this study concern ice margin position(s) and meltwater dynamics, respectively, although these two sets of conditions are spatio-temporally inter-related. Meltwater landforms have often been used with ice-marginal landforms to infer past ice sheet geometry and dynamics, and these interpretations are necessarily based on relatively simple assumptions regarding landform formation (Greenwood *et al.* 2016). In contrast, inference of palaeo-hydrological systems using meltwater landforms are much less common and usually target prominent landform-assemblages such as eskers and tunnel valleys; recent examples include Nitsche *et al.* (2013), Phillips & Lee (2013), Storrar *et al.* (2014), Burke *et al.* (2015), Lee *et al.* (2015) and Livingstone *et al.* (2015).

The overwhelming signature of meltwater activity during the mid-Holocene in this part of west Greenland is that related to proglacial meltwater, in the form of ice marginal lakes, perched deltas into these lakes, spillways feeding, connecting and draining these lakes, perched fans and glacifluvial terraces. Most of this landform evidence can be explained by hypothesis of normal ice ablation-fed river flows. However, some evidence such as box canyons and streamlined bedrock mounds, together with consideration of the necessary dam to impound the lake water, has suggested high-magnitude glacier outburst flood ('jökulhlaup') activity (c.f. Carrivick *et al,* 2004; Carrivick 2007). Evidence of a dynamic and varied proglacial runoff routing and style during the mid-Holocene, and a likely direct association of this with ice margin dynamics, comes from the multiple shorelines around several of the (probably ice-dammed) palaeolakes (Carrivick & Tweed 2013). Specifically, a shoreline has to be formed where lake levels are relatively stable, and if the lake was ice-dammed, that requires a quasi-stable ice dam thickness.

The hundreds of minor palaeochannels that exist in stacked successions throughout the study area and that are particularly pervasive in the Ørkendalen valley (Fig. 11) delineate progressive recession and thinning of the ice margin. In direct association with kame terraces, these channels illustrate drainage of supraglacial meltwater along the mid-Holocene ice margin, especially where ice has been pinned against a topographic slope (c. f. Kleman et al. 1992; Dyke, 1993; Greenwood et al.

2007; Margold et al. 2013). The fact that the larger (up to a few tens of metres wide) of these lateral channels are apparently at least partially cut into bedrock means that they probably represent former subglacial drainage routeways and thus should be considered to be submarginal channels, i.e. formed at the lateral margin but beneath the ice surface (c. f. Greenwood *et al.* 2007; Syverson & Mickelson 2009; Lovell *et al.* 2011; Margold *et al.* 2011, 2013b). For comparison, the contemporary ice margin on the northern flank of Russell Glacier is known to hold a submarginal channel that connects the large ice-dammed lake with 'overspill lake 1' (Russell *et al.* 2011). Palaeochannels with a distributary arrangement (e.g. Fig. 5B) are not concordant with moraines in position or orientation, so do not correspond to likely former ice margins and therefore could mark former minor subglacial channels.

Lateral channels have usually be attributed to cold-based ice margins, where percolation of meltwater in inhibited (Kleman *et al.* 1992; Dyke 1993; Kleman & Borgström 1996). The pervasive and dominant character and widespread distribution of lateral and submarginal meltwater channels in this study is similar to that found in parts of Scandinavia, the Canadian Arctic and the North American Cordillera, where cold-based or polythermal ice prevailed during deglaciation (Kleman *et al.* 1992; Dyke 1993; Sollid & Sørbel 1994). The relative paucity of landform evidence of *subglacial* meltwater in this study suggests that this part of the west Greenland ice sheet did not have a widely developed subglacial hydrological system during the mid-Holocene. An apparent lack of a developed subglacial hydrological system is not at all unusual in considerable parts of ice sheets (Kleman & Glasser, 2007) and together with the relatively ubiquitous lateral meltwater channels, can most simply be explained if the ice were cold-based or polythermal (c. f. Kleman *et al.* 1992; Dyke 1993; Sollid & Sørbel 1994).

Controls on spatial distribution of landforms

The spatial pattern of moraine ridges, palaeochannels and kame terraces across our study area is relatively coherent, with no quantifiable (statistically significant) change in spatial density, orientation or geometrical size of landforms, such as along the entire length of Ørkendalen (Fig. 11). Therefore, and in terms of the moraines, ice margin retreat has apparently been consistent in style for a time period of hundreds to thousands of years. In terms of meltwater, we find that the geological legacy of runoff generation, routing and magnitude-frequency style changed little for ~1000 years during the mid-Holocene. This uniformity of geomorphological evolution due to meltwater runoff contrasts with that during the Laurentide Ice Sheet deglaciation (Storrar *et al.* 2014) and can tentatively be attributed to a lack of abundant sediment and to a lack of subglacial bed-channel coupling if compared with the historical (decadal-scale) landform evolution at Breiðamerkurjökull, for example (Storrar *et al.* 2015).

The asymmetric distribution of moraines within major moraine complexes, such as between the contemporary termini of Russell Glacier and Isunnguata Sermia (Fig. 7) is probably due to glaciers in troughs having very different dynamics to that part of the ice margin situated at higher elevation between valleys, and on the timescales of tens to hundreds of years. Such a control of local topography on local ice sheet terminii is similar to that interpreted for the Canadian Arctic Archipelago (Storrar *et al.* 2014) and has been noted by Weidick *et al.* (2012) for the Nuup Kangerlua region in southern West Greenland. It is also evident from contemporary ice surface velocity measurements (e.g. Palmer *et al.* 2011; Morlighem *et al.* 2013) and ice margin retreat rates in this part of west Greenland.

The small portions of the landscape that we find in this study with a till/drift mantle/veneer (Fig. 6D) that is indicative of subglacial deposition probably represent transition from an actively retreating ice margin to a more complex and stagnant ice body (c. f. Margold *et al.* 2013). A present-day example of this transitional situation is shown in Figure 6E. Further support for the idea that this part of the west Greenland ice sheet was active throughout its mid-Holocene retreat perhaps comes from the absence of eskers. Eskers tend to be a key focus of glacial geomorphology-based studies on ice sheet margins and are common at other ice sheet margins (e.g. Brennand 1994; Clark & Walder 1994; Winsborrow *et al.* 2010). However, mindful that eskers have been associated with active ice margin retreat in Iceland (e.g. Evans & Twigg 2002) other factors additional to ice flow are likely to have been important in west Greenland in the mid-Holocene for restricting esker formation. We consider that such other factors include a distributed meltwater drainage system (c. f. Livingstone *et al.* 2015) and restricted sediment supply due to a lack of supraglacial debris and due to a cold-based subglacial thermal regime, as described and interpreted earlier in this study, respectively. Additionally, a lack of deformable sediment could explain the lack of eskers (Burke *et al.* 2015) but both factors are contrary to the contemporary situation in this part of west Greenland (e.g. Russell *et al.* 2011 and Adam & Knight 2003, respectively).

In this study we did not find any relationship between moraine or palaeochannel distribution (spatial density) and geological variability. This fact, together with the presence of several major and many minor channel systems incised into the hard crystalline bedrock, suggests that the mid-Holocene ice margin system in this part of west Greenland was more similar to that understood for Antarctic continental shelves than for the periphery of the northern hemisphere Quaternary ice sheets. Indeed Antarctic continental shelf systems also have a general lack of evidence of subglacial drainage and an absence of eskers (e.g. Wellner *et al.* 2006; Graham et al. 2009) as has been found in this study. However, the Antarctic system has been suggested to have meltwater delivery through small canals or a deforming till aquifer (Graham *et al.* 2009; Noormets *et al.* 2009) and that cannot be evaluated by this study but is intriguing in the context of west Greenland and deserves careful consideration.

The northern part of our study area, along the Isunnguata Sermia valley, is peculiar because any discernible glacial geomorphology is extremely sparse (Fig. 7). Notwithstanding that this valley is deeper and wider there is no difference in topography or geology along the Isunnguata Sermia valley in comparison to the other valleys of this study we consider it useful to speculate on the major types of events that could explain such a pattern. In brief, during the mid-Holocene the Isunnguata Sermia terminus could have: (i) retreated very rapidly without sufficient time in any one configuration for discernible moraine ridges to be deposited and for palaeochannels to develop, or (ii) any moraine ridges and palaeochannels have become buried with the abundant glacifluvial sediment that is now and probably has been throughout the mid-Holocene accumulating on the valley floor, as a product of ablation-fed river flows and jökulhlaups.

Regional ice margin dynamics

Our mapping supports a general mid-Holocene regional eastward migration of the western margin of the Greenland Ice Sheet (GrIS) from Kangerlussuaq to the Russell Glacier area (c. f. Weidick 1968; Ten Brink 1975; van Tatenhove *et al.* 1996). Our mapping also reveals considerably more detail in the position, size and geometry of former ice margins associated with this general retreat by including not only major, contiguous, advances as by Ten Brink (1975) but also minor discontinuous moraine ridges. Numerous minor readvances are evidenced by the number of moraine ridges, although the spatial patterns of them suggests that these ice margin advances were short-lived. A low sediment supply as well as a short period of time of formation more than likely explains why these moraine ridges are far smaller in local relief than those at the contemporary ice margin. We have not sought to produce a qualitative reconstruction of landscape evolution because we realised that there were too many ambiguities and permutations, specifically that: (i) our sub-sites do not unfortunately have (cross-cutting) stratigraphical evidence between them, and (ii) as discussed above considerable asymmetry in (intra-complex) moraine positions is apparent across local valleys and adjacent hillsides and local plateaux so simply using the dated moraine complexes to link our sub-sites would be speculative at best. Such 'metachronous' problems are not unusual in landform records of ice sheets and probably require strategic and cautious application of an 'inversion model' to unravel them (Klemen & Borgström, 1996).

Regional meltwater dynamics

Meltwater has left a more widespread and arguably a more pervasive landscape record than deposition of sediment directly from a glacier in this study area. This landscape record is dominated by landforms related to transitory proglacial meltwater systems, all existing within a time period of ~1000 years, yet each with long-lasting geological legacy. In particular, this evidence includes temporary storage of meltwater in large (ice-dammed?) lakes such as Aajuipsup Tasia, drainage via deeply and progressively incised gorges such as from the Leverett Glacier proglacial area, and progressive ice margin thinning and retreat as in Ørkendalen, for example. Several major lakes (systems) must have had glacier termini abutting them to form dams, and several of these lakes had multiple levels probably indicating glacier thickening/thinning. The number and the size(s) of the palaeolakes revealed in this study and their transitory nature is comparable to the present day situation in west Greenland (Carrivick & Quincey 2014). Some lakes drained via spillways, and whilst it is presently ambiguous as to whether these were lake maintenance spillways or outburst flood spillways (c. f. Perkins & Brennand 2015), they illustrate the exchange of water between local valleys and thus major hydrology reorganisation as a response to changing ice margin configurations. The fans and deltas associated with these spillways are not huge, which is to be expected because any of the numerous lakes upstream would have acted as a sediment trap. The most spectacular sedimentation is actually at the distal end of the gorge emanating from the Leverett Glacier proglacial area and comprises sedimentation in the form of stacked sloping terrace edges as well as a major distal fan (Fig. 5D). The longitudinal transition from incision in proximal reaches of this gorge to sedimentation in the distal reaches is striking and demonstrates a rapid attenuation of (palaeo)flow transport capacity and energy (Fig. 5D).

Whether the lateral and submarginal palaeochannels represent a product of steady-state down-cutting by normal ice ablation-fed river flows, or else a single high magnitude or multiple lower magnitude erosional events is ambiguous, not least due to a lack of modern analogues of these sorts of channels (Greenwood et al. 2016). Where palaeochannels persist with a distributary planform, especially on inter-valley and plateaux areas, the position and nature of minor subglacial channels is suggested. Although minor in individual size(s), these could have been pressurised and with ability to affect ice dynamics. In contrast, whilst the larger palaeochannels mapped in this study at least partially cut into bedrock and thus are associated with Nye channels (see section 2.2.3 of Greenwood *et al.* 2016) they are isolated and do not link topographic basins so are without an obvious ability to affect major (palaeo)ice flow dynamics.

Conclusions

This study has greatly improved the spatial resolution of data and knowledge of topography, geomorphology and associated geology and landcover for the Kangerlussuaq – Russell Glacier area, west Greenland. These data will be useful for future work on deglaciation, which is likely to continue to focus on geochronology, landscape stability and development/succession. Specifically, future work in this study area should look to target the previously undocumented geomorphology revealed in this study for sedimentological and geochronological analysis of the composition and structure of the moraines, the lake sediments and the fan-shaped deposits.

This study has distinguished push-squeeze moraines, composite block moraines and De Geer moraines based on a set of topographical and geomorphological criteria developed for west Greenland. These moraines record minor ice advances and perhaps also some glacier terminus thickening, transport of (frozen) subglacial sediment, and grounding line deposition with ice marginal lakes, respectively. Asymmetry and discontinuity between intra-moraine ridges across adjacent valley hillsides and valley floors is attributed to a control of local topography on former ice dynamics.

Meltwater generated from this part of the west Greenland ice sheet during the mid-Holocene and likely over just ~1000 years has a legacy of landforms that reveal major reorganisations of proglacial routing and of runoff frequency-magnitude regime. In particular, large palaeolake systems such as the Aajuitsup Tasia complex had multiple shorelines and major spillways associated with them. The lake-related spillways and shorelines and dry gorges such as that emanating from the Leverett Glacier proglacial area evidence a spatio-temporally dynamic proglacial hydrology. Several of the lakes are on local inverse slopes and would likely have been ice-dammed and would thus have exerted a control on the mid-Holocene ice margin configuration and behaviour. Hundreds of minor palaeochannels have an intimate association with minor moraines on hillsides and are attributed to former ice marginal or lateral channels. They record progressive ice margin retreat and thinning, especially in the Ørkendalen valley, and are indicative of a cold-based ice margin. The few larger channels at least partially cut into bedrock are interpreted to be submarginal Nye channels. Minor palaeochannels comprise networks with a distributary planform that is discordant with moraines and are most likely the position of former (likely inefficient) subglacial meltwater channels.

This suite of glacial landforms perhaps has more similarity with that of Antarctic continental shelves than with most northern hemisphere ice sheet margins and whilst several topographical, geological and glaciological controls must have been important, such as a reverse bed slope, hard crystalline rock and a cold-based ice margin, respectively, a lack of sediment supply seems very evident and important.

Overall, the lack of any statistically significant difference in spatial density, and of landform size and orientation across this study site, means that the most pervasive impression given by this suite of landforms is that of considerable spatio-temporal variability of meltwater routing and runoff regime persisting for ~1000 years during the mid-Holocene, despite a relatively consistent pattern and style of ice margin retreat. Landforming events during the mid-Holocene in this part of west Greenland were very similar to those of the present day. A better understanding of the timescales involved is needed to examine whether the meltwater system re-organisations correspond to changes in ice margin dynamics or vice versa.

Acknowledgements

JLC acknowledges fieldwork funding from the School of Geography, University of Leeds, for field work in 2008, 2010, 2012 and 2015, and the Royal Institute of Chartered Surveyors (RICS) (administered via the RGS-IBG) for fieldwork in 2014 (grant no. 474: DJQ). Richard Hodgkins is thanked for his provision of the IPY07-03 airborne LiDAR data. Steve Carver assisted in fieldwork in 2012. Daniel Carrivick is thanked for his field photographs of the landscape east of Lake Fergusen and Andrew Tedstone and Neil Ross are thanked for photographs of the Leverett Glacier proglacial area. M. Winsborrow, two anonymous reviewers and Editor J. Piotrowski are thanked for their careful scrutiny and constructive criticism.

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Figure captions

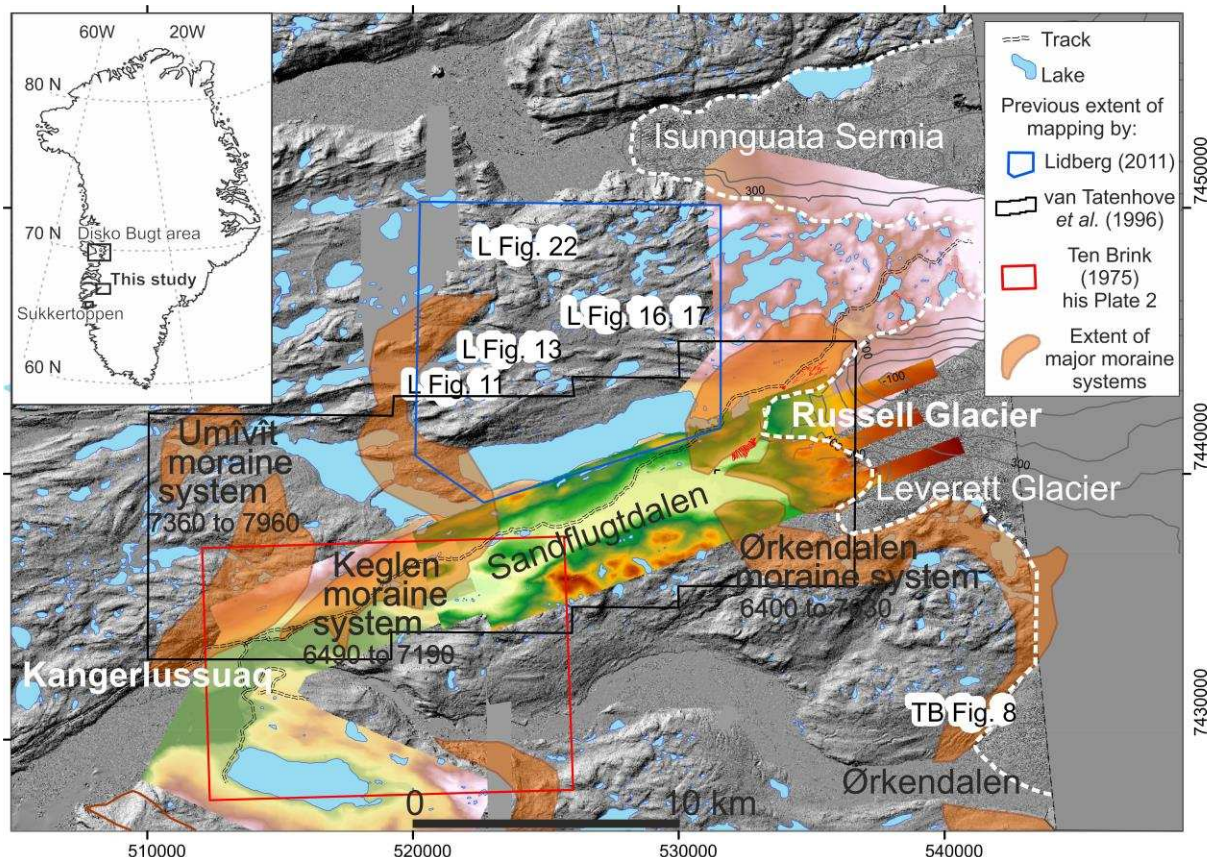


Figure 1. Study area, topographic data and extent of previous glacial geomorphology mapping. Topographic data includes SETSM 2 m grid DEM (b/w background), 5 m photogrammetry DEM (pastel colours), ALS 2 m DEM (vivid colours) and dGPS 3D points (red dots). ASTER DEM is not shown for clarity. Note glacier bed elevation (contours) is an extract from IceBridge data. The present day ice margin is represented by white dashed line. Field photographs by Lidberg (2011) and Ten Brink (1975) that were found to be useful to this study are geolocated on the map and with their figure numbers and with authorship denoted by ‘L’ and ‘TB’, respectively. Grid coordinates are in UTM zone 22N format and dates pertaining to major moraine complexes are in cal. a BP.

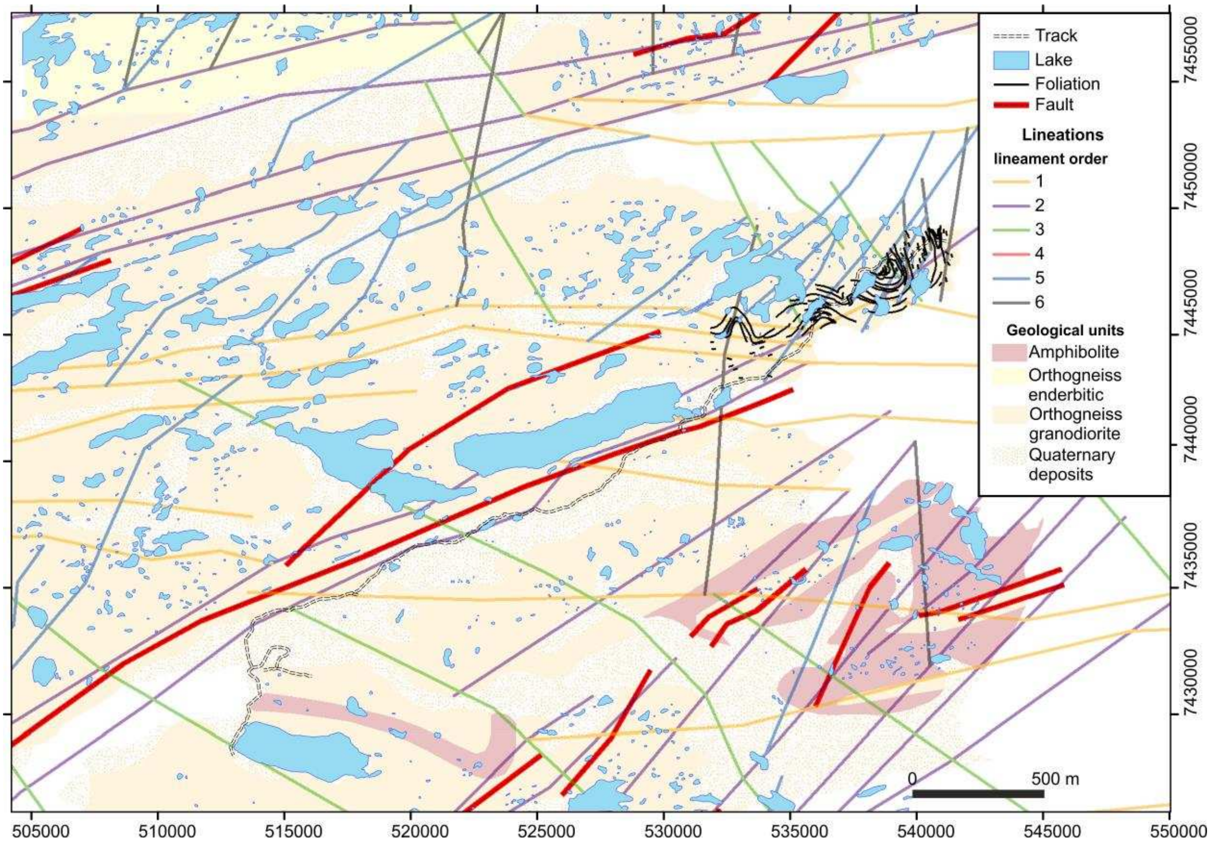


Figure 2. Geology of the Kangerlussuaq – Russell Glacier area adapted from the 1:500 000 mapping of Pedersen *et al.* (2013) and GEUS (2013). Pedersen *et al.* (2013) report that the lineaments are hierarchical and relate to structural features in crystalline rocks such as faults and shear zones, rock fabrics and discontinuities due to differences in rheology or competence.

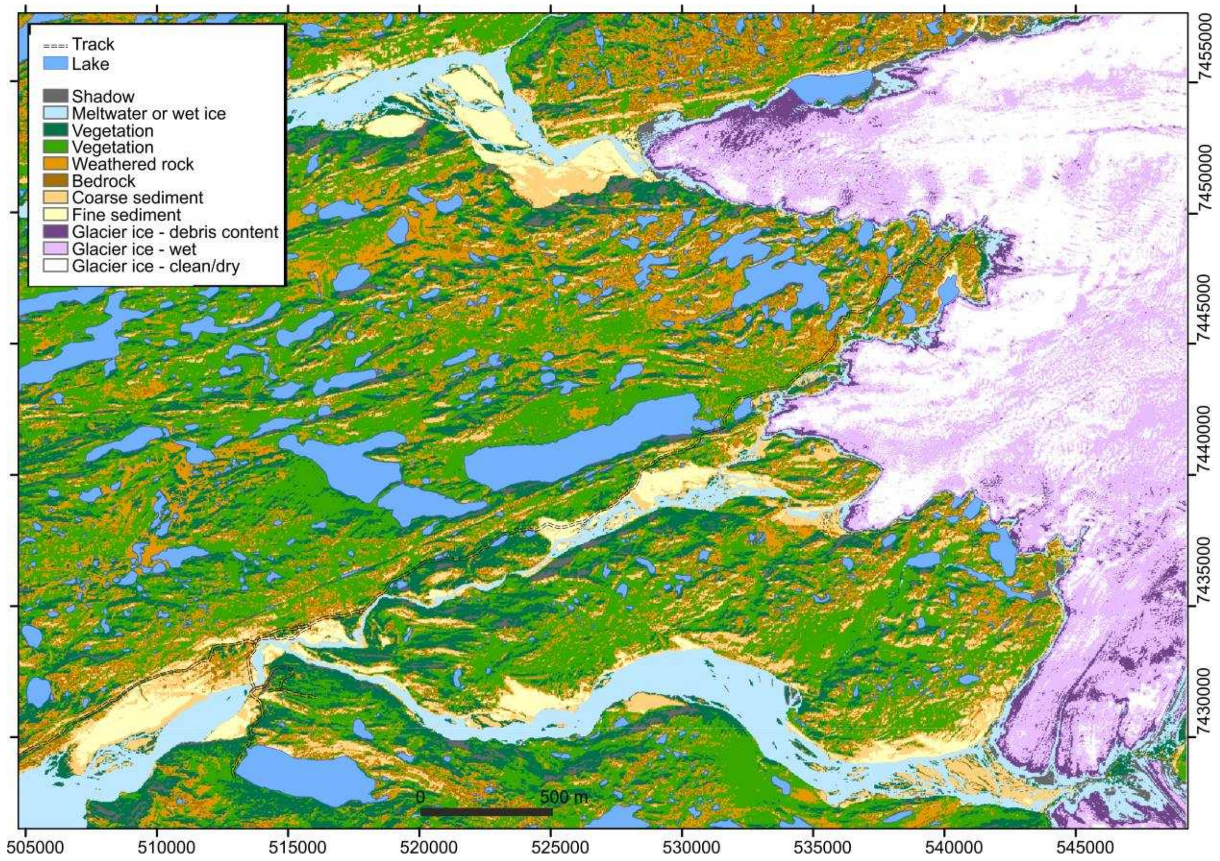


Figure 3. Landcover classification of the Kangerlussuaq – Russell Glacier area, achieved using 30 m cell size bands 2 - 7 of a Landsat 8 image and an ISODATA clustering algorithm.

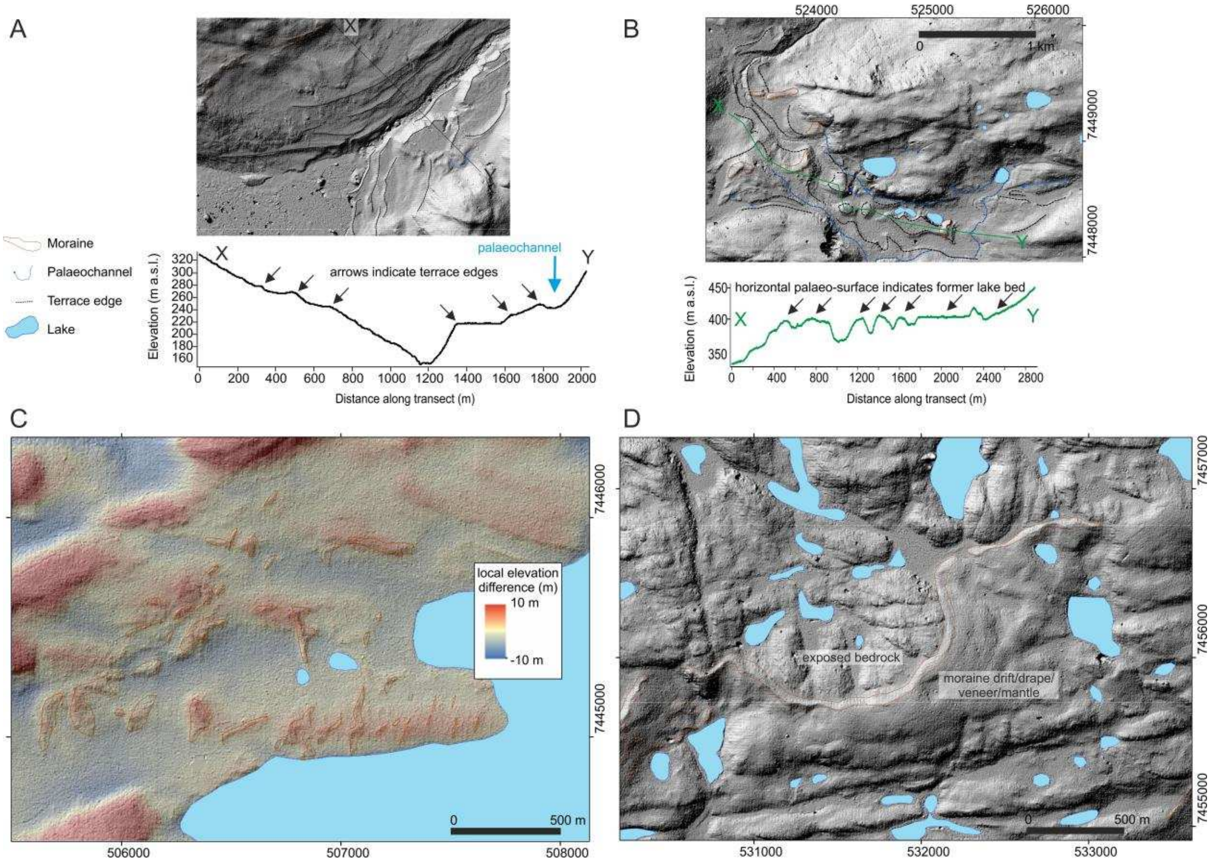


Figure 4. Examples of digital criteria used to identify glacial landforms, including elevation transects, in this case of asymmetric terrace edges (A), 3D visualisation of palaeochannels and transect of elevation (green line) (B), elevation deviations from a local trend to identify subdued moraine ridges (C), and visual assessment of surface texture to identify moraine drift/veneer/drape deposits (D). The location of these panels is indicated in Fig. 7.

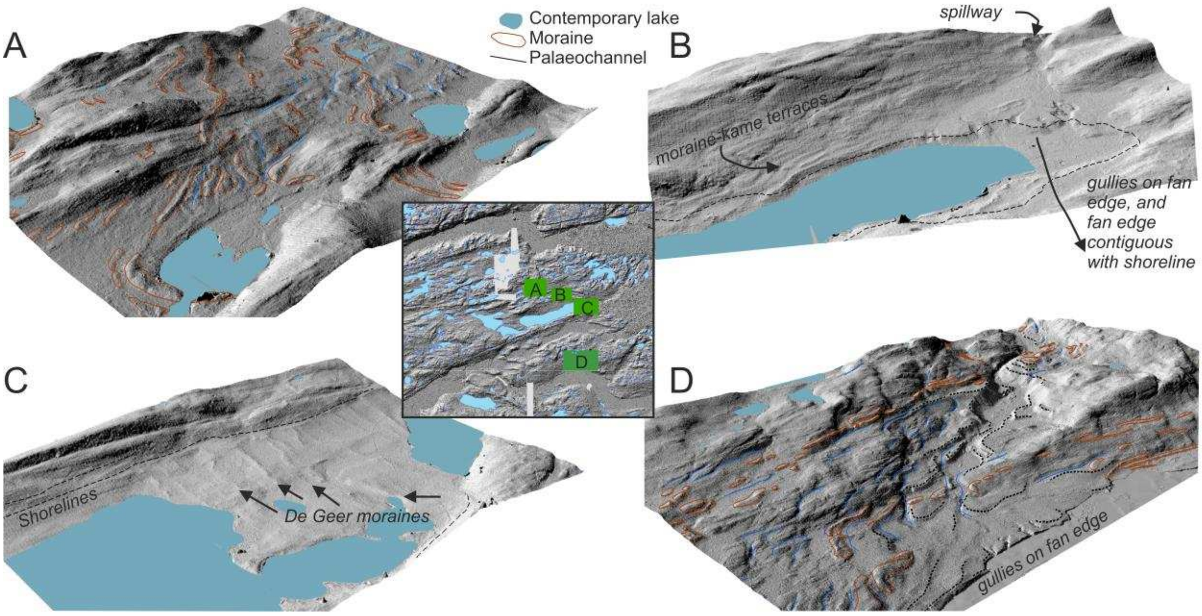


Figure 5. 3D visualisation of four sub-areas to illustrate landform type, geometry, position and context, specifically: the occurrence and character of major and minor moraine ridges (A), stacked terraces within a bedrock gorge and palaeochannels on hillsides and plateau surfaces (B), shorelines and De Geer moraines (C), and shorelines and perched fan-shaped deposits with incised edges (D). Note moraines and palaeochannels are visible in panels C and D but not encircled to maintain clarity of their topographic signature. Note that scale varies due to perspective of view.

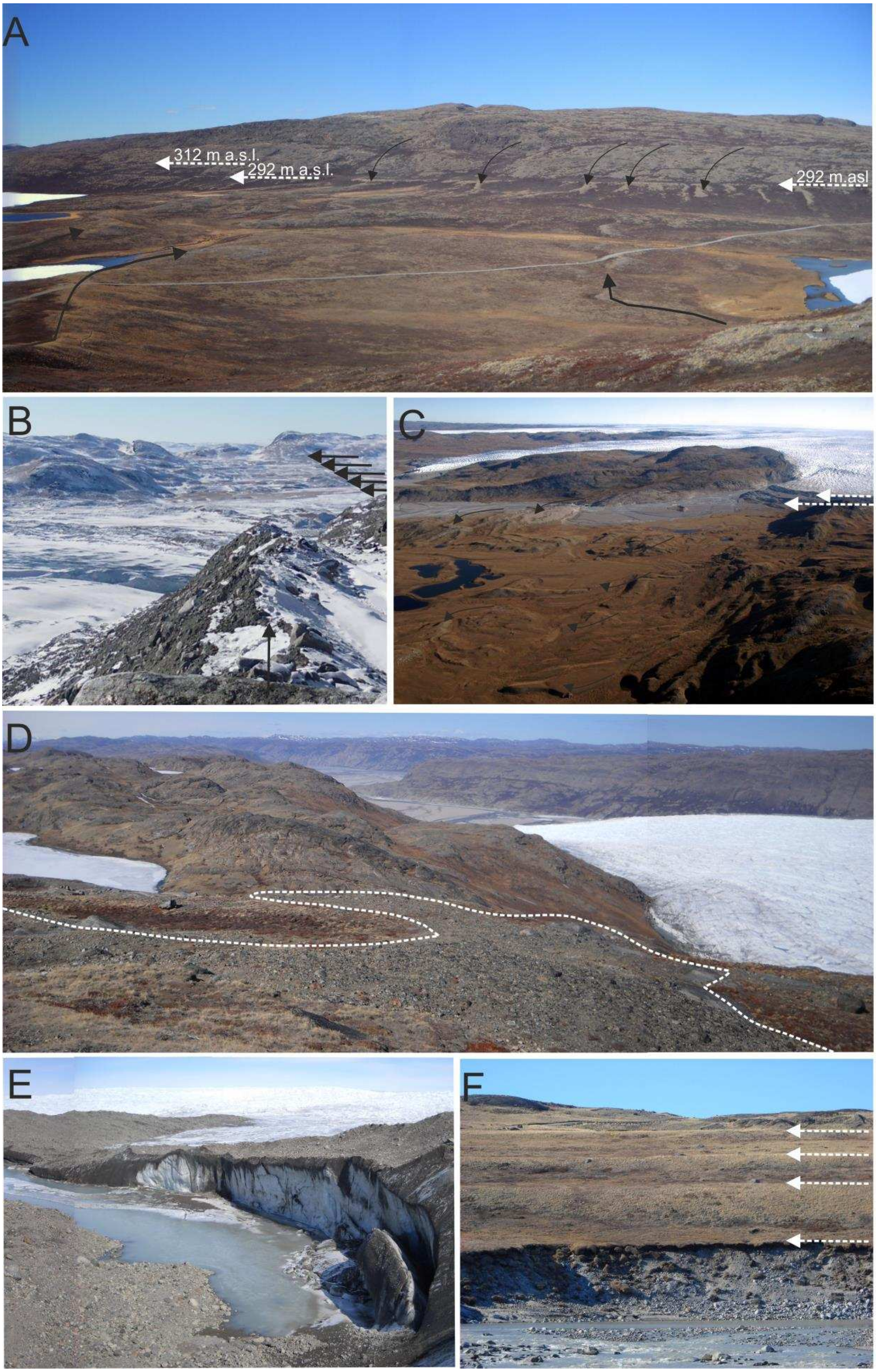

A
312 m a.s.l.
292 m a.s.l.
292 m.asl
B
C
D
E
F

Figure 6. Field photographs depicting examples of: shorelines (white arrows) and De Geer moraine ridges (black arrows) at easternmost end of Aajuitsup Tasia (A), and contemporary lateral moraine (vertical line) and moraine-kame terrace complex (horizontal arrows) in Leverett Glacier proglacial area (B), push moraine complex (white arrow) and mid-Holocene moraines (black arrows) at Leverett Glacier (C), patch of till/drift mantle/veneer on otherwise scoured bedrock surface immediately south of Isunnguata Sermia (D), stagnating ice with veneer of melt-out sediment at easternmost end of track (E), shorelines immediately adjacent to contemporary river draining northern flank of Russell Glacier (F). Image forming panel B is courtesy of Neil Ross C and image forming panel is courtesy of Andrew Tedstone.

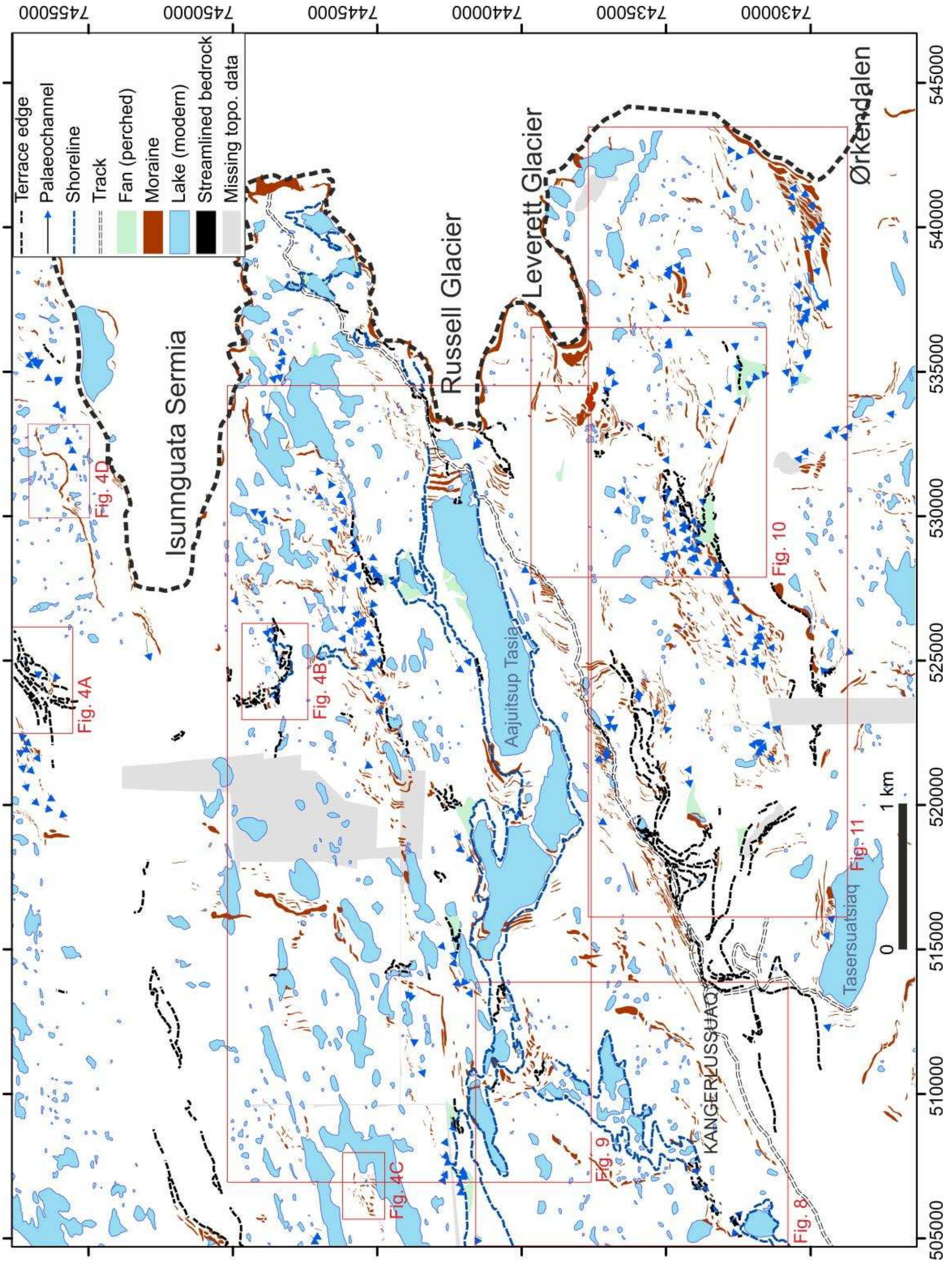


Figure 7. Overview of the glacial geomorphology in the Kangerlussuaq – Russell Glacier area. The present day ice margin is represented by black dashed line. A .pdf version of this map (without the other figure location outlines) enabling zoom, panning and with layers that can be switched on and off is available as Supporting Information (Fig. S7).

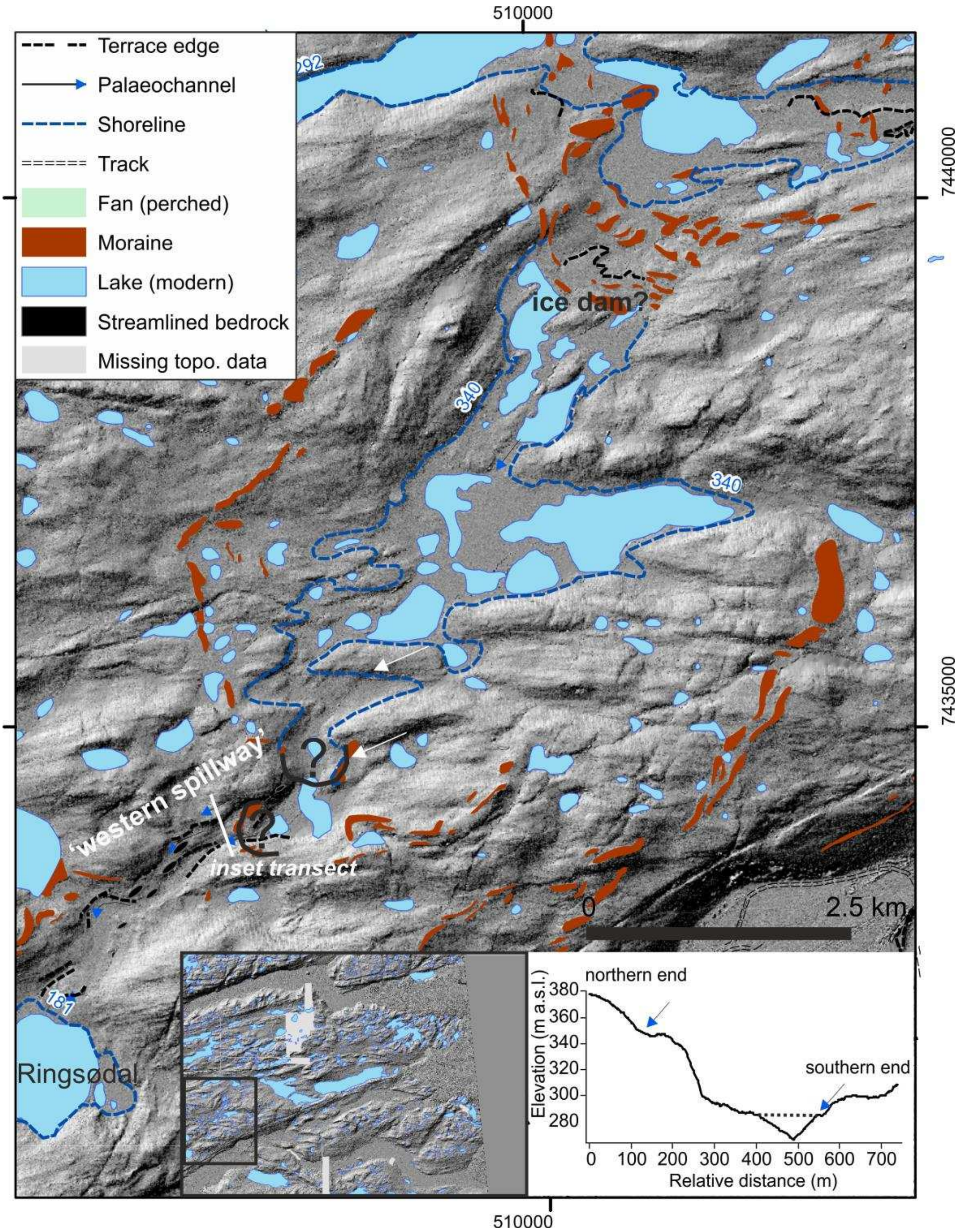


Figure 8. Glacial geomorphology of the western spillway and associated palaeolake.

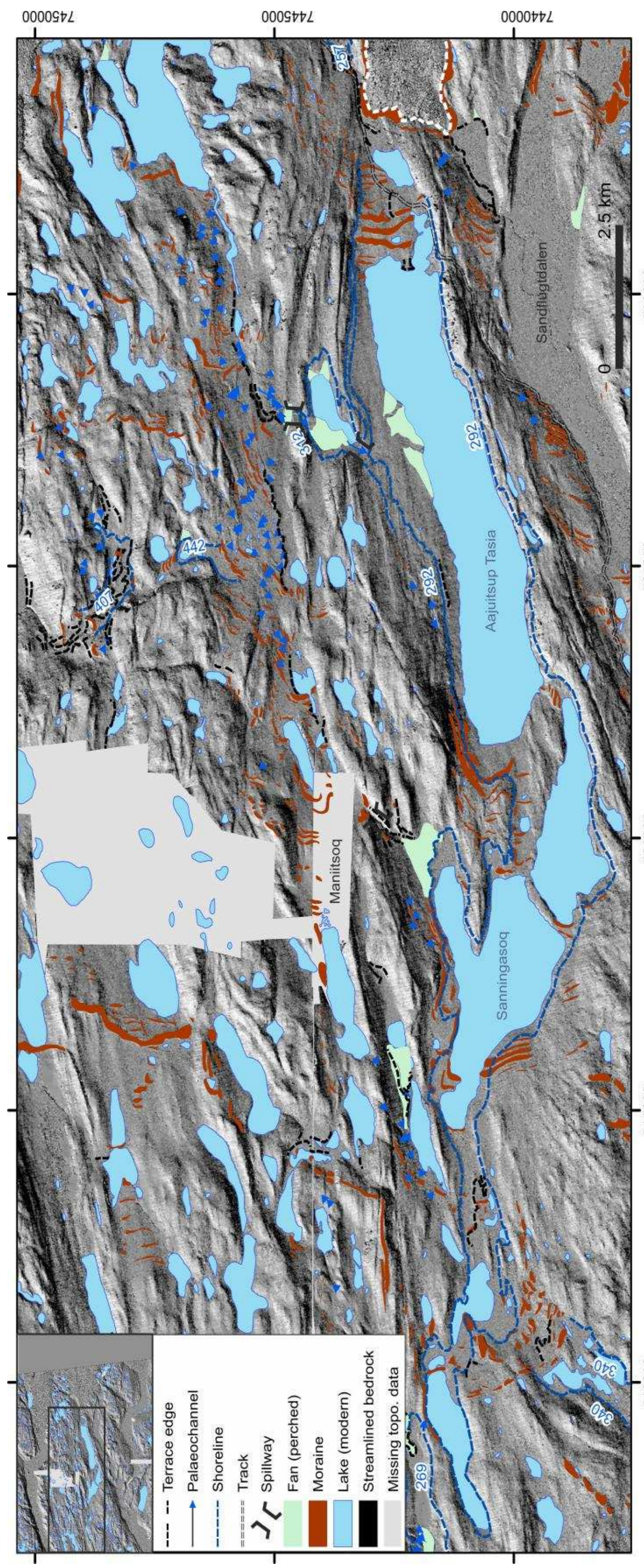


Figure 9. Glacial geomorphology of the Aajuitsup Tasia complex, highlighting evidence of shorelines and meltwater routing over several major spillways. The present day ice margin is represented by white dashed line.

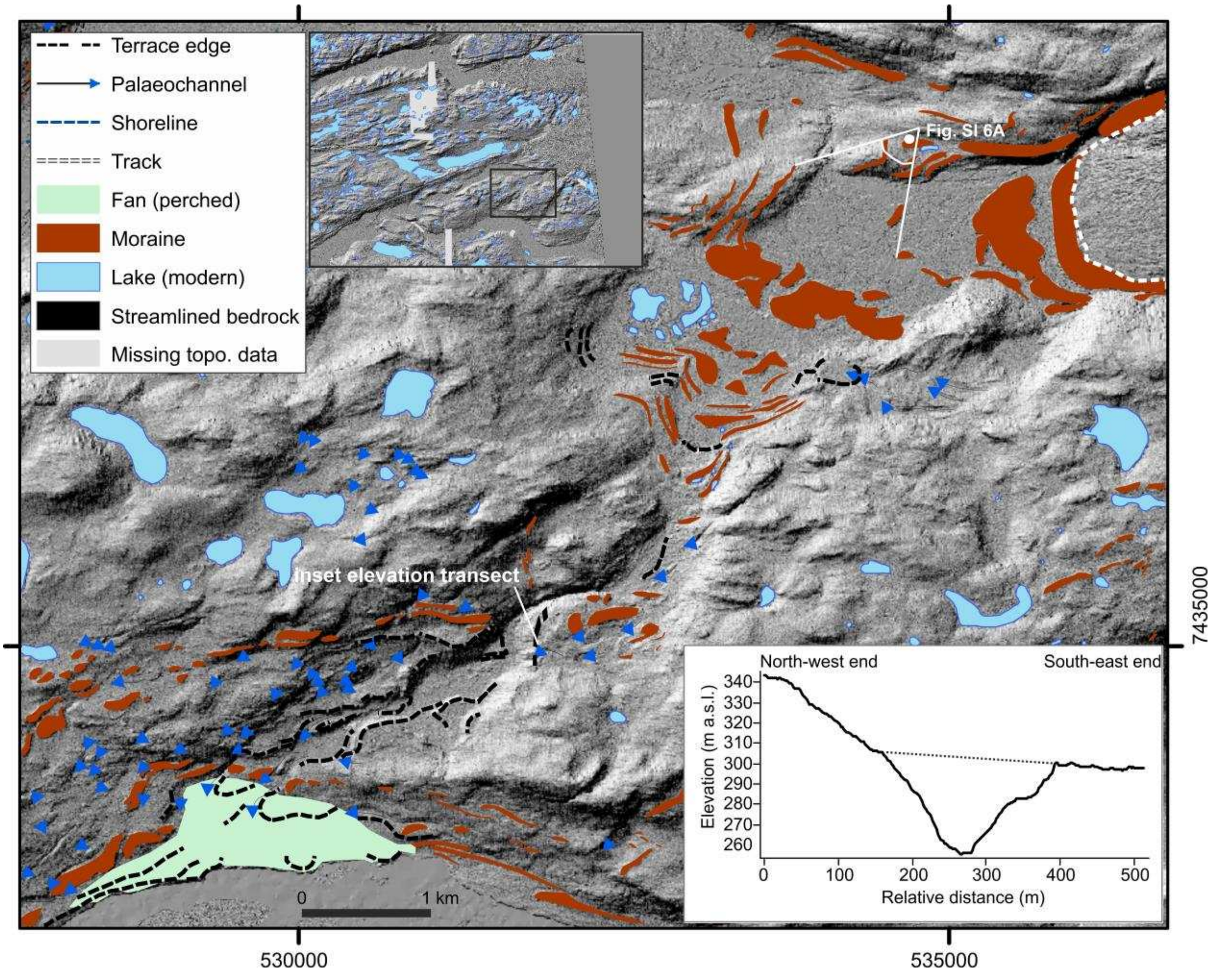


Figure 10. Glacial geomorphology of the palaeochannel emanating from the Leverett Glacier proglacial area. The present day ice margin is represented by white dashed line.

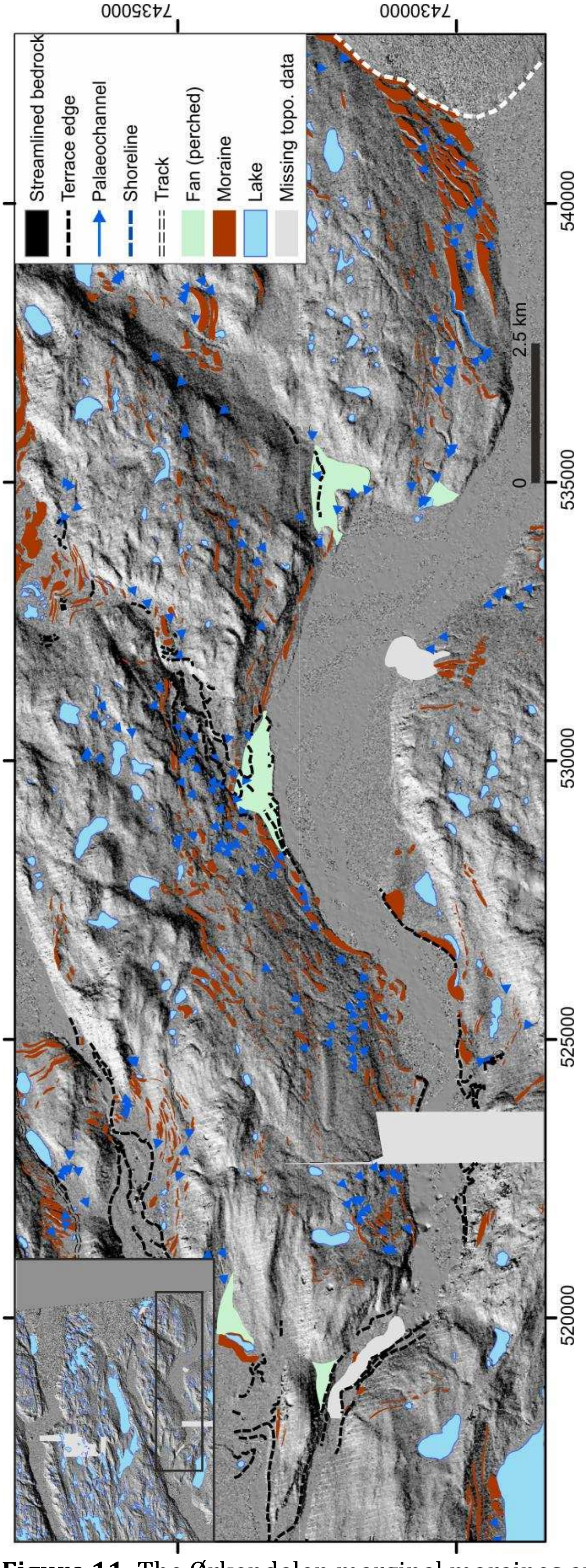


Figure 11. The Ørkendalen marginal moraines and meltwater channels. The present day ice margin is represented by white dashed line.

Description	Landform	Interpretation
Discontinuous ridges, often sinuous and hummocky along crest, and extends across landscape, especially in subparallel sets linking those on shallow gradient slopes with those across valley floors	Moraine ridge	Major complexes indicative of ice margin advance and minor ridges indicative of still-stand during recession
Sub-parallel ridges with sub-parallel and curved crests that trend transverse to (palaeo) ice flow and situated altitudinally below palaeolake shoreline	De Geer moraine	Indicative of seasonal advances during grounding line retreat
Discontinuous ridges on valley sides with relative smooth elevation profile along crest and especially with numerous parallel sets	Moraine-kame terrace complex	Indicative of ice marginal meltwater reworking moraine during ice surface lowering
Nested, sinuous and incised channel sets subparallel to topographical contours	Meltwater palaeochannel	Channel formed during ice surface lowering. Possibly subglacial if partly in bedrock
Pitted and hummocky surface, often with outsized boulders	Till/drift mantle/veneer	Subglacial deposition during glacier stagnation and passive ice margin retreat
Near-horizontal surface with smooth texture, often with identifiable bench especially in embayments. Exactly parallel to topographic contours and partially encircling a topographic basin	Shoreline	Former ice- or moraine-dammed lake
Sinuous channel with steep sides and streamlined bedrock hummocks within it, often over a col	Spillway	Major palaeochannel, probably carved during sudden lake outburst flood
Fan-shaped landform comprising gently-sloping top and steep foreslope, situated above modern lake level	Perched delta	Indicative of former sediment-charged fluvial system and of former lake level/base level
Horizontal terrace edges	Incised braidplain or marine terrace	Change in base level and/or meltwater-sediment regime
Sloping terrace edges	Incised fluvial fan or delta	Change in base level and/or meltwater-sediment regime

Table 1. Geomorphological criteria to identify ice marginal and subglacial landforms in west Greenland, developed in part from Tables 1 and 2 of Perkins & Brennand (2015) and in part from the digital topographical analyses of this study.

Supporting Information

Word document containing a description of method used to obtain, process and correct fine-resolution topography, and geology and landcover datasets

All Supporting Information figures are available as separate image files, except figure S 7 which is a .pdf format.

Figure S 1 Dates of imagery used to contruct the 2m grid SETSM DEM of the Kangerlussuaq – Russell Glacier area. Colours denote year groups.

Figure S 2A SETSM DEM tiles *before* relative vertical correction

Figure S 2B SETSM DEM tiles *after* relative vertical correction

Figure S 3 Comparison of grid cell elevations of SETSM (2m grid) with airborne laser scanner (ALS) data (2m grid), *before* (A) and *after* (B) horizontal co-registration.

Figure S 4 Comparison of grid cell elevations of SETSM (2m grid) DEM tiles with dGPS 3D points, a 2m ALS DEM, a 5m DEM from traditional photogrammetry, and an ASTER DEM (A). The ASTER DEM and the 5m DEM are apparently uncorrected for the geoid. The western-most SETSM tile was 3.5 m lower than the dGPS and ALS data (B). In contrast the eastern-most SETSM tile had elevations that were in good agreement with the dGPS data, as indicated by the ellipse in panel C.

Figure S 5 Comparison of ALS 2m DEM (A) with the adjusted SETSM 2m DEM (B). Note SETSM DEM is rougher, arguably more noisy, due to interpolation between matched points identified in the SETSM photogrammetry algorithms.

Figure S 6 Field photographs of Leverett forefield (A) and of ice margin between Isunnguata Sermia and Russell Glacier (B).

Figure S 7 Mapped geomorphology overlain on hillshaded DEM, in a .pdf file with zoom and pan functions and with layers (e.g. moraines, palaeochannels) that can be switched on and off.

<please see accompanying .pdf file>

Supporting Information

Datasets and methods

Fine-resolution topography

Topography at high-resolution (2m grid) was downloaded in four tiles to cover the study area from the University of Minnesota Polar Geospatial Center (PGC) website: http://www.pgc.umn.edu/elevation/stereo. These digital elevation models (DEMs) were produced by the PGC using photogrammetric processing of stereo-pairs of DigitalGlobe imagery, specifically via the Surface Extraction with Triangulated Network-based Search-space Minimization (SETSM) algorithms (Noh and Howat, 2015). Note that this is a composite DEM, the seamless coverage being constructed from multiple image pairs from multiple flight lines from multiple dates (Fig. S 1).

Readers may wish to note that similar quality of glacial geomorphology mapping can now be achieved elsewhere in west Greenland (and in some other parts of the Alaskan and Canadian arctic) because SETSM DEM production by the PGC is ongoing. As high-resolution topographic data because available for more remote regions, via automated processing such as SETSM, the opportunity to exploit the excellent preservation of landforms within semi-arid arctic, sub-arctic and sub-polar environments will develop further.

We made a three-step evaluation of the 3D quality of the SETSM DEMs. Firstly, all datasets were projected to UTM zone 22N. We then compared elevations of overlaps of the four adjacent SETSM DEM tiles for relative consistency. One tile was found to have grid cell elevation values that were on average 3.44m higher than the other tiles and this was adjusted to best-fit (Fig. S 2). Following the relative vertical shift, the four SETSM DEM tiles were mosaicked and then this mosaic was horizontally adjusted, or 'co-registered' to our ALS data, the horizontal shift applied was -3.61m in the x direction, i.e. westwards, and +4.10m in the y direction, i.e. northwards (Fig. S 3). The vertical and horizontal adjustments made individually and in combination in this study are within the 4.44 Circular Error (CE) reported by Noh and Howat (2015) for the flight strip WV02_20100819 (their Table 2).

Finally, for absolute elevation checks we compared the SETSM mosaic DEM (2m grid) elevations to those of (i) an ASTER DEM (30m grid), (ii) a DEM (5m grid) produced using standard photogrammetry on 1:10,000 scale aerial photographs and as described in Carrivick et al. (2013), (iii) an airborne laser scan (ALS) dataset (gridded at 2m) as obtained from the UK Natural Environment Research Council (NERC) Airborne Research and Survey Facility (ARSF) campaign IPY07-03, and (iv) to ~ 10,000 differential Global Positioning System (dGPS) 3D coordinates as obtained over multiple field seasons and as utilised in Russell et al. (2011) and Carrivick et al. (2013), for example. The spatial coverage of each of these four topographic datasets is given in Figure 1 and notably encompasses wide swaths of land, which is in stark contrast to the NASA Operation IceBridge data, which is predominantly over ice and entirely composed of narrow strips, but nonetheless used by Noh and Howat (2015) for vertical elevation checks of the SETSM DEM. Our (absolute) elevation analysis identified excellent agreement in dGPS-derived elevations and in ALS-derived elevations and thus realised the necessity for a +3 m vertical shift of the SETSM DEM mosaic (Fig. S 4).

The resultant SETSM mosaic DEM as modified and utilised in this study is of an unprecedented fine-resolution given its spatial extent, but surfaces tend to be rougher, perhaps more noisy, than similar resolution ALS data (Fig. S 5). This 'roughness' is most likely due to interpolation between matched points that were identified in the SETSM photogrammetry algorithms (Noh and Howat, 2015).

Additionally, 1080 lake polygons were digitised to (i) provide a map layer for ease of navigation/orientation, and (ii) to mask them from the DEM, because although the SETSM algorithm was designed to reconstruct a water surface (Noh and Howat, 2015) we found that across this study site there were big errors in the DEM over and around lakes, probably due to reflectance issues in the optical imagery, shadow, partial snow cover, frozen or partially frozen water and boulders protruding through shallow water surfaces. We could not use the binary raster grids of calculated/interpolated elevation as provided automatically by the SETSM algorithm because at fine-resolution the interpolated grid cells included a lot of land surface other than lakes.

Finally, the SETSM DEM and the associated hillshade image of this DEM were visually inspected for remaining errors, most notably those caused by snow patches. Whilst we erred on the side of retaining as much data as possible particularly bad (massive spikes or sinks) errors were manually removed.

Geology and landcover

Geological unit lithology (Figure 2) was digitised as polygons from that mapped at 1:500,000 (Pedersen et al., 2013; GEUS, 2013). Major faults, regional lineations and local foliation (Figure 2) were digitised as polylines from that mapped by Klint et al. (2013). Due to the relatively coarse scale of the original mapping, both the geological polygons and the geological polylines were manually adjusted (by eye) to best-fit the position of the same features evident in the high-resolution topography.

Landcover (Figure 3) was classified using an ISODATA clustering algorithm applied to a Landsat 8 (operational Land Imager) scene acquired on 12th July 2014. Bands 2 – 7 (all 30 m spatial resolution) were used and the required number of classes was set at fifteen. This number of classes provided the optimum balance between class separation and class redundancy. The classification was subsequently validated in the field, primarily along a 30 km transect in the vicinity of the vehicle track (Figure 3). Dry southern slopes and plateau are dominated by steppe vegetation, specifically grasses and sedges (e.g. Carex supina, Kobresia myosuriodes, Poa glauca, Calamagrostis purpurascens, C. poluninii and Salix glauca). Wetter northern slopes and hollows were dominated by slightly taller vegetation (e.g. Betula nana, Carex norvegia, Juncus arcticus and Rhododendron tomentosum or 'Ledum palustre'). Aeolian sediments often appeared to be overgrown with Calamangrostis purpuracens, Artemisia borealis and Rumex actosella (Génsbøl, 2004; Willemse 2003). Some manual editing of the landcover was necessary because some snow patches, which we identified with comparisons to optical imagery and to the high resolution topography, were misclassified as water. The final landcover classification (raster image) was adjusted in horizontal position by -26.00 m in the X direction (i.e. west) and by +18.3 m in the Y direction (i.e. north) to best-fit our high-resolution topography.

All Supporting Information figures are available as separate image files, except figure S 7 which is a .pdf format.

Figure S 1

Dates of imagery used to construct the 2m grid SETSM DEM of the Kangerlussuaq – Russell Glacier area. Colours denote year groups.

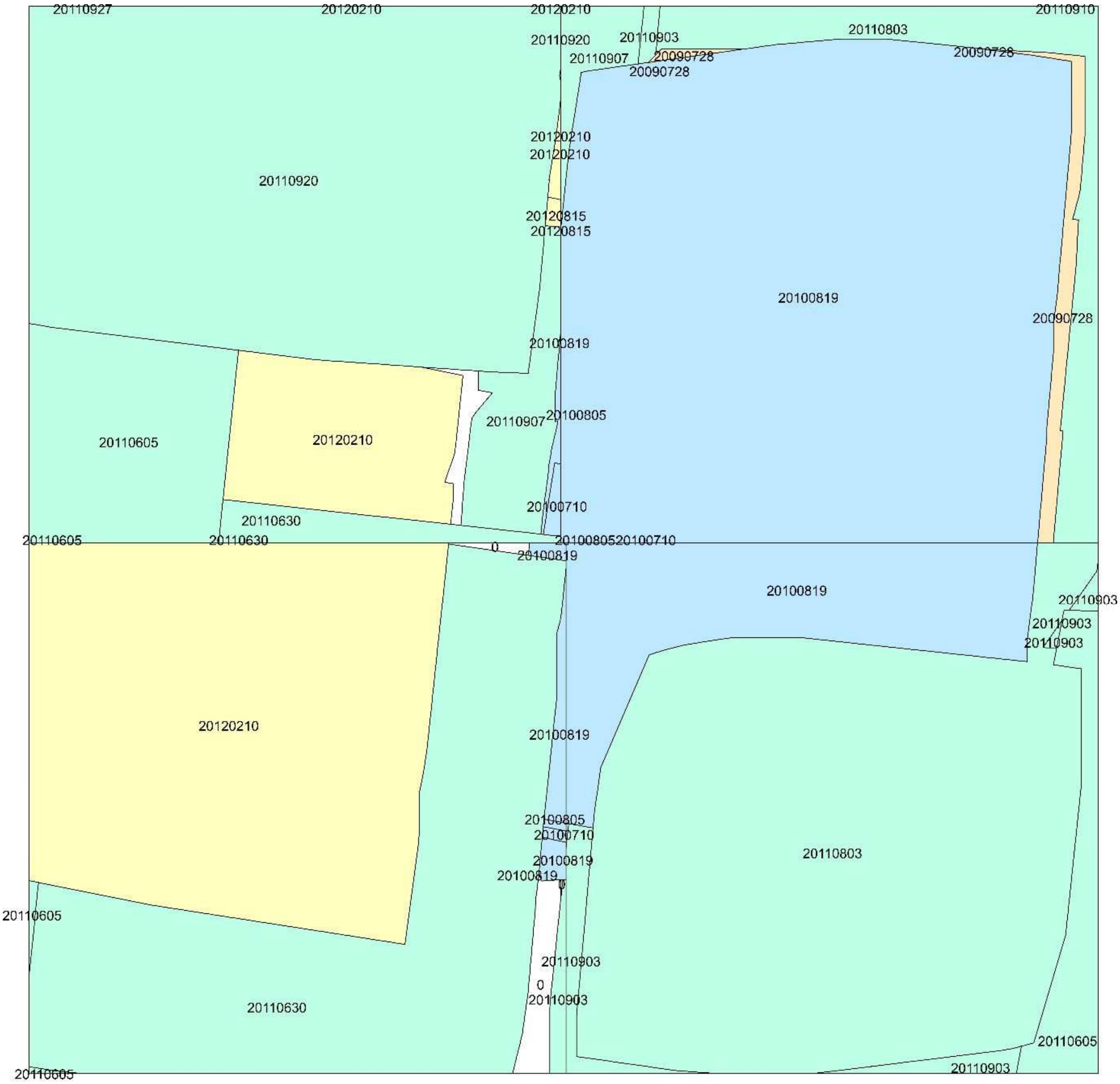


Figure S 1
Dates of imagery used to contruct the 2m grid SETSM DEM of the Kangerlussuaq – Russell Glacier area. Colours denote year groups.

Figure S 2A

SETSM DEM tiles *before* relative vertical correction

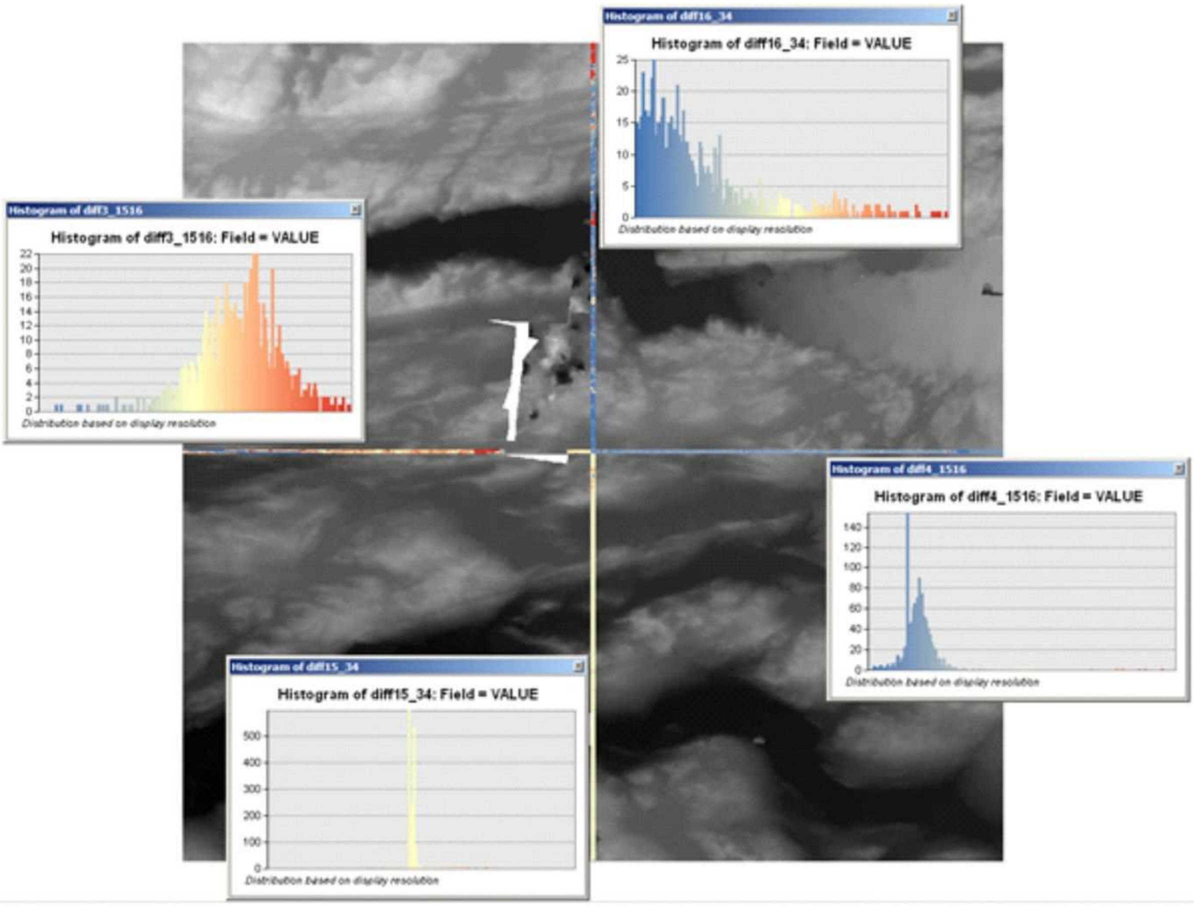


Figure S 2A

SETSM DEM tiles *before* relative vertical correction

Figure S 2B

SETSM DEM tiles *after* relative vertical correction

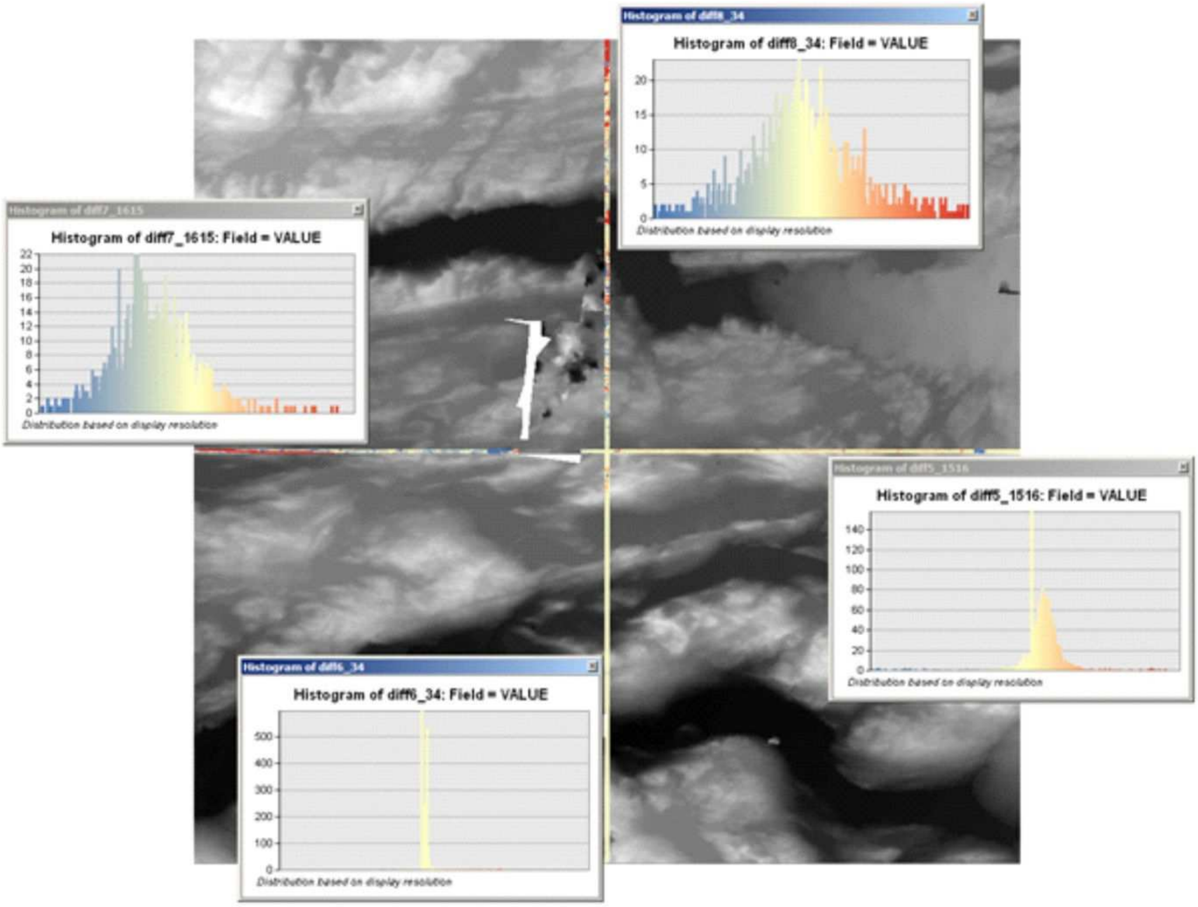


Figure S 2B

SETSM DEM tiles *after* relative vertical correction

Figure S 3

Comparison of grid cell elevations of SETSM (2m grid) with airborne laser scanner (ALS) data (2m grid), *before* (A) and *after* (B) horizontal co-registration.

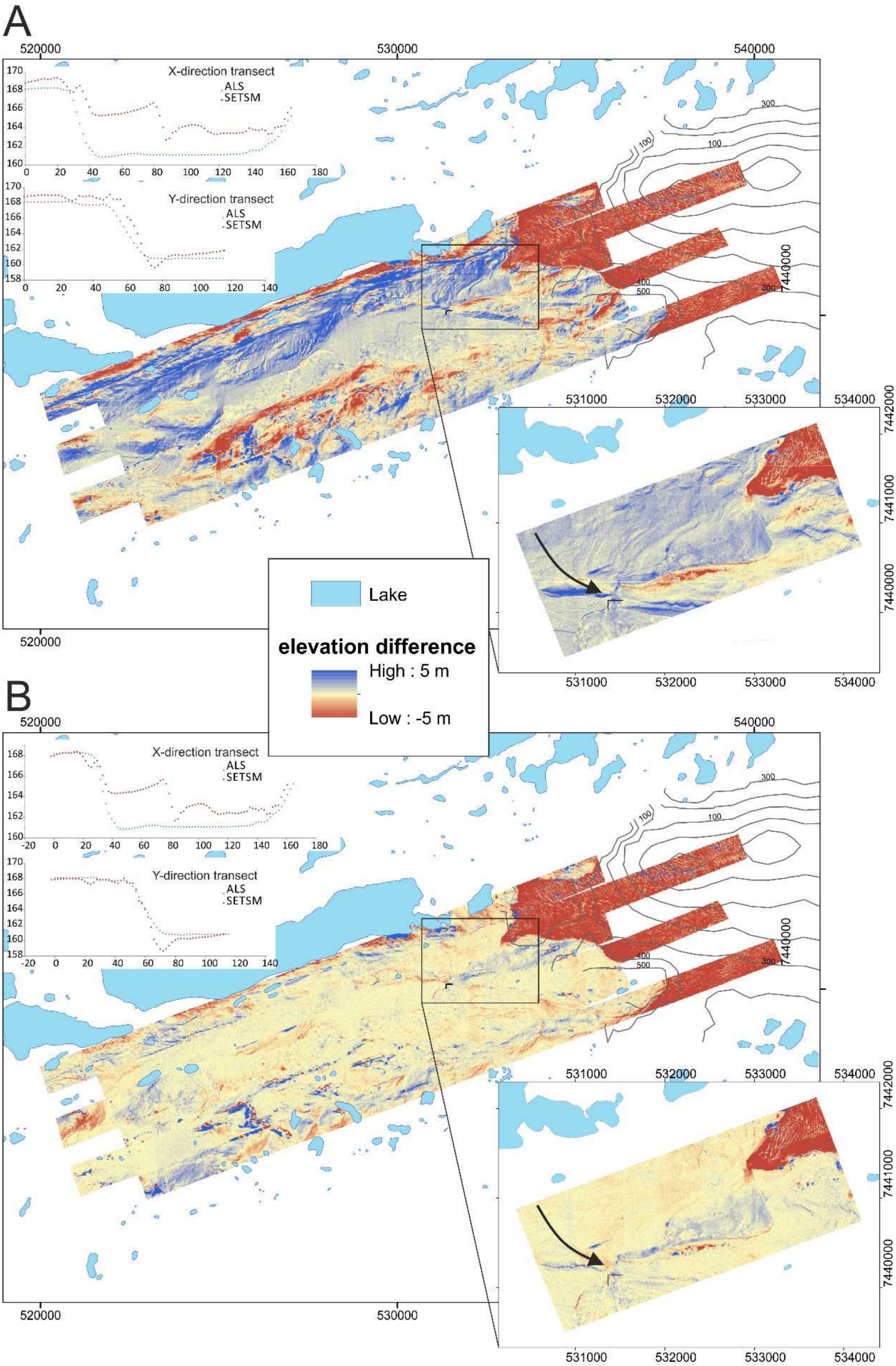


Figure S 3

Comparison of grid cell elevations of SETSM (2m grid) with airborne laser scanner (ALS) data (2m grid), *before* (A) and *after* (B) horizontal co-registration.

Figure S 4

Comparison of grid cell elevations of SETSM (2m grid) DEM tiles with dGPS 3D points, a 2m ALS DEM, a 5m DEM from traditional photogrammetry, and an ASTER DEM (A). The ASTER DEM and the 5m DEM are apparently uncorrected for the geoid. The western-most SETSM tile was 3.5 m lower than the dGPS and ALS data (B). In contrast the eastern-most SETSM tile had elevations that were in good agreement with the dGPS data, as indicated by the ellipse in panel C.

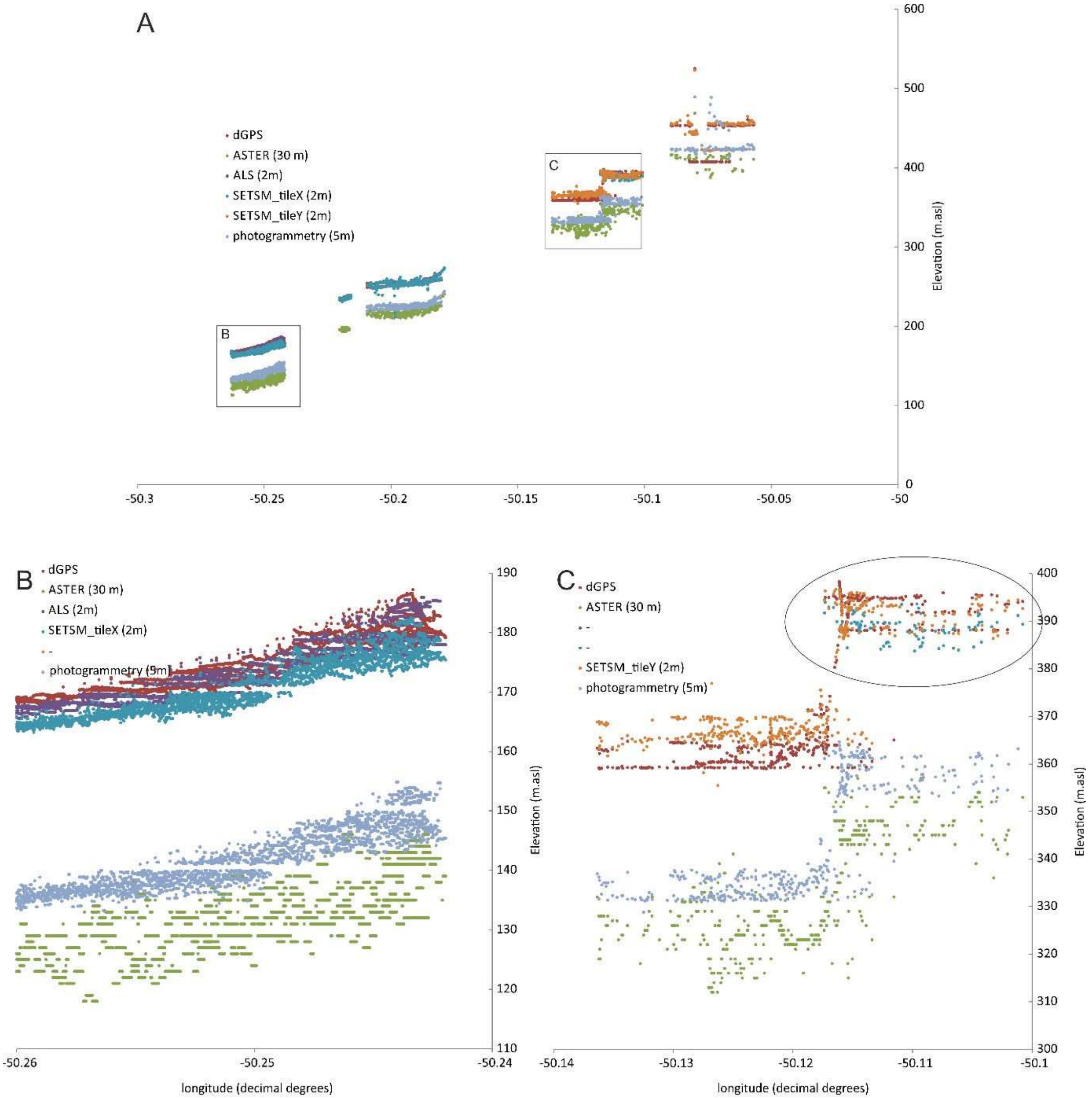


Figure S 4

Comparison of grid cell elevations of SETSM (2m grid) DEM tiles with dGPS 3D points, a 2m ALS DEM, a 5m DEM from traditional photogrammetry, and an ASTER DEM (A). The ASTER DEM and the 5m DEM are apparently uncorrected for the geoid. The western-most SETSM tile was 3.5 m lower than the dGPS and ALS data (B). In contrast the eastern-most SETSM tile had elevations that were in good agreement with the dGPS data, as indicated by the ellipse in panel C.

Figure S 5

Comparison of ALS 2m DEM (A) with the adjusted SETSM 2m DEM (B). Note SETSM DEM is rougher, arguably more noisy, due to interpolation between matched points identified in the SETSM photogrammetry algorithms.

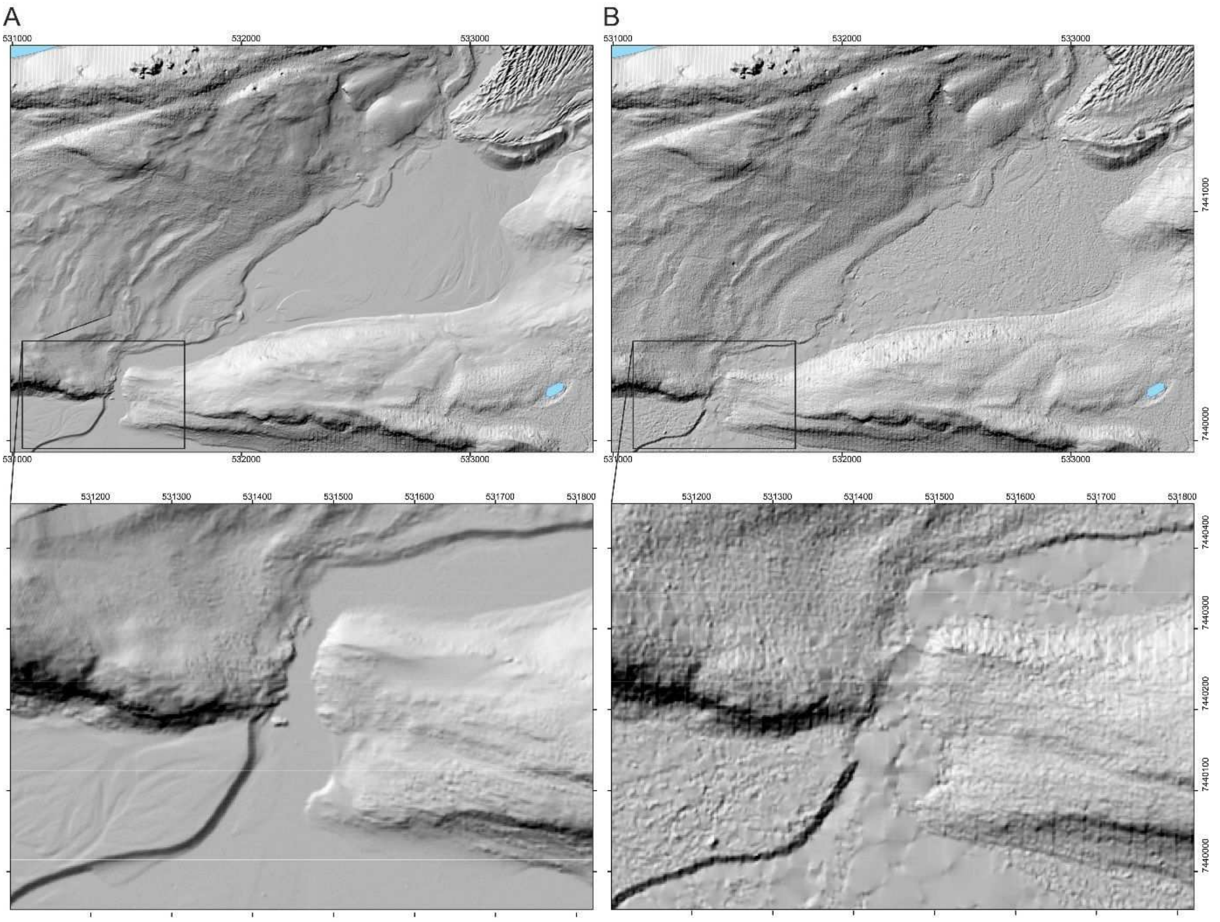


Figure S 5

Comparison of ALS 2m DEM (A) with the adjusted SETSM 2m DEM (B). Note SETSM DEM is rougher, arguably more noisy, due to interpolation between matched points identified in the SETSM photogrammetry algorithms.

Figure S 6

Field photographs of Leverett forefield (A) and of ice margin between Isunnguata Sermia and Russell Glacier (B).

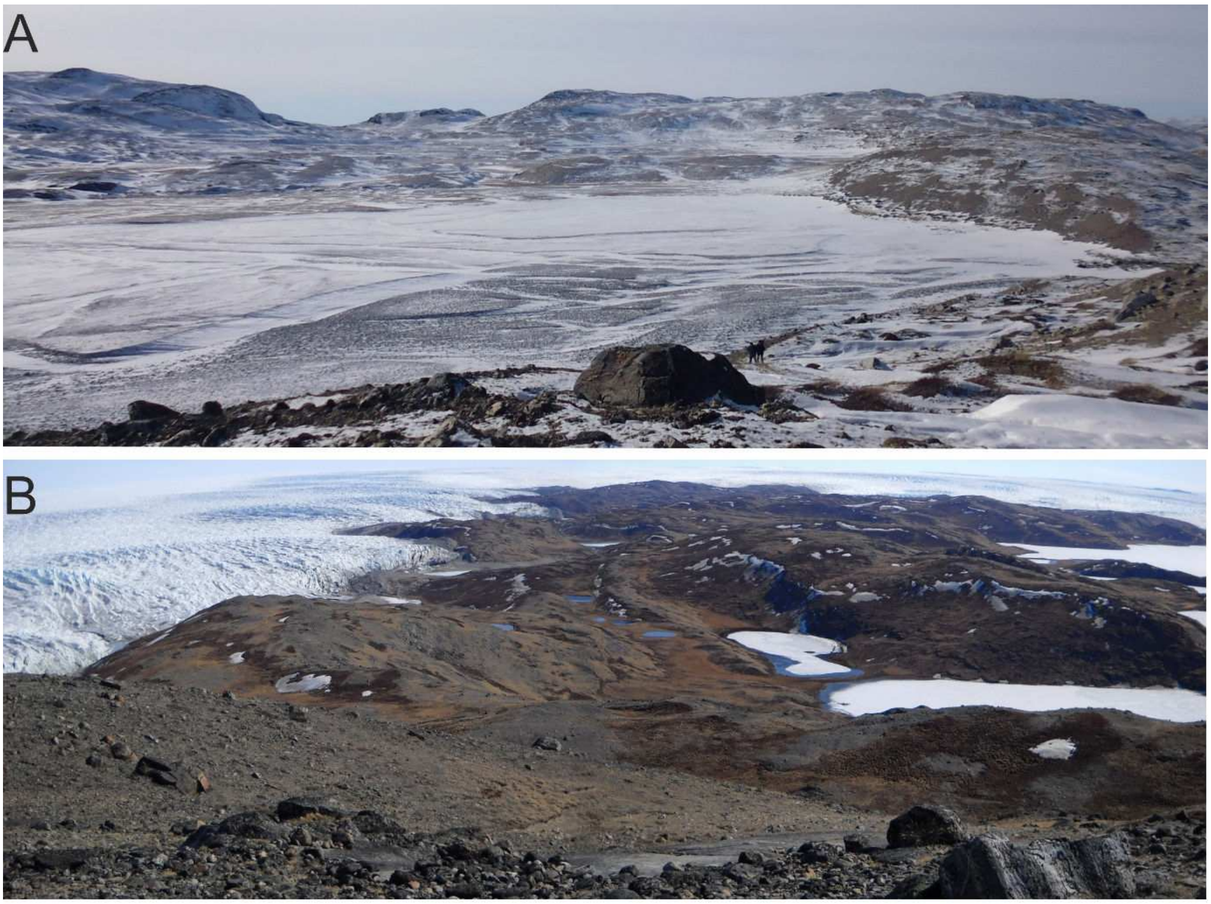


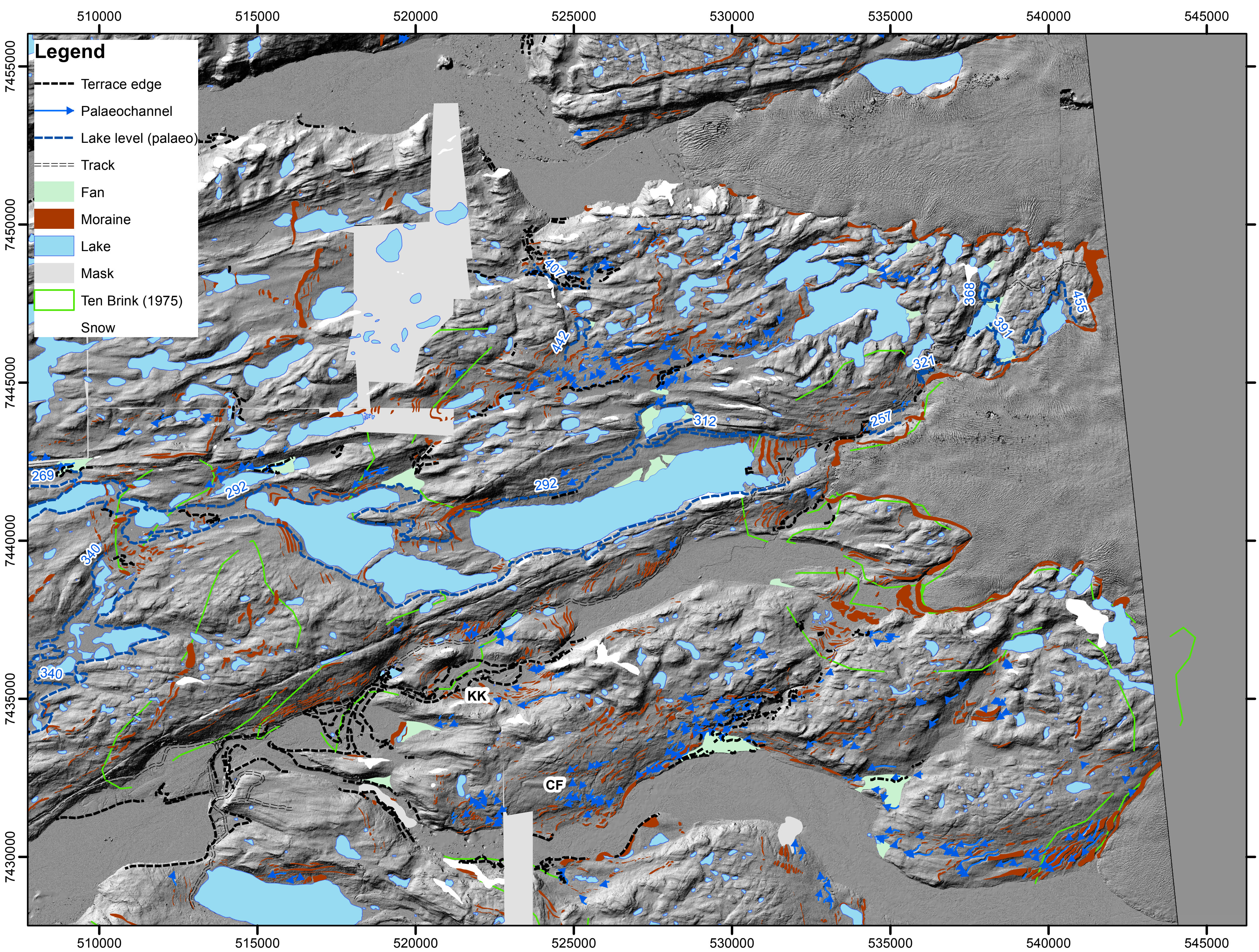
Figure S 6

Field photographs of Leverett forefield (A) and of ice margin between Isunnguata Sermia and Russell Glacier (B).

Figure S 7

Mapped geomorphology overlain on hillshaded DEM, in a .pdf file with zoom and pan functions and with layers (e.g. moraines, palaeochannels) that can be switched on and off.

<please see accompanying .pdf file>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